Groundwater studies

An international guide for hydrogeological investigations

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Preface

Understanding of groundwater has developed significantly since 1972 when the first part of the original volume of *Groundwater Studies* was published by UNESCO. Yet for someone who is just commencing the study of groundwater, there is still a need for a text which will help in starting their work. For those with a greater experience in hydrogeological investigations, there is a need to increase awareness both of more recent work and information about techniques which are outside their previous experience. This new volume is intended to meet both of these needs and therefore it has the subtitle 'An International Guide for Hydrogeological Investigations'.

This document was prepared in the framework of the Fourth Phase of the International Programme as Project M-1-3 that was supervised and directed by Habib Zebidi, Water Science Specialist, Division of Water Sciences, UNESCO and after his retirement in 1999 by his successor Ms Alice Aureli.

Dr Habib Zebidi has drawn contributors and members for an editing committee from different countries. They brought to this volume their own distinctive perspectives, for all are acknowledged experts with extensive experience of practical groundwater issues. The members of the editing committee have provided a certain unity of style end presentation and also they have tried to keep the size of this document in hand. Each chapter is intended to provide sufficient information to comprehend the fundamentals of the topic; in addition reference is made to publications where further information can be obtained for more detailed study. The contributing authors are listed below per chapter and details are given in Appendix A-3.

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1 Occurrence of groundwater, regime and dynamics

1.1 Hydrological cycle

Most water on our planet occurs as saline water in the oceans and deep underground or is contained in polar ice caps and the permanent ice cover of the high mountain ranges. So, only 30 million km^3 of fresh water, that is only 2 percent of all water, plays an active part in the hydrological cycle and in the maintenance of all life on the continents.

The hydrological cycle (Figs 1.1a and 1.1b) depicts how part of the ocean water evaporates, the water vapour turns into fresh water precipitation (rain, hail, snow) on the earth's surface (seas, land), then flows over the land surface (glaciers, runoff, streams) and partly infiltrates into the soil (soil water) to be used by the vegetation (evapotranspiration), or to recharge the groundwater bodies. Subsequently, most groundwater returns either by being pumped or by natural outflow, to surface water bodies which subsequently discharge back into the sea.

Figure1.1a The hydrological cycle (After De Wiest,1965)

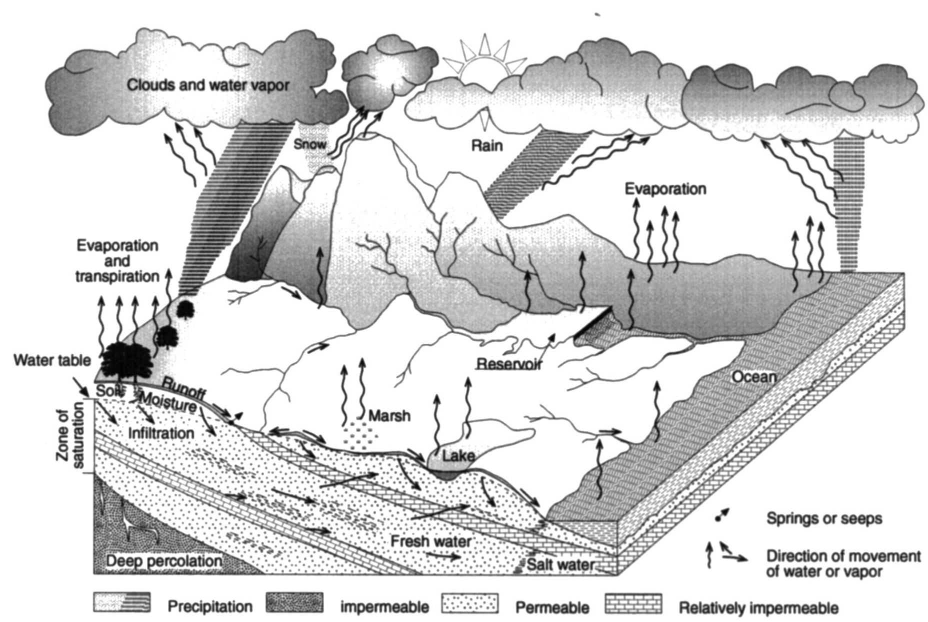
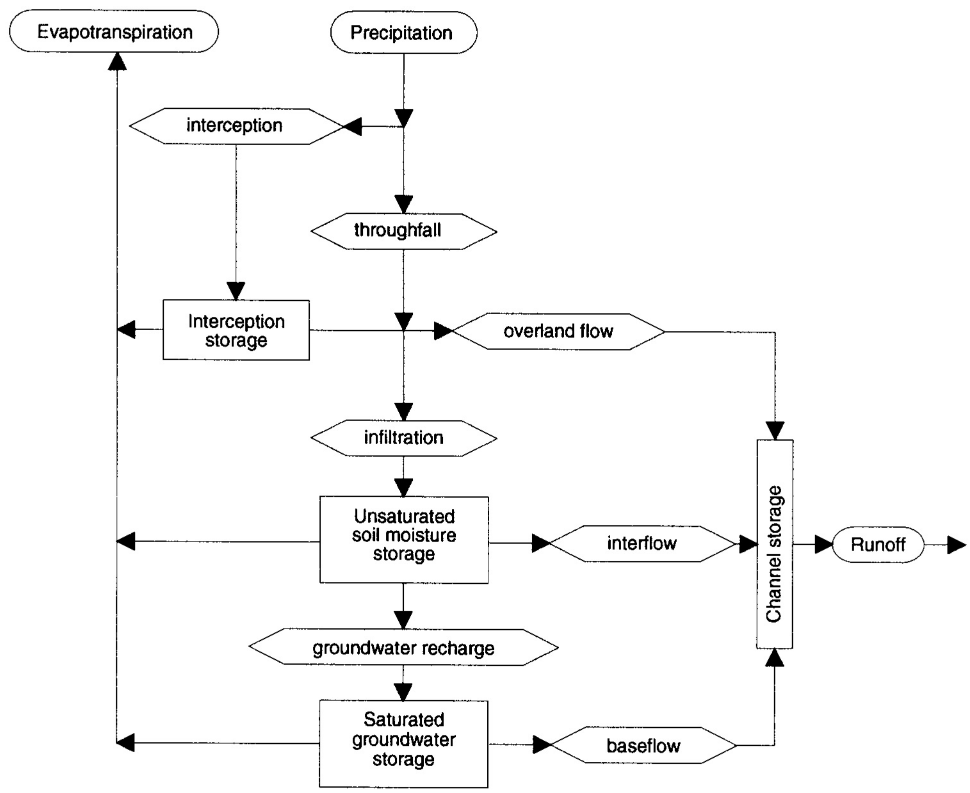


Figure1.1b Flow chart of the hydrological cycle (After Freeze and Cherry, 1979)



Water on land masses is always in motion, either moving quickly (vapour transport, precipitation, surface flow) or slowly (groundwater flow, glaciers). The slowness of groundwater flow means that most fresh water is in the form of groundwater. Consequently, groundwater is the main storage reservoir of fresh water, while surface water can be considered as the surplus precipitation that could not infiltrate, or that has been rejected as overflow from the groundwater reservoir (springs and other outflows).

When water enters the soil (infiltration) it becomes soil water; soil water may not completely fill the pores between the soil particles. Thus the zone through which the water moves is unsaturated (unsaturated or vadose zone). The flow through the unsaturated zone is essentially vertical. At the top of the vadose zone the vertical flow may be downward under the influence of gravity, or it may be upward (capillary flow) resulting from the evapotranspiration processes. To a depth of 5 m below the ground all unsaturated flow is downward (deep-rooting trees and shrubs, e.g. salt cedar, may still abstract water from the downward flow).

When vertical flow is impeded by an impervious layer, the pores above this layer fill completely and become saturated. The voids filled by the groundwater may have different origins and may occupy a smaller or larger part of the gross volume. The porosity is the percentage of a gross volume of soil or rock that is filled by air or water; the soil or rock containing such pores is called a porous medium. When the pores have their origin in the genesis of the rock, the rock is said to have primary porosity (sedimentary deposits, weathered hard rocks). When their origin is from events that occurred during a later stage (jointing, faulting, dissolution) the porosity is said to be secondary. Most sedimentary rocks with primary porosity also contain some joints and fracture that add secondary porosity to the total porosity of the porous medium. Similarly dense rocks (quartzites and granites) may have some primary porosity which is subordinate to the secondary porosity. Occasionally the effects of both forms of porosity can be recognised in the behaviour of a single groundwater body; this phenomenon is called dual porosity.

1.2 Groundwater flow

1.2.1 Groundwater flow processes and continuity

Groundwater is an important source of water; it may provide the base flows for rivers, or act as an underground reservoir from which water can be pumped as a location into which water can be drained. Consequently it is the *flow of groundwater* which must be examined.

Usually, groundwater travels very slowly; one hundred metres per year is a typical average horizontal velocity and one metre per year is a typical vertical velocity. When these velocities are multiplied by the cross-sectional areas through which the flows occur, the quantities of water involved in groundwater flow are often substantial. Consequently, the essential feature of an aquifer system is the balance between the inflows, outflows and quantity of water stored.

Unlike a surface reservoir, the upper surface of the groundwater (the water table or phreatic surface) is not horizontal; a sloping water table results from the resistance to flow caused by the hydraulic conductivity. Due to the slow movement of groundwater, care is necessary when positioning any man-made outflows, such as pumped boreholes, to ensure that they collect water efficiently from the aquifer system. The water balance of the aquifer system is the key to the identification of the aquifer resources and the consequences of changes in exploitation. The water balance is based on the principle of the *continuity of flow*.

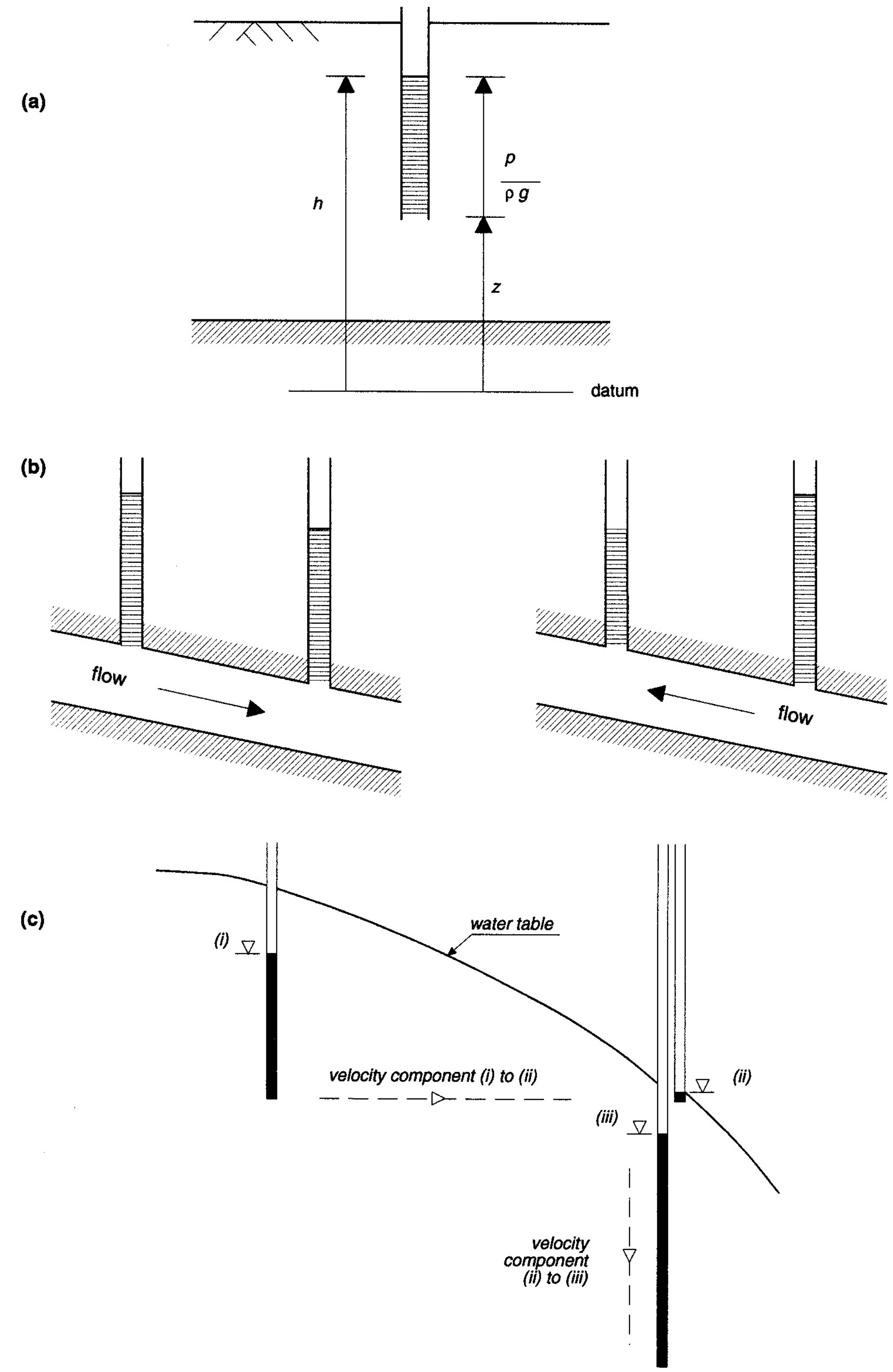
1.2.2 Groundwater head

Although the flow of groundwater is an important process, the actual groundwater flows cannot be measured directly. Consequently, an alternative method of identifying groundwater conditions is required and this is provided by the *groundwater head* (or groundwater potential).

The groundwater head at a location in an aquifer is the height to which water will rise in a piezometer (or observation well). So that the conditions at a specific location in an aquifer can be identified, the open section of the piezometer which monitors the conditions should extend for no more than a metre. Figure 1.2(a) shows a typical piezometer; the groundwater head, h, equals the sum of the pressure head $p/\rho g$, plus the datum head, z. Groundwater flows from a higher to a lower groundwater head. Typical examples of the use of the groundwater head to identify the direction of the groundwater flow are shown in Fig. 1.2. In Fig. 1.2(b) a confined aquifer is represented in which there are two piezometers to identify the direction of flow. In the left-hand diagram the flow is from left to right because the lower groundwater head is in the piezometer to the right, that is in the direction of the dip of the strata. For the right- hand diagram the groundwater gradient is to the left, since the water level in the left-hand piezometer is lower; consequently the direction of the flow is up-dip in the aquifer. This flow could be caused by the presence of a pumped well or a spring to the left of the section.

The diagram in Fig. 1.2(c) refers to an unconfined aquifer in which the water table has a significant slope. There are three piezometers (which can be considered as two pairs) provided to identify the *horizontal and vertical components* of the flow. Piezometer (ii) penetrates just below

Figure 1.2 Significance of groundwater head; (a) definition, (b) identifying flow directions in one dimension, (c) identifying flow directions in two dimensions



the water table, hence it provides information about the water table elevation; the other two piezometers are within the aquifer and are only open at the bottom of the piezometer. Therefore they provide information about the groundwater head at the bottom of the piezometer. Since the open sections of piezometers (i) and (ii) are at the same horizontal elevation, they provide information about the *horizontal flow component* (or *velocity component*); the horizontal flow is clearly from left to right. Piezometer (iii) is positioned directly below piezometer (ii) and provides information about the *vertical flow component* on that section. Since the groundwater head in piezometer (iii) is below that of piezometer (ii) there is a vertically downward component of flow. The horizontal and vertical flow components can be combined vectorially to give the magnitude and direction of the flow.

1.2.3 Darcy's Law, hydraulic conductivity and transmissivity

The magnitude of the flow can be calculated from *Darcy's Law*; the Darcy velocity, v, can be determined from

$$v = -K \cdot i \tag{1.1}$$

where

K is the *hydraulic conductivity* or (permeability) [L/T],
i is the hydraulic gradient between two piezometers,
the *minus sign* signifies that flow is in the direction of falling groundwater head.

This equation can be used to estimate the flows for the two examples of Figure 1.2(b). The same equation can be used to calculate the horizontal and vertical flow components for the example of Fig 1.2(c).

The Darcy velocity calculated from Darcy's Law is an artificial velocity, since it assumes that flow occurs through the whole cross section. In practice the flow occurs only through certain pores and fissures; consequently, an approximation to the actual velocity can be obtained by dividing the Darcy velocity by the effective porosity.

In regional groundwater studies, an important quantity is the horizontal flow through the aquifer system, Q_h; this flow can be calculated from the equation

$$Q_h = -T \cdot i \tag{1.2}$$

where *T*, the *transmissivity* is the sum of the permeabilities over the saturated depth. For example, the transmissivity in the *x* direction,

$$T_x = \sum_{\text{sat depth}} k_x \, dz \tag{1.3}$$

1.2.4 Inflows and outflows

A study of the groundwater flow within an aquifer requires information about *inflows* and *outflows*. The term recharge is used for the inflow to an aquifer system arising from precipitation, return flow from irrigation and flows from various surface water bodies such as rivers, canals and lakes. The magnitude of the recharge is likely to change significantly with time. Two books published by the International Association of Hydrogeologists provide extensive information about recharge; *Groundwater Recharge* (Lerner et al., 1990) reviews the methods of estimating recharge in a range of climates whereas in his book Simmers (1997) focuses on semi-arid and arid areas. There can also be inflow from other aquifers.

Outflows from the aquifer system can be divided into natural outflows and man-made outflows. Natural outflows occur when water leaves the aquifer at springs or into rivers. Other natural outflows include low-lying areas which act as a sink to groundwater systems; this form of outflow may be associated with areas of evapotranspiration especially from deep-rooting vegetation. These low-lying areas often form wetlands which have a high ecological value. One further natural outflow occurs when water flows into other aquifers.

There are also man-made outflows. Pumped wells and boreholes are the main means of withdrawing water from an aquifer; different designs of wells and boreholes are required for different types of aquifer and different discharge rates. Since the velocities in the vicinity of the pumped borehole are far higher than the natural groundwater velocities, there is a risk of deterioration of the aquifer in the vicinity of the well or borehole and a deterioration of the borehole structure. Horizontal wells or adits provide alternative means of collecting water from an aquifer; this approach is especially suitable for shallow aquifers or for aquifers with thin lenses of good quality water.

1.2.5 Storage coefficients and time dependency

Decreases in the volume of the water *stored* in an aquifer release water to flow through the aquifer, especially during periods with little recharge. There are two types of storage coefficients (see Figure 1.3):

- *storage coefficient* of a confined aquifer S_A [dimensionless]; this is the amount of water released from a column of unit cross-sectional area of a confined aquifer for a unit decline of the piezometric surface.
- *specific yield* of an unconfined aquifer S_Y [dimensionless], this is the amount of water released from a column of unit cross-sectional area of an unconfined aquifer for a unit decline of the water table (phreatic surface)

The storage properties of an aquifer allow continuing exploitation of the aquifer during periods of poor recharge. Consequently it is necessary to consider the *time-variant* behaviour of an aquifer. In periods of high recharge, inflows to the aquifer system in excess of the outflows may be stored in the aquifer although the resultant rise in the groundwater heads may lead to increased outflows to springs or rivers. During periods with little or no recharge, water is withdrawn from storage. Due to the important time-variant response of aquifers, it is essential to obtain all data on a time-variant basis and it is advisable to study aquifer conditions over a number of years before reaching conclusions about the aquifer's behaviour

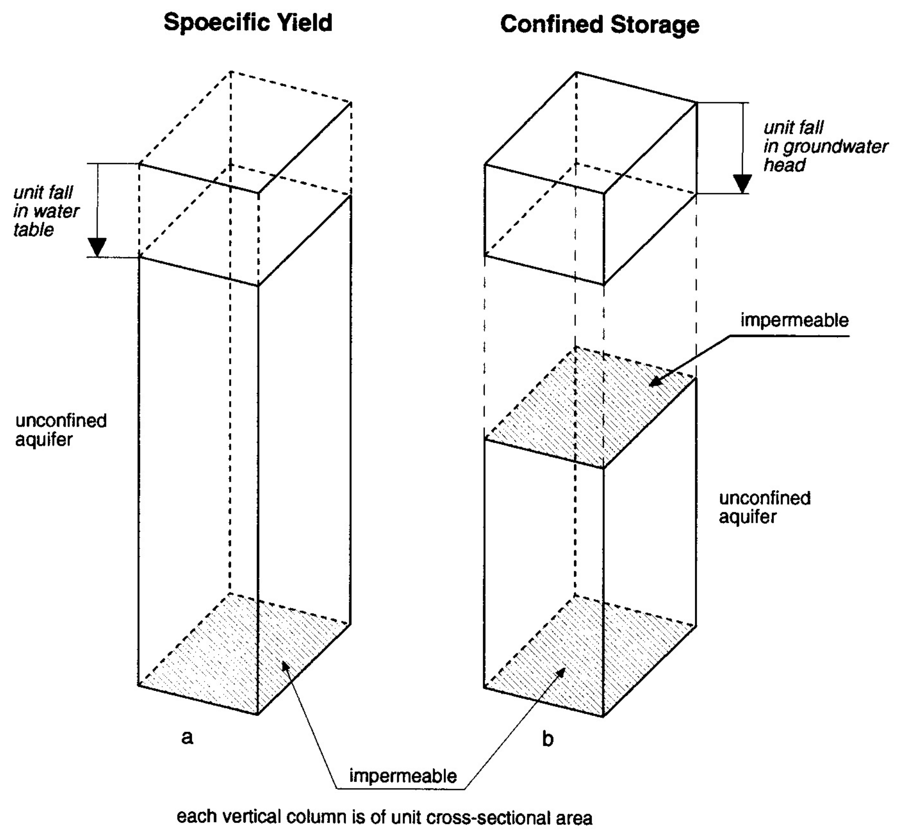
1.3 Groundwater composition

1.3.1 Physical and chemical properties

A water molecule consists of two atoms of hydrogen (H) and one atom of oxygen (O), so it has the chemical formula H_2O. At sea level its freezing point is 0°C and its boiling point is 100°C. Water is a good solvent and natural water always contains some chemicals in solution. The total amount of dissolved solids in a water sample (TDS) is expressed in mg/l and water is classified according to the TDS as fresh, brackish, saline or brine. The limits in this classification vary from country to country and even from study to study (see also Chapter 2).

The major cations in groundwater are usually sodium (Na^+), potassium (K^+), calcium (Ca^{++}), magnesium (Mg^{++}) and the major anions are chloride (Cl^-), bicarbonate (HCO_3^+),

Figure 1.3 Description of storage effects; (a) specific yield, (b) confined storage



sulphate (SO_4^{--}) and nitrate (NO_3^-). Of the solutes that occur in minor amounts the following are mentioned because of their influence on the water use: iron (Fe; taste, staining), boron (B; toxicity to plants), fluoride (F; health risk), aluminium (Al; health risk), nitrate (NO_3^-; health risk). Very small amounts of other ions, usually called trace elements, are often present in natural water. Furthermore, small amounts of the isotopes of hydrogen, such as deuterium ^2H, tritium ^3H and oxygen ^{18}O occur in all natural waters. The use of the analysis of the isotope content of these environmental isotopes is discussed in Chapter 9.

Groundwater that comes from deep aquifers (> 2000 m) or from aquifers in contact with subterranean (volcanic) heat sources may have high temperatures and may be used as a source of geothermal energy; this topic will not be discussed in this publication.

Water with a particular chemical composition may be exploited as 'mineral water' for bottling and for 'medical' use in health resorts; this is another topic not covered in this publication.

1.3.2 Risk of groundwater pollution

Groundwater pollution from human activities has become a major topic of groundwater research and large amounts of money are currently being invested in the prevention of groundwater pollution and in the rehabilitation of polluted groundwater bodies. The contaminants that may pollute groundwater are grouped according to their physico-chemical characteristics in order to characterise their fate in the groundwater environment:

- metals;
- oxy-anions;
- dissolved organics;
- non-aqueous phase liquids (NAPLs);
- colloids and radionuclides;
- bacteria and viruses.

(i) Metals

Dissolved metals usually occur as cations in groundwater, but important exceptions exist such as chromium (Cr) and uranium (U), which may also occur as oxy-anions. The mobility of metal cations often increases with decreasing pH, for a combination of two reasons. Firstly, most minerals that are formed by metals are less soluble at increasing pH: carbonates, oxy-hydroxides, sulphides. Secondly, the sorption capacity of solid phases for cations increases with increasing pH.

(ii) Oxy-anions

Oxy-anions have less singular characteristics in groundwater than metals. The sorption capacity for anions generally increases with decreasing pH. However, the importance of sorption strongly varies for individual anions and oxy-anions and is still an active topic of research. However, it has become clear that sorption is relevant for oxy-anions that can be considered as weak bases, like phosphate, arsenate and chromate. Sulphate, being a strong base, is weakly adsorbed, and adsorption of nitrate, which is also a strong base, is negligible.

Oxy-anions that behave as weak bases are most mobile under weakly acid conditions (pH approximately 5 to 6). At higher pH the mobility is limited by solubility of minerals and at lower pH it is limited by sorption.

Several oxy-anions are not stable within the entire range of redox conditions that can be found in groundwater which makes them susceptible to redox processes. With decreasing redox potential , the major anions, nitrate and sulphate may be converted to N_2 and H_2S, respectively. The former is unreactive in groundwater, the latter may form sulphides with Fe or other heavy metals. Arsene occurs in three redox states in groundwater: arsenate, arsenite and arsenic. Their mobility is distinctly different, since arsenic binds to sulphides. This limits As concentrations in groundwater in strongly reduced environments. The redox state for chromium also varies from Cr(VI) as $Cr_2O_7^{2-}$ to Cr(III) as Cr^{3+}. The latter is also much more susceptible to sorption and precipitation than the former.

(iii) Dissolved organics

Dissolved organics are relevant in the groundwater environment in two different ways. First, dissolved organic matter may be a reductant of the groundwater system , i.e. its decomposition brings about anaerobic conditions in the groundwater. Second, organic molecules may be undesirable in groundwater because of their toxicity. The first condition is normally indicated by the concentration of Dissolved Organic Carbon (DOC) and refers to organic matter as a major contributor to the overall groundwater composition. The latter refers to individual organic species at low concentrations and these species are referred to as micro-organics. Examples are pesticides, polycyclic aromatic hydrocarbons and chlorinated aliphatic hydrocarbons.

The transport of micro-organics is controlled by aqueous solubility, sorption and degradation. As soon as the organic carbon content of the aquifer exceeds 0.1% solid organic matter is the major sorbing compound. Degradation of micro-organics may be controlled biotically or abiotically. Often, biotic degradation is faster than abiotic. The degradation rate strongly depends on the redox type of the groundwater/aquifer system.

(iv) Non-aqueous phase liquids

Following spills on the surface, non-aqueous phase liquids (NAPLs) may occur as immiscible fluids in the subsurface. The flow of these liquids is hydrodynamically not geochemically controlled. Their behaviour depends on their density; fluids like petrol, diesel, etc. are lighter than water and form floating layers. On the other hand, several solvent fluids like trichloroethene are heavier and form sinking layers. Non-aqueous phase liquids are important as a source of dissolved organic matter in groundwater. The DOC may change the redox status of the aquifer system and the soluble compounds of oil derivatives may cause deterioration in the groundwater quality. Benzene, toluene, ethylbenzene and xylene, for example, are compounds in oil that show relatively high mobility in groundwater.

(v) Colloids and radionuclides

Contamination associated with nuclear activities deserves special attention. Radionuclides behave chemically in an identical manner to their non-radioactive isotopes, but physically they may show some distinct features. Radionuclides may be attached to colloids.

Colloidal particles can be transported faster than the average linear groundwater velocity. The reason is that a colloid has to be considered as a distinct particle having a specific radius. This makes it impossible to enter pores having smaller radius and also impossible to move along the edge of pores. Effectively, the flow rate is higher than for water molecules themselves or dissolved species. The charge characteristics of the colloid compared to that of the solid matrix complicate the description of colloidal transport. Colloids may be repelled from the edge when they have a similar charge and may be attracted if they have opposing charges.

Contaminants that are adsorbed to colloids can thus be transported at unexpectedly high flow velocities. One of the most relevant examples is radioactive caesium that is adsorbed by colloidal Fe-hydroxide particles. Radioactive contamination via colloid-facilitated transport usually has a local nature, since the aquifer acts as a filter for colloidal particles. Aquifers having large pores such as gravel deposits, fractured zones or karstified zones, are more susceptible to contamination caused by colloids that carry contaminants than fine-grained aquifers.

(vi) Bacteria and viruses

Bacteria and viruses may cause diseases if the contaminated water is used for drinking. These bacteria are referred to as pathogenic bacteria. The contamination is often related to sewage or waste water. Short-circuit flow from the surface to well screens is a well-known cause of bacterial contamination, which can often be attributed to poor well construction.

Bacteria can be considered as living colloidal particles. Viruses are particles that have a smaller radius than bacteria: in the order of 0.01 μm versus 1 μm, for bacteria. Like colloids, the mobility of bacteria is primarily controlled by filtration. The movement is enhanced by large pores, i.e. coarse matrices having small sorption capacities. Survival of pathogenic bacteria is encouraged by high moisture contents, low temperature and neutral pH.

Filtration is less important for viruses due to their small size. The dominant factor affecting their movement is adsorption. Survival of viruses is favoured by high moisture contents and low temperatures. The mobility of bacteria and viruses is largely determined by the same factors. Greatest movements occur in coarse aquifers and infiltration areas with thin unsaturated zones. Fractured and karstified rock have the highest potential risk. Contamination by bacteria or viruses is often local, for similar reasons to colloidal-facilitated radionuclide transport.

Both pathogenic bacteria and viruses decay in groundwater, but this needs time. The decay has been described as:

$$N_t = N_0 \, e \, {-kt} \qquad (1.4)$$

where:

N refers to the number of species,
k is the degradation constant, and
t is time.

Observed values for the degradation constant range from 0.001 to 0.06 hr^{-1} for the groundwater environment.

Table 1.1 presents an overview of the types of geochemical and biogeochemical reactions that control the fate of contaminants in groundwater.

Table 1.1 Overview of the types of (bio)geochemical reactions that control the fate of contaminants in groundwater

	metals / cations				(oxy)anions									
	K, NH_4, R_b, Cs	Al	transition	heavy	NO_3	SO_4	PO_4	As, Se	DOC	micro organics	radio nuclides	bacteria/viruses	inorganic colloids	NAPLS
acid/base	NH_3						+							
redox			Cr,Fe Mn	U				+	+	+			+	(+)
precipitation/dissolution														
carbonates			(+)				(+)							
oxides		+	Fe,Mn				+							
sulphides			(+)	Pb				As						
sulphates		+	Fe			+								
aqueous complexing		+	+	+										
non-specific sorption	+	+												
specific sorption	+		+	+			+	+						
decay/degradation									+	+	+			(+)
NAPL dissolution										+				+
filtration												+	+	

1.4 Groundwater assessment and exploration

1.4.1 Aquifer and groundwater systems

Any groundwater assessment study requires knowledge of the nature of the aquifer system and groundwater conditions.

(i) Aquifer system

The aquifer system comprises:

- the geometry (extent and thickness) of the aquifer or aquifer system and possible interlayered aquitards,
- the boundary conditions: head controlled, flow controlled or no-flow boundaries
- the aquifer type(s): confined, semi-confined (leaky), unconfined (phreatic), or perched unconfined (Figure 1.4)
- the hydraulic parameters are derived from the properties of the aquifer material:
  - porosity n (dimensionless),
  - intrinsic permeability k (dimension L^2) is a function of the grain-size diameter, rounding and sorting,
  - compressibility of the rock matrix, a, that ranges from 10^{-6} to 10^{-8} for clay to 10^{-8} to 10^{-10} m^2/N or Pa^{-1} for gravel.

For fresh water these fundamental parameters combine into the commonly used terms:

hydraulic conductivity: $$K = \frac{k \rho g}{\mu} \ [L^1T^{-1}] \qquad (1.5)$$

where:

ρ and μ are the density and viscosity of the water, and
g is acceleration of gravity

transmissivity $$T_x = \sum_{\text{sat depth}} k_x \, dz \ \ [L^2T^{-1}] \qquad (1.3)$$

The transmissivity for vertical flow through an aquitard is:

$$T_v' = K_v'/M' \ [T^{-1}]; \qquad (1.6a)$$

where

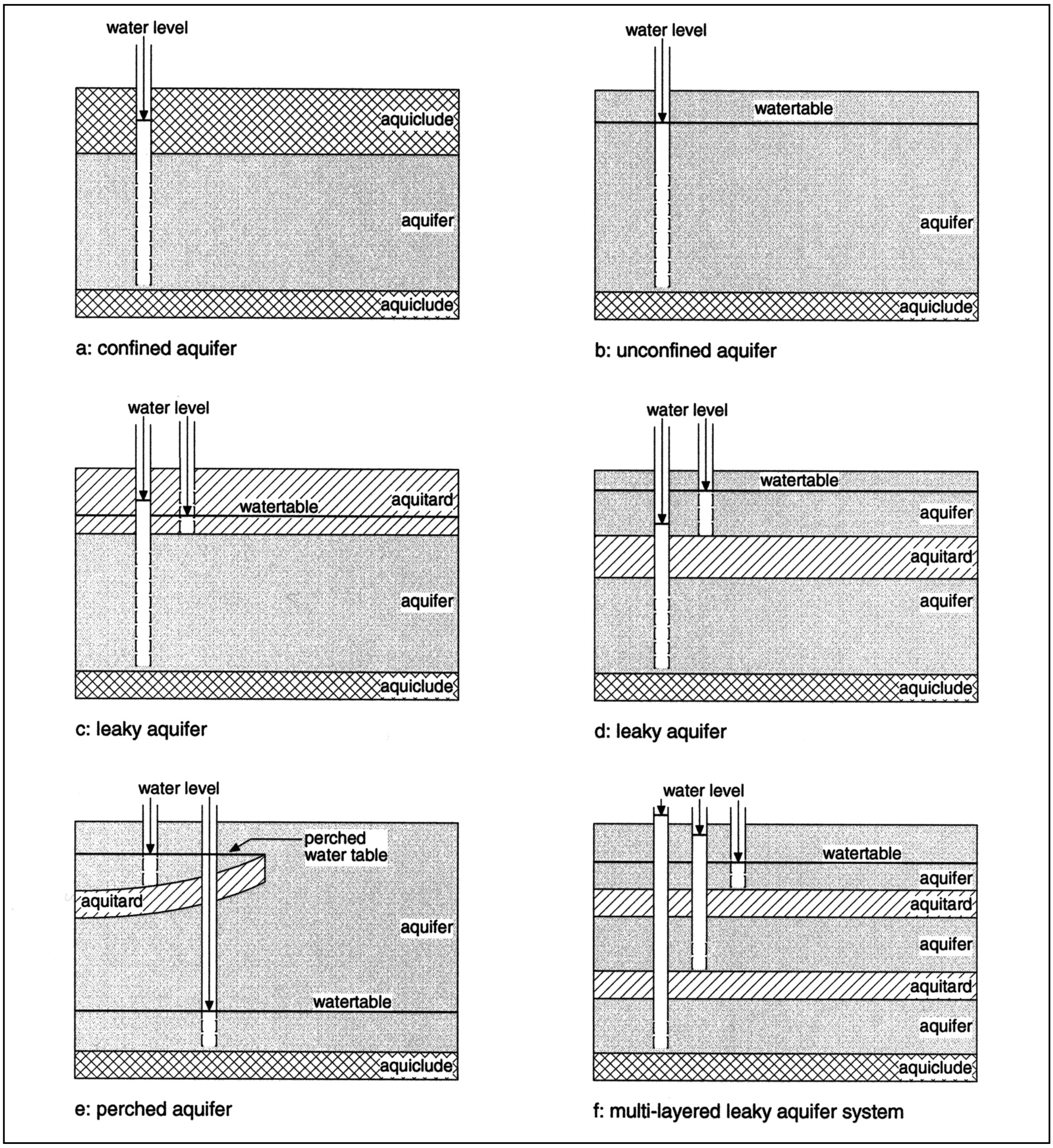
K_v' and M' are the vertical hydraulic conductivity and thickness of the aquitard, respectively.

The reciprocal of this value is known as the 'hydraulic resistance' of the aquitard:

$$c = M'/K_v' \ \ [T] \qquad (1.6b)$$

Fractures in a rock formation (Figure 1.5) strongly influence the fluid flow in that formation. Consequently, conventional well flow equations developed primarily for homogeneous aquifers do not adequately describe the flow in fractured rocks. An exception occurs in hard rocks of very low permeability if the fractures are numerous enough and are evenly distributed

Figure 1.4 Aquifer types: (a) confined aquifer, (b) unconfined aquifer, (c and d) leaky aquifer, (e) perched aquifer, (f) composite aquifer system



throughout the rock; then the fluid flow will only occur through the fractures and will be similar to that in an unconsolidated homogeneous aquifer.

A complicating factor is the fracture pattern, which is seldom known precisely. This means that, based on the geological data, a fracture model must be assumed (Fig. 1.6). Although many theoretical models have been developed in recent years, few of the associated well functions have been tabulated. Therefore the discussion will be restricted to fracture models for which the tables have been published (Kruseman et al., 1990).

The double-porosity concept regards fractured rocks as consisting of matrix blocks with a primary porosity and low hydraulic conductivity, separated by fractures with a low storage

Figure 1.5 Porosity systems: (a, b) primary single porosity, (c, d) secondary single porosity, (e, f) double porosity

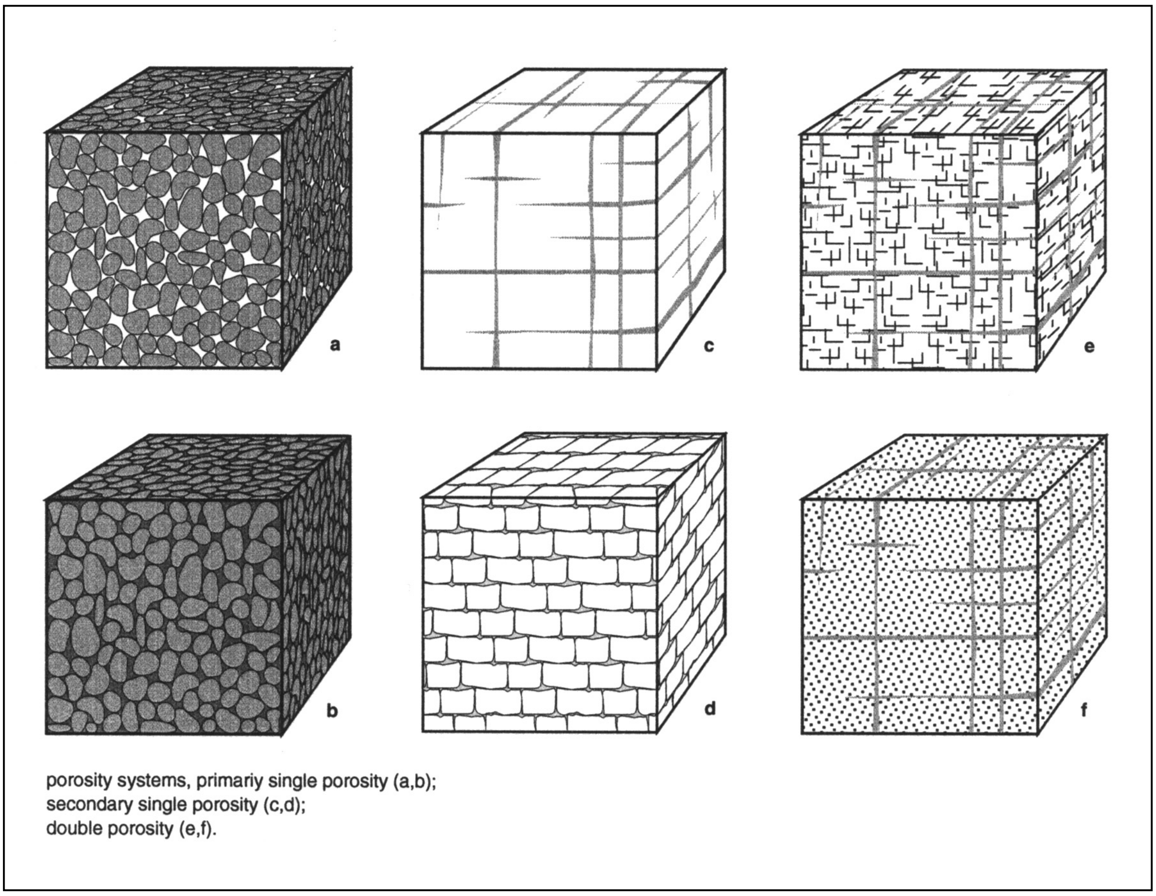
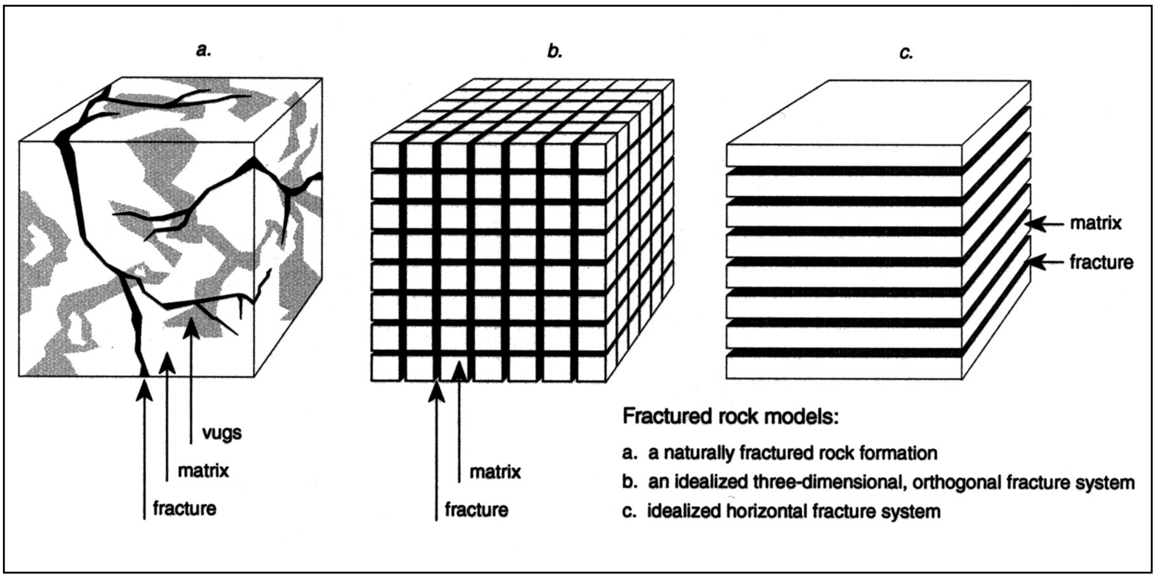


Figure 1.6 Fractured rock models: a) a naturally fractured rock formation, b) an idealised three-dimensional, orthogonal fracture system, c) idealised horizontal fracture system



capacity but a high hydraulic conductivity. This concept assumes that no variation of head occurs in the matrix blocks, i.e. the inter-porosity flow is in pseudo-steady state. The flow through the fractures in the vicinity of a pumped borehole will be radial and non-steady, Figure 8.1. Curve-fitting methods and straight-line methods have been developed to analyse pumping tests in double-porosity aquifers (Chapter 8).

(ii) Groundwater system

The groundwater system comprises many components:

- the quantity of groundwater stored in the aquifer system and its quality,
- water table (phreatic) levels and their fluctuations over time indicating changes in the amount of water stored in the aquifer,
- groundwater head (piezometric level) fluctuations of confined or semi-confined aquifers indicating the changes over time of the hydraulic pressure in the aquifer,
- recharge and discharge sources and the time-dependent rates of discharge and recharge from each source (hydraulic stress),
- groundwater budget, being a comparison between the sum of all recharge and other inflow components and the sum of all discharge components plus the change in storage over a specific period of time (e.g. six months, one year, etc.),
- chemical composition.

(iii) Groundwater models

As an understanding of the groundwater system develops, mathematical groundwater flow simulation models can be used to check the consistency of the data. Mathematical models range from the Theis equation for radial flow to a well to complex three-dimensional flow and contaminant transport models. Adjustment of the parameters and variables within physically realistic limits allows the refinement (calibration) of the model so that the model reproduces the historical changes in groundwater heads and groundwater flows.

In the same way a mass transport model can be developed to integrate the hydrochemical data.

As the precise parameter values are often scarce and are obtained indirectly (e.g. by pumping tests, see Chapter 8) and recharge values may be difficult to estimate, adjustments may be necessary before the model is properly calibrated. This implies that sufficiently long time series data (groundwater heads, recharge and discharge rates) are required to develop a groundwater flow simulation model. Without historical data a model may have no real similarity with the actual physical problem.

When a calibrated model has been prepared, it is assumed that because it can reproduce the past behaviour, it can also predict the future behaviour under changing hydraulic stress conditions. For further details on models and modelling the reader is referred to Anderson and Woessner (1992).

1.4.2 Data collection

A successful groundwater investigation depends on field data (see also Chapters 3 and 4). However, to minimise additional fieldwork each groundwater study should start with the collection and analysis of the available information and documentation. Based on the results of this analysis, it will be decided whether this information is sufficient to carry out the assessment or whether additional information must be collected in the field. At that time, a preliminary report is prepared that contains the available information and a plan of additional field studies. The following sections consider the kind of information that is required and how and from where it can be obtained.

(i) The geometry of the aquifer

The geometry of the aquifer or aquifer system and possible interlayered aquitards, is determined by the extent and thickness of the aquifer. These geometrical parameters are derived from the study of the geology of the area, borehole drilling data (for details see Chapter 7), geophysical well logs and geophysical surface studies, e.g. geo-electrical surveys and seismic surveys (for details, see Chapter 6).

(ii) Water table and piezometric level

The water table and piezometric level data are obtained from dedicated observation wells and piezometers that are regularly monitored and from incidental measurements.

(iii) The aquifer type(s)

Figure 1.4 shows the main different aquifer types: confined, semi-confined (leaky), unconfined (phreatic), perched (i.e. two independent unconfined aquifers, one above the other); in many real life the aquifer system is often more complex (see Chapter 11).

A *confined aquifer* is completely filled with water and bounded above and below by an impervious layer. The water level in a piezometer that taps the aquifer rises to a level above the top of the aquifer (the piezometric level). If the pressure in the aquifer is such that the piezometric level lies above the land surface, the piezometer may become a free-flowing (artesian) well.

An *unconfined aquifer* is bounded below by an impervious layer, but is not restricted by a confining layer above it. Its upper boundary is the water table that is free to rise and fall. Water in a well which just penetrates an unconfined aquifer is at atmospheric pressure and does not rise above the water table.

A *semi-confined*, or *leaky*, aquifer is an aquifer whose upper and lower boundaries are aquitards, or one boundary is an aquitard and the other is an aquiclude. Water is free to move up or down through the aquitards. If a leaky aquifer is in hydrological equilibrium, the water level in a well tapping it may coincide with the water table. The water level in the well may also stand above or below the water table, depending on the recharge and discharge conditions.

(iv) The hydraulic parameters

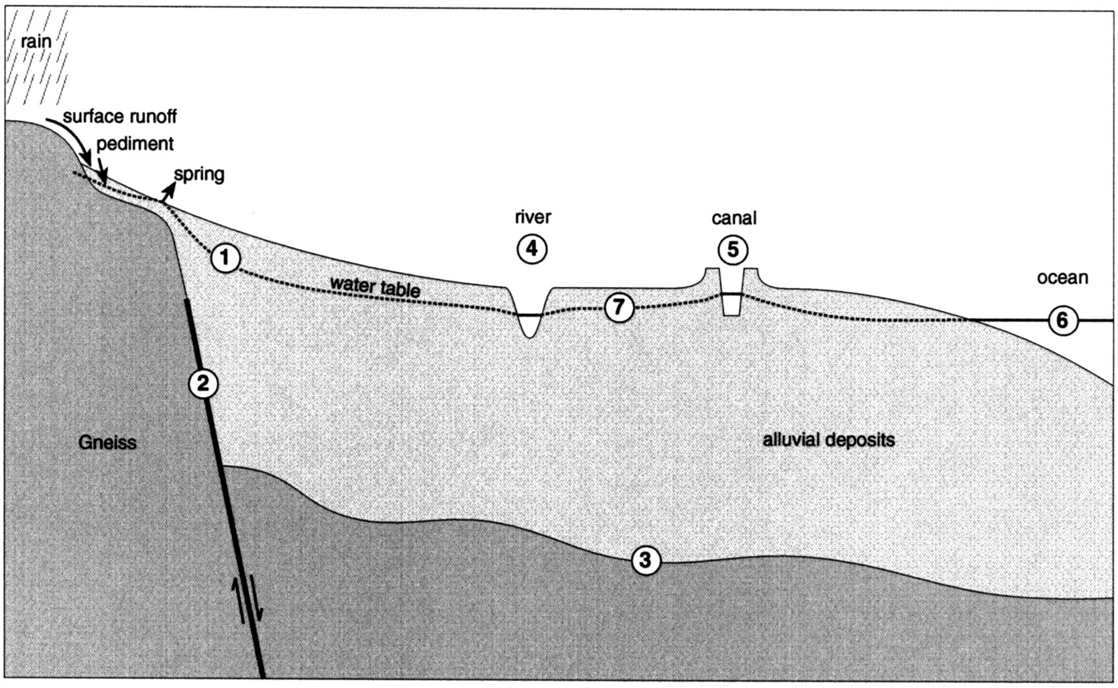
Methods to determine the hydraulic parameters mentioned in section 1.4.1 are discussed in Chapter 8.

(v) Boundary conditions

Figure 1.7 shows the three boundary types. If the boundary of an aquifer consist of an impervious barrier, there will be no flow across that boundary; i.e. it is a no-flow boundary. If the boundary is pervious, flow may occur if there is a head difference between the groundwater on either side of the boundary. The amount of flow is determined by this head difference and the transmissivity at the boundary; i.e. it is a head-controlled boundary. If the flow across the boundary is not determined by a head difference, the boundary is said to be flow-controlled, e.g.

the inflow from a karst aquifer into an alluvial aquifer. The boundary conditions are determined from an analysis of the hydrogeological context.

Figure 1.7 Aquifer boundary conditions; 1) flow-controlled boundary, 2) external zero-flow boundary, 3) internal zero-flow boundary, 4) and 5) internal head-controlled boundaries, 6) external head-controlled boundary, 7) free surface boundary (After Boonstra and De Ridder, 1981)



Aquifer boundary conditions (after Boonstra and de Ridder, 1981).

1 flow-controlled boundary
2 external zero-flow boundary
3 internal zero-flow boundary
4 5 internal head-controlled boundaries
6 external head-controlled boundary
7 free surface boundary

(vi) Groundwater storage

The volume of groundwater stored in an aquifer is calculated as the product of the thickness of the saturated part of the aquifer and its effective porosity; this volume is much larger than the exploitable amount of groundwater, because:

- part of the water is retained in the pores, i.e. the difference between the total porosity and the specific yield,
- the yield of a pumped well in an unconfined aquifer may decrease with the diminishing thickness of the saturated part of the aquifer.

The yield of a confined aquifer is determined by the confined storage coefficient, which is several orders of magnitude smaller than the porosity. In deep aquifers, the pumping lift may become uneconomically large before the pumping water level falls to the top of the aquifer.

(vii) Groundwater budget

The groundwater budget is given by:

groundwater recharge − groundwater discharge = change in groundwater storage

The *recharge* components include:

- percolation of precipitation,
- percolation of irrigation water,
- percolation from other surface water bodies, e.g. rivers, lakes,
- lateral subsurface inflow,
- vertical subsurface inflow through an underlying or overlying aquitard, artificial recharge.

Discharge components include:

- direct evapotranspiration from the water table,
- spring flow and exfiltration,
- lateral subsurface outflow,
- vertical subsurface outflow,
- discharge by pumping or other man-made devices.

(viii) Groundwater quality

Sampling of groundwater for water quality analysis not only comprises the sampling technique itself but also the set-up of the sampling well, the materials used, the type of field measurements and conservation techniques prior to analysis in the laboratory (see also Chapters 2 and 3). The form of the sampling depends on the purpose of the water quality determination, e.g. regional differentiation between aquifers, identification of recharge sources, study of local groundwater pollution, monitoring of the supply of water for domestic use, etc.

A critical issue is the correct choice of the materials for the sampling equipment. Teflon materials are preferred for the completion of the sampling well because Teflon is the most chemically inert plastic and is therefore superior to other materials such as polypropylene and polyethylene or rigid PVC. When the content of metals has to be determined, steel and iron – particularly the latter – should be avoided.

The manner of collecting the groundwater sample may influence the outcome of the analysis and is most critical for volatile compounds, in particular for volatile organic chemicals. In order to collect a sample that is not influenced by a long presence in the well the latter should be flushed repeatedly before a sample is collected. A practical rule of thumb is flushing for three times the well volume. All materials that come into contact with the groundwater during sampling should be cleaned between two samplings to avoid cross-contamination. Special care should be taken to avoid contact with oil or grease from the engines or other parts of the sampling equipment.

Sampling for the analysis of possible microbiological contamination requires sterile equipment and sampling bottles, as well as procedures that prevent contamination of the sample during its handling.

1.5 Groundwater exploitation and management

1.5.1 Monitoring conditions within the aquifer system

When considering groundwater exploitation and management it is essential to have a versatile system of *monitoring* conditions within the aquifer system; a detailed presentation on aquifer monitoring can be found in Chapter 3. This brief résumé concentrates on three important aspects of groundwater monitoring.

(i) Monitoring groundwater heads

As explained in Section 1.2, groundwater heads provide invaluable information about the flow conditions in the aquifer system. Consequently, the changing conditions in an aquifer can be identified by the monitoring of groundwater heads preferably in purpose drilled observation boreholes and piezometers. The use of boreholes which are open to the aquifer over a significant proportion of their depth is likely to lead to erroneous results, especially if the borehole is close to a pumped borehole. Figure 1.8 compares the response of an open borehole and the same borehole when a number of individual piezometers were installed in the borehole; these piezometers were influenced by the pumping from a nearby supply borehole. From the difference between the continuous line which represents the individual piezometers and the broken line which refers to the open borehole, it is clear that the single curve for the open borehole is misleading (Rushton and Howard 1982). In fact the open borehole acts as a large vertical fissure transferring water from top to bottom of the aquifer; this so disturbs the aquifer flows that the result from the open borehole is certain to be misleading.

The frequency and duration of groundwater head measurements must reflect the changes in aquifer conditions. Close to a pumped borehole, frequent readings are required, whereas at greater distances less frequent readings are acceptable. However it is essential to take sufficient readings to identify the peaks and troughs of groundwater heads; the minimum frequency of readings should be every two months.

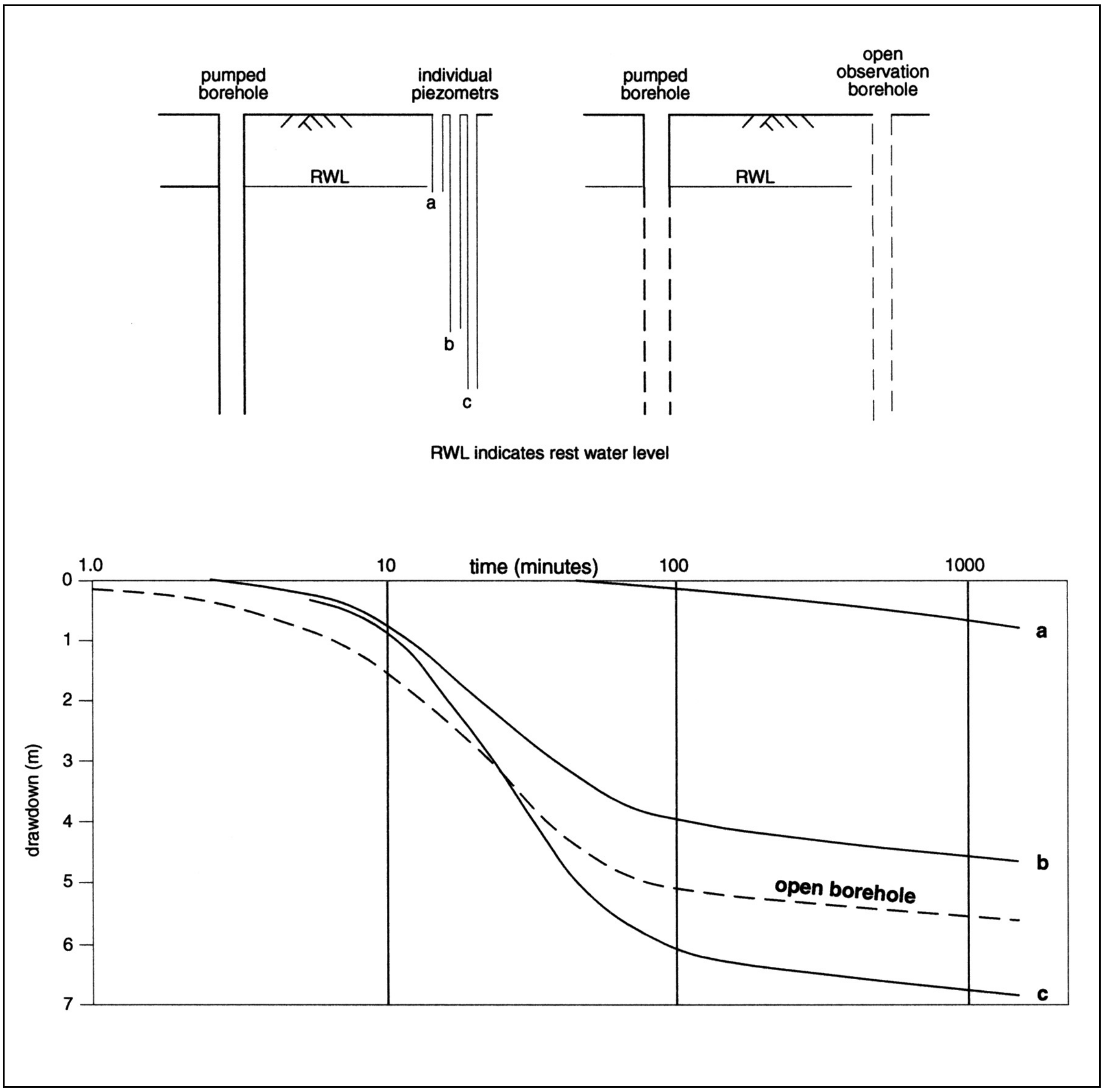
(ii) Monitoring groundwater flows

The direct measurement of flows in an aquifer is rarely possible, although valuable information can be gained from the use of tracers. Estimates of inflows such as recharge can be derived from tracers and from lysimeters; these estimates suffer from the limitation that they are point readings. Integrated values of groundwater outflows can be gained from estimates of the change in river flow between two locations; monitoring of spring flows can also provide information about local outflows. One component of groundwater flow which can be measured accurately is the quantity abstracted from boreholes.

(iii) Monitoring groundwater composition

The chemical composition of groundwater provides crucial information about the historical aquifer conditions and current changes within the aquifer system. Because groundwater under natural conditions moves so slowly, the overall distribution of the chemical components often provides information about the long term conditions in the aquifer system. However for parts of the aquifer in the vicinity of sources of inflow or outflow, changes in the chemical composition can give clear information about the current groundwater flow processes. In determining the chemical composition within the aquifer or at outlets such as springs, rivers or pumped

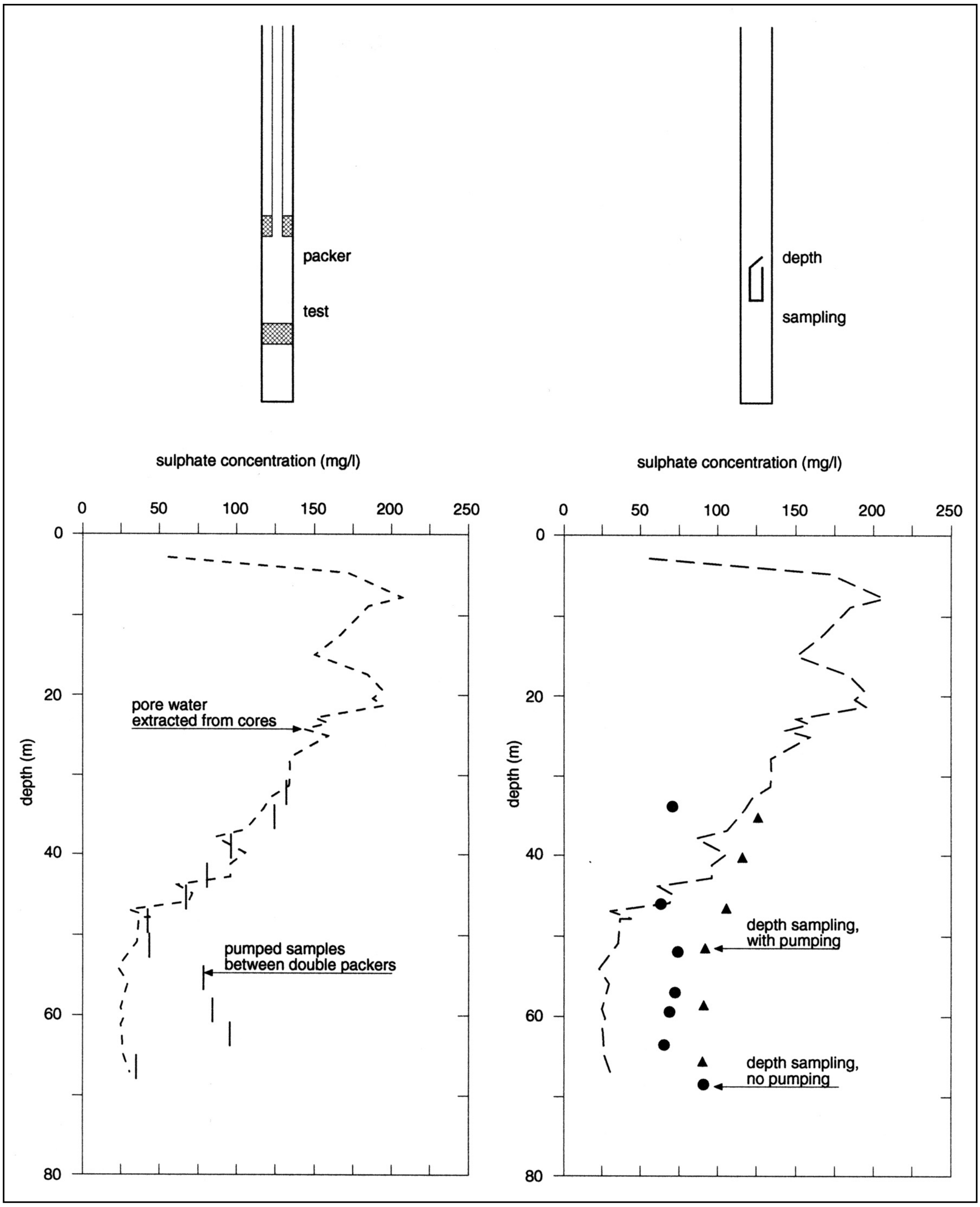
Figure1.8 Unreliability of open boreholes; hydraulic head readings of an open borehole compared to individual piezometers close to a pumped borehole



boreholes, great care must be taken to ensure that the sample is not contaminated or allowed to change in composition before testing.

The determination of the groundwater quality within an aquifer is often attempted using open observation boreholes. Again, serious errors can arise due to the mixing of groundwaters in open boreholes. Price and Williams (1994) compared sulphate concentrations obtained from depth samples in an open borehole, both with porewater samples obtained when the borehole was drilled and also with samples obtained from sections of the aquifer which were isolated by packers. Figure 1.9 shows that the depth samples in an open borehole do not reflect accurately the conditions within the aquifer. Also, samples obtained from pumping between packers fail to reflect the water quality at depth, since the higher sulphate water from the upper part of the aquifer flows into the lower part of the aquifer, causing contamination.

Figure 1.9 Unreliability of open boreholes: chemical sampling in an open borehole compared to cores and packer testing



1.5.2 Water balances and simulation models

Before any changes are made to the exploitation of an aquifer, or management rules are introduced, it is necessary to understand the flow processes within the aquifer system. The first step is usually to develop a *water balance*; subsequently a *simulation model* may be devised. Water balances should be prepared for regular periods, such as every six months, or every year. The initial attempt at a water balance is unlikely to give components which sum to zero; however, any lack of balance should provide insights into the components which require further investigation. There are two main purposes in developing a simulation model:

- To confirm whether there is a valid understanding of the overall aquifer system; this is achieved by bringing together all the parameter values and estimates of quantities such as the recharge and river/aquifer interaction to see whether historical responses can be reproduced. If there is difficulty in reproducing certain historical responses, the simulation model can be used to explore alternative parameter values and may lead to the need for further fieldwork. Until this stage of the model development is completed, the simulation model should not be used for making predictions.
- Once the validity of the simulation model has been confirmed, it can be used for predictions related to master planning of a large system or examining the impact of a limited number of boreholes. In the predictive simulations it is possible to examine the consequences of various exploitation scenarios; alternatively the objective may be to see how deteriorating conditions within the aquifer can be corrected.

1.5.3 Potential consequences of changed groundwater conditions

Groundwater exploitation, the construction of irrigation schemes and urban development can all lead to major changes in groundwater conditions.

Groundwater development leading to heavy exploitation can cause significant falls in the water table and a reduction in the yields of boreholes due to the extra height over which the water has to be raised; if the aquifer is over-exploited it may become unusable. Another consequence of heavy exploitation and the removal of large volumes of water can be settlement of the ground (Downing, 1994).

- Rising water tables can occur due to excessive irrigation, losses from canals or return flow from irrigated fields; the result of the rising water table may be water-logging and salinisation. In urban areas, the natural drainage by surface water courses and through the aquifer system may be unable to take away excess water; unless sufficient additional drainage is provided ,water tables may rise to the ground surface. Riyadh in Saudi Arabia provides a typical example where it was thought that the aquifer would provide natural drainage but the vertical permeability of underlying strata is too small to take away the recharge which results from losses from the water supply system and the irrigation of gardens; the result is that the water table is close to the ground surface in many areas of the city (Rushton and Al-Othman, 1994).
- In a number of major cities the original source of water for domestic and commercial purposes was groundwater; due to heavy exploitation in the past there have been significant falls in the water table elevation. Because of the unreliability of the groundwater (both in terms of quantity and quality) water is imported from another catchment. Consequently, the rate at which water is pumped from the aquifer has decreased, resulting in a stabilisation and subsequent rise in the water table. This water table rise is enhanced by leakage from water mains and sewers and may lead to flooding of basements and underground services and a reduction in the load-bearing capacity of foundations (Johnson, 1994; Wilkinson and Brassington, 1991).
- Changing abstraction from an aquifer can influence the flows to springs, rivers, and other areas of seepage; this can lead to a severe deterioration in the environmental conditions. The maintenance of wetlands has become an important issue in a number of countries. An increase in the volume of water pumped from an aquifer will affect

surface water features and will therefore have an environmental impact; however, it is sometimes possible to relocate abstraction boreholes and modify abstraction patterns so that the environmental impact can be minimised.

- A further consequence of increased exploitation of an aquifer is that poor quality water can be drawn in from other parts of the aquifer system. There are numerous field examples of saline water being drawn into coastal aquifers or poor quality old water being drawn up from depth (Wheater and Kirby, 1998).

1.5.4 Cost of investigations and groundwater development

A groundwater investigation can be costly since it may involve:

- drilling trial boreholes,
- monitoring groundwater heads, flows and chemical characteristics for a number of years,
- carrying out a thorough analysis of all the data and
- developing a simulation model.

Nevertheless, investigations which are carefully planned, implemented and interpreted are usually cost effective, since they can lead to a suitable abstraction regime minimising the risk of providing too few or too many boreholes and can identify any undesirable effects on the environment.

When a groundwater scheme has been implemented it is important to continue with monitoring to ascertain whether the consequences, as identified in the initial investigation, are observed in the field. If there are unforeseen results following the implementation, any necessary modifications can be incorporated.

1.5.5 Management of groundwater[1]

The management of groundwater resources is a difficult task because it requires the co-operation of all the users. Frequently there is competition for the good quality groundwater and there is insufficient water to meet all the demands. This brings a serious risk of over-exploiting and damaging the aquifer, but the consequences of over-exploitation may take many years to become unambiguously apparent. A fall in groundwater heads may be due to a period of reduced recharge or a redistribution of abstraction sites and may not be a sign that abstraction exceeds recharge. Careful monitoring of the water table elevation is the key to the identification of unsustainable abstraction from the aquifer (Rushton, 1994).

Suitable control rules often involving legislation are needed for the management of an aquifer. One approach is a system of licences to exploit the groundwater, but an aquifer may be over-licensed historically. Furthermore, when considering a licence application it may be difficult to be certain about the consequences of the increased abstraction; time-limited licences are a sensible way of determining the effect of increased abstraction. In countries where there is no enforceable legislation about groundwater abstraction, over-abstraction is likely to occur.

Studies have been made of the optimal use of aquifers, but they tend to have little relevance in real-life situations. It is difficult to reduce or stop exploitation, since large investments are required for drilling boreholes, with serious financial losses if the boreholes cannot be used. If the borehole capacity exceeds the inflows into the aquifer system, one approach is to limit the drawdown in all the pumped boreholes; this is similar to defining a minimum water level in a surface reservoir.

1. Management of groundwater is considered in detail in Chapters 12 and 13.

1.6 References and additional reading

ANDERSON, M. P.; WOESSNER, W. W., 1992. *Applied Groundwater Modelling*, Academic Press, 381 pp.

BOONSTRA, J.; DE RIDDER, N. A., 1981. *Numerical Modelling of Groundwater Basins; A User-oriented Manual.* International Institute of Land Reclamation and Improvement, Publication No. 29, Wageningen, 226 pp.

DE WIEST, R. J. M., 1965. *Geohydrology.* John Wiley and Sons, New York, 366 pp.

DOWNING, R. A., 1994. Falling Groundwater Levels, a Cost-Benefit Analysis. In: *Proceedings, ICE International Conference on Groundwater Problems in Urban Areas,* Thomas Telford, London, pp. 213–36.

FREEZE, R. A.; CHERRY, J. A., 1979. *Groundwater.* Prentice-Hall, Englewood Cliffs, NJ, 604 pp.

JOHNSON, S. T., 1994. Rising Groundwater Levels: Engineering and Environmental Implications. In: P*roceedings, ICE International Conference on Groundwater Problems in Urban Areas,* Thomas Telford, London, pp. 285–98.

KRUSEMAN, G. P.; DE RIDDER, N. A.; VERWEIJ, J. M, 1990. *Analyses and Evaluation of Pumping Test Data (Second edition).* International Institute of Land Reclamation and Improvement, Publication No. 29, Wageningen, 371 pp.

LERNER, D. N.; ISSAR, A. S.; SIMMERS, I., 1990. Groundwater Recharge. A Guide to Understanding and Estimating Natural Recharge. *International Contributions to Hydrogeology,* Verlag Heinz Heise, Vol. 8, 345 pp.

PRICE, M.; WILLIAMS, A., 1993. The Influence of Unlined Boreholes on Groundwater Chemistry: A Comparative Study Using Pore-Water Extraction and Packer Sampling. *J. Inst. Water and Eng. Man,* Vol. 7, pp. 651–9.

RUSHTON, K. R., 1994. Assessing Aquifer Exploitation Using Observation Boreholes. *J. Inst. Water and Env. Man.,* Vol. 8, pp. 156–64.

RUSHTON, K. R.; HOWARD, K. W. F., 1982 The Unreliability of Open Observation Boreholes in Unconfined Aquifer Pumping Tests. *Ground Water,* Vol. 20, pp. 546–50.

RUSHTON, K. R.; AL-OTHMAN, A. A. R., 1994. Control of Rising Groundwater Levels in Riyadh, Saudi Arabia. In: *Proc. ICE International Conference on Groundwater Problems in Urban Areas,* Thomas Telford, London, pp. 299–309.

SIMMERS, I., 1997. Recharge of Aquifers in (Semi-)arid Areas. *International Contributions to Hydrogeology,* A. A. Balkema, Rotterdam, Vol. 19, 277 pp.

WHEATER, H.; KIRBY, C., 1998. *Hydrology in a Changing Environment,* Vol. II, John Wiley and Sons, New York, 604 pp.

WILKINSON, W. B.; BRASSINGTON, F. C., 1991. Rising Groundwater Levels – An International Problem. In: *Applied Groundwater Hydrology: a British Perspective*. Oxford University Press, pp. 35–53.

2 Groundwater quality

2.1 Introduction

2.1.1 Definition and scope of chemical hydrogeology

Chemical hydrogeology is an interdisciplinary science that relates the quality of groundwater to various processes and reactions in aquifers and aquitards: the quality of groundwater in a particular aquifer is described, evaluated and explained primarily by application of principles of aquatic chemistry to a hydrogeological system. In other words chemical is the study of the geological and hydrological controls on the groundwater chemistry. The objectives of chemical hydrogeological investigations are to determine the sources, concentration, and fate of dissolved constituents within the physical framework of flow and transport. This chapter emphasises the application of chemical hydrogeology, including the techniques, and procedures required to investigate groundwater quality. For general background information on chemical hydrogeology see Back and Freeze (1983).

In the early phases of chemical hydrogeology, research focused on developing analytical techniques to describe the solute concentrations and chemical characteristics of water. The approach is generally referred to as hydrochemical facies mapping. The next phase was to provide an explanation of these observations by identifying the chemical reactions and hydrogeological controls that are associated with the solute concentrations in water and with the chemistry of the water. The approach is called reaction modelling. Now increased exploitation of groundwater resources along with contamination of some supplies has brought a need for greater understanding of the concentration limits and the variations in the behaviour of the dissolved substances. We are now entering a phase in which research is directed toward predicting the solute concentrations and chemical character of water at a specified point in some future time. The ultimate goal is multicomponent geochemical transport modelling.

2.1.2 General approach

A schematic flow chart in Table 2.1 shows the objectives of a typical quality of groundwater investigation. The ultimate goal, rarely achieved at present, is to predict the quality of groundwater changes in space and time and the rates of these changes.

As in any other project, in a groundwater quality project, the desired objectives must e clearly formulated. For research, these objectives are the testing, verification or refuting of one or more hypotheses; for applied studies they are identification of the problem, and of the controlling parameters and an evaluation of alternative ways of solving or coping with the problem. Hydrogeologists are frequently required to assess the quality of groundwater on a

Table 2.1 Schematic flow chart for a chemical hydrogeological project

Needed for sampling and monitoring

(1) Preliminary interpretation of the hydrogeological system, and
(2) Reliable techniques for chemical and isotopic analysis

• In order to prepare hydrochemical maps and diagrams that will:

• *DESCRIBE*	• *IDENTIFY*
The spatial and stratigraphic distribution of dissolved constituents	Probable chemical reactions
Relation of chemical constituents to hydrogeological parameters	Significant parameters to monitor
	Mineralogical controls
	Source and flow of groundwater
	Source of contaminants

• *TO EXPLAIN*	• *TO MONITOR*
Source of constituents	Chemical changes in space
Concentration and distribution of constituents	Chemical changes over time
Hydrochemical heterogeneity	Rate of chemical changes

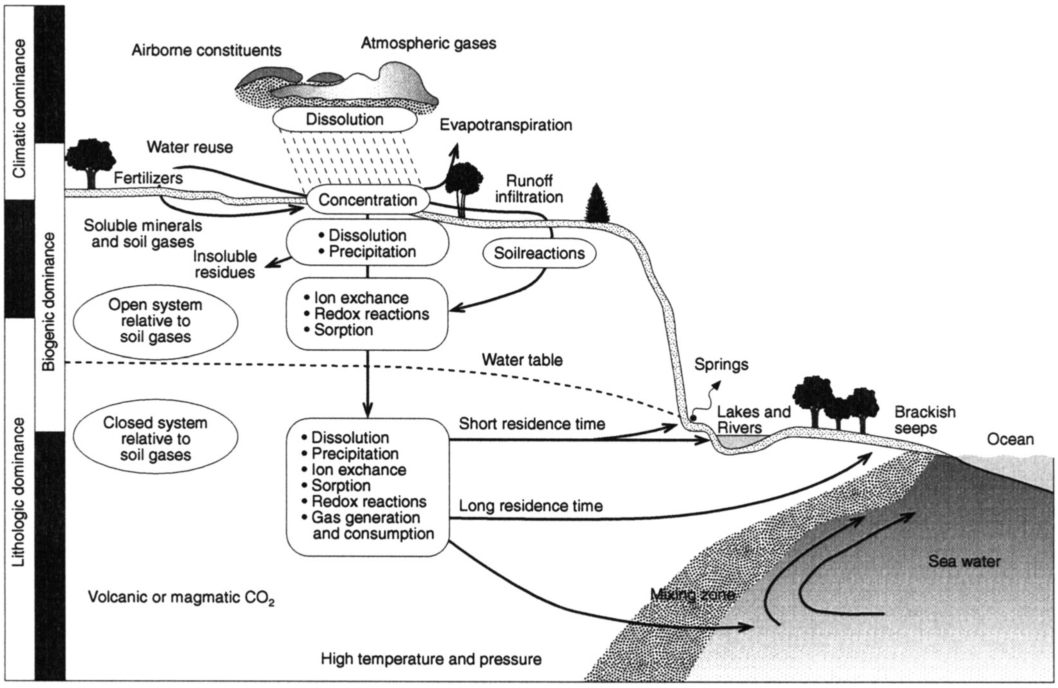
• *PROVIDES ALL OF THE ABOVE*

A better understanding of the functioning of the hydrogeological system

regional basis. This requires a sampling and monitoring programme, that is significantly different spatially and temporally from a site-specific study. When hydrogeological and geochemical principles are applied appropriately, the result is efficient, meaningful and therefore successful groundwater assessments.

To be able to identify the controlling hydrogeochemical reactions, one must know the present and past geological, hydrological, biological, soil chemical, meteorological, and human factors that affect the chemistry of water (Fig. 2.1). The input of chemical or biological species in

Figure 2.1 Factors controlling groundwater composition



the unsaturated zone of the recharge area is largely controlled by atmospheric deposition, land use (including vegetation and soil cover), the frequency, amount and duration of precipitation and irrigation, the mineralogy of the soil, the air temperature, the soil gas/air exchange rate, and transport properties. The quality of infiltrating surface water is controlled by the composition of the surface water together with the temperature, the composition of the surface water bottom sediments and the residence time within these bottom sediments. Factors that affect chemistry of groundwater in the saturated zone are chemical reaction rate, residence time within the saturated zone, and mineralogy of the rock matrix. Here, the residence time and flow path are determined by factors like aquifer thickness, permeability, porosity and amount of recharge. The phenomena of mixing of water from different areas, aquifers or confining beds, from seawater intrusion or trapped saline water, or contaminants impose a hydrological control on the chemical character of groundwater. Superimposed on all these primarily natural factors are anthropogenic effects leading to chemical and physical stresses on the hydrogeological system that may be dominant in some areas.

To be able to formulate and carry out regional studies, the following scientific and technical aspects associated with hydrogeology must be understood:

- the basic concepts of organic and inorganic chemistry;
- the use of environmental isotopes and geochemical modelling;
- principles of advective/dispersive transport and the coupling with reactions
- the consequences of surface water/groundwater interaction or soil moisture/groundwater interaction on groundwater quality;

- the effects of land use on water quality; and application of geochemical principles to regional aquifer systems;
- salt water intrusion; and specific contamination problems in karst aquifers.

A publication that thoughtfully demonstrates the application of these many concepts and principles for regional assessment of water quality is Alley (1993).

2.2 Basic principles of hydrochemistry

2.2.1 Precipitation/dissolution reactions

(i) Dissolution of neutral salts

Dissolution of a neutral salt is a reversible reaction in which the acid/base equilibria of water are not affected. An example is the dissolution of gypsum:

$$CaSO_4 \cdot 2H_2O \leftrightarrow Ca^{2+} + SO_4^{2-}$$

A typical example of dissolution of neutral salts is the dissolution of evaporites. These salts are very soluble and their dissolution strongly increases the salinity of the groundwater. When saline groundwater exfiltrates in closed semi-arid or arid basins precipitation is likely.

(ii) Weathering reactions

In these reactions primary minerals become dissolved or altered and secondary minerals may be produced. The secondary minerals are closer to chemical equilibrium with the earth surface conditions than the primary ones. Weathering reactions are associated with shifts in the acid/base equilibria. Consider for example the dissolution of Ca-carbonate in carbonic acid:

$$CaCO_3 + H_2CO_3 \leftrightarrow Ca^{2+} + 2HCO_3^-$$

Carbonate reactions are reversible. Note that dissolved CO_2 is hydrated in groundwater and exchange with the gas phase may be limited or impossible. CO_2 in groundwater is therefore indicated as H_2CO_3. Another example of weathering is the incongruent dissolution of Al-silicates. At neutral pH, Al remains in the solid and the process is represented in a general way as:

$$\text{Aluminium silicate} + H^+ \rightarrow Me^{m+} + \text{Aluminium mineral} + H_4SiO_4$$

where Me^{m+} refers to metals such as Na, K, Ca, and Mg. Silicon dissolves in pore water and/or remains in the solid. Weathering reactions involving silicates can be considered as irreversible under earth surface conditions.

2.2.2 Redox reactions

In these reactions ions electrons are transferred from one species to another. Redox half-reactions are used to describe the change of an element from the reduced state to the oxidised state or vice versa. Consider for example the half-reaction for the reduction of nitrate to nitrogen gas:

$$NO_3^- + 5e^- + 6H^+ \leftrightarrow \tfrac{1}{2} N_2 + 3\, H_2O$$

All elements that exist in more than one valence state in nature are susceptible to this type of reaction. In order to obtain a true redox reaction which occurs in aquatic systems two half reactions have to be combined. Redox reactions also affect the acidity of the system, since during the reaction protons are transferred. These reactions may co-occur with precipitation/dissolution reactions, yielding redox-controlled dissolution/precipitation reactions. For example, pyrite contains two reduced species, Fe(II) and S(–I), which may be oxidised to Fe(III) and SO_4, respectively. Oxidation of pyrite by dissolved oxygen may, therefore, be complete or incomplete, depending whether or not Fe(II) is oxidised:

$$FeS_2 + 3\tfrac{3}{4}\,O_2 + 3\tfrac{1}{2}\,H_2O \rightarrow Fe(OH)_3 + 2\,SO_4^{2-} + 4H^+$$

or

$$FeS_2 + 3\tfrac{1}{2}\,O_2 + H_2O \rightarrow Fe^{2+} + 2\,SO_4^{2-} + 2H^+$$

Many redox reactions in groundwater systems are microbially mediated and kinetically controlled. The bacteria use redox-sensitive compounds as a source of energy for their metabolism. The thermodynamic feasibility of a redox reaction is determined by the redox status of the system. That status is expressed in terms of the redox potential, the Eh. The combination of Eh and pH determine whether or not a redox reaction is feasible.

A somewhat special type of redox reaction in the subsurface is the degradation of organic matter. Organic matter is unstable and is oxidised to CO_2 by the oxidants present, or is degraded to CO_2 and CH_4. Organic matter is often simplified as CH_2O. The oxidation of organic matter by SO_4 is represented by:

$$2CH_2O + SO_4^{2-} \rightarrow HCO_3^- + HS^- + H_2CO_3$$

Methanogenesis is represented by:

$$2CH_2O + H_2O \rightarrow H_2CO_3 + CH_4$$

The degradation of organic matter is commonly a sequence of reactions that produces a variety of carbon compounds. The degradation usually induces inorganic reactions, and in many subsurface systems it is the primary control for the redox status. Degradable organic matter occurs in the solid state as a primary material, or it enters the system in the dissolved state. In the latter case it may be the result of natural or anthropogenic processes.

2.2.3 Sorption reactions

(ii) Cation exchange reactions

Cation-exchange reactions are the result of charge compensation of negatively charged clay minerals and organic matter. The occupation of the exchange complex is determined by the aqueous composition together with the affinity of the exchange complex for the cations present. The charge compensation occurs by means of an increased concentration of cations near the surface, possibly accompanied by a decreased concentration of anions. The cation-exchange reaction has been formulated in several ways. A common thermodynamically justified way to represent the exchange-reaction between Na and Ca, for example, is:

$$2NaX + Ca^{2+} \leftrightarrow CaX_2 + 2Na^+$$

where X^- represents the solid phase exchanger with charge –1. The cation-exchange capacity (CEC) indicates the total amount of exchangeable cations on the porous medium. It depends on the type of clay minerals and the amount of organic matter, and it is also pH-dependent. Cation-exchange happens as soon as the native pore water composition is displaced by water of another composition. This happens, for example, during seawater/fresh water displacement associated with seawater intrusion, etc.

(ii) Surface complexation

Surface complexation reactions are reactions in which species are transferred from the solid surface to the aqueous phase and vice versa. They typically occurs for oxides, humic and fulvic acids. Sites at the surface of the solids are protonated or deprotonated, for example:

$$\equiv S - OH \leftrightarrow \equiv S - O^- + H^+$$

where $\equiv S$ refers to a surface site. The protonation/deprotonation is strongly pH-dependent. Additionally, dissolved metals and oxyanions become sorbed at the surface sites. If the amount of proton transferred differs during sorption, several types of reaction happen. Consider, for example, sorption of copper and arsenate, which are strongly favoured by Fe-hydroxide, as examples:

$$\equiv S - OH + Cu^{2+} \leftrightarrow \equiv S - OCu^+ + H^+$$

and

$$\equiv S - OH + AsO_4^{3-} + 3H^+ \leftrightarrow \equiv S - OH_2AsO_4 + H_2O$$

Sorption of metals and oxyanions is strongly pH-dependent. Sorption of oxyanions is strongest at acid to neutral pH, and sorption of metals is strongest at neutral to alkaline pH. This behaviour is related to increasing deprotonation of the surface sites with increasing pH, which gives rise to a gradual change in charge from positive to negative.

2.2.4 Aqueous complexing

Aqueous complexing does not involve solid minerals. However, it indirectly influences reactions with the porous medium. It gives rise to an increased concentration of species, since the free species react together to form complexes. The solubility of Ca-carbonate is increased by the forming of the aqueous Ca-bicarbonate complex:

$$Ca^{2+} + HCO_3^- \leftrightarrow CaHCO_3^+$$

In groundwater systems aqueous complexing reactions are instantaneous.

2.2.5 Gas transfer

Some of the gases dissolved in groundwater are important reactants with the solid matrix. Consider for example the role of CO_2 in dissolving carbonates and that of O_2 as an oxidant of reduced compounds. Gases become dissolved in pore water proportionally to the partial gas pressure in the associated gas phase:

$$CO_{2g} \leftrightarrow CO_{2(aq)}$$

where subscripts g and aq refer to the gaseous and aqueous phases. Gases remain dissolved as long as the sum of partial gas pressures does not exceed the hydrostatic pressure. Groundwater can often be considered as closed to gas transfer, i.e., gases that react are not replenished and gas that is produced remains in the dissolved state during groundwater flow. Exceptions to this rule are groundwater near the water table and groundwater in hydrothermal or volcanic systems.

2.2.6 Ion filtration and osmosis

These reactions occur in clays and other fine-grained sediments in which the unchanged water molecule can pass freely through the semi-permeable membrane but the anions are repelled by the negative charge on the clay layers and the cations remain with the anions to maintain electrical balance. It is important in large, thick sedimentary basins, in areas with a long residence time and at interfaces between seawater and fresh water in clayey deposits.

2.2.7 Radioactive decay

Radioactive decay and fractionation of isotopes produce radioactivity in the groundwater and alter the isotopic composition of the water. This is discussed in Chapter 9.

2.3 Acquisition of chemical data

2.3.1 Sampling procedures

Ideally, the sampling programme for a geochemical groundwater investigation will collect the minimum number of samples required to have adequate three-dimensional spatial and stratigraphic coverage of the area being investigated. The fundamental task is to obtain samples that are representative, diagnostic, and characteristic of the aquifer and to analyse them with minimal change in composition. The volatile and reduced compounds are the most sensitive to shortcomings in the sampling and conservation procedures. Poor procedures that can render unreliable results include obtaining samples of water standing in a well casing, or from a well having cross flow, the use of unclean or reactive bottles, the lack of an airtight seal to prevent aeration or degassing, the lack of adequate preservation of non-stable constituents, contamination of the samples with the materials used in well construction (casing, cement, drilling mud, or other additives), the lack of proper filtration of particulate matter from the sample, and long storage time.

Before starting the groundwater sampling, the chemists in the laboratory who will perform the analysis, should be consulted for advice on sample size and the preservation techniques to be used in the field. The concise handbook on analytical techniques for water published by APHA-AWWA-WPCF is useful. Volatile compounds must be collected in gas bottles. Samples for analysis of dissolved cations must be collected in plastic bottles, filtered through a 0.45 μm filter, and acidified to pH 1–2 to prevent precipitation after sampling and sorption to the bottle.

The alkalinity (bicarbonate concentration) together with the pH should preferably be measured immediately in the field, together with temperature and electrical conductivity, since these two change rapidly after sampling. Aeration of the groundwater sample gives rise to entrance of O_2 and associated oxidation of dissolved reduced species, and dissolved CO_2 escapes when the partial pressure of dissolved CO_2 is higher than atmospheric. When sampling for reduced species, special precautions must be taken in order to preserve the actual concentrations. When sampling for dissolved gases degassing must be prevented. Positive displacement pumps are to be preferred above suction type of pumps.

2.3.2 Groundwater analysis

(i) Inorganic constituents

An analysis of groundwater may range from a value for a single component to a long list of inorganic, organic and biological measurements. The selection of the constituents for analysis is determined both by the overall objective of the investigation and the specific purpose of the chemical and biological analysis. Because of the complexity of many diverse problems for which analyses may be useful, there is no 'standard' analysis. For example, the suite of inorganic constituents for geochemical prospecting is appreciably different from the constituents used to monitor a landfill or to determine the effects of mineral diagenesis. However, the conventional point of view has been that a standard inorganic analysis consists of determination of the four major cations (calcium, magnesium, sodium, and potassium) and the three major anions (bicarbonate, sulphate, chloride) together with pH, electrical conductivity and temperature. If there is anthropogenic contamination, nitrate must also be determined. Aluminium becomes increasingly important when the pH is 6 or lower. Dissolved Fe(II) and Mn(II) may occur as major species in anaerobic groundwater. Ammonium, F, PO_4, H_2S and CH_4 are of secondary importance; their analysis is often crucial to resolve the evolution of groundwater quality along flow lines. A criterion for an accurate analysis is that the difference between the sum of cations and the sum of anions is less than 5% of the sum of these two sums. Note that this criterion only checks the major ions, and does not guarantee that the minor species have been properly analysed.

(ii) Organic constituents

Due to increasing anthropogenic contamination with organic microspecies (pesticides, petroleum hydrocarbons, chlorinated hydrocarbons, etc.) it is becoming increasingly important to obtain samples of groundwater for determination of organic compounds. The first step towards obtaining insight into the status of groundwater with respect to organic compounds determination of the total amount of dissolved organic carbon (DOC). Common compounds that make up DOC include fatty acids, humic substances, carbohydrates, amino acids and uronic acids. Organic material is frequently present in rainfall, groundwater, streams, rivers, lake and seawater. A properly carried out investigation will explain how and why the DOC varies over time and space and what processes have controlled these variations. Principles of organic geochemistry of water are increasingly applied for this purpose.

DOC is of little diagnostic value when dealing with organic microcontaminants. There are special analytical procedures for volatile and non-volatile microspecies. In the first instance, we should distinguish three major groups of interest:

- chlorinated hydrocarbons that are associated with dry-cleaning facilities, metal-working industries, chemical industry and military areas,
- petroleum hydrocarbons (including the BTEX compounds; benzene, toluene, ethylbenzene and the xylenes) which are related to petrol stations, petrochemical industry, crude oil and natural gas exploitation sites, and military areas,
- pesticides, herbicides, fungicides, which represent an extremely broad group of compounds mainly related to agricultural activities, but also used to maintain railways, parks, golf courses, etc.

(iii) Bacteria, viruses and microbial activity

Microbiota in the subsurface are studied either to find out whether pathogenic bacteria and viruses are present in groundwater or to determine the role of microbial activity in predominantly redox processes. If possible groundwater must be sampled under sterile conditions, to avoid any contamination from the earth surface. Note that during drilling of wells contamination with allochtonous microorganisms may occur.

There are routine techniques t for analysing pathogenic microorganisms. Isolation procedures and plate counts are often the basis for identifying and counting the bacteria present.

The role of microorganisms in controlling quality of groundwater is increasingly being demonstrated, especially their potential to degrade organic contaminants. An excellent book that addresses this topic is Chapelle (1993). Microorganisms and microbial activity have been studied in two different ways: the microbial activity is characterised by means of incubations and temporal analysis of the compound of interest, or the microorganisms themselves are cultivated, isolated, etc. We must realise that the latter analyses are not completely representative of the subsurface microbial community, since cultivation is selective and most bacterial species have not been a identified. Instead of studying groundwater it may be worth studying the microorganisms in a solid matrix, however, the spectrum of groundwater microorganisms is not identical to that of the solid matrix and biased results will be obtained when characterising groundwater.

2.4 Evaluation of groundwater chemical data

2.4.1 General procedure

Chemical groundwater data are interpreted to find out where the ions come from, how they reach their concentration, what is their form and behaviour, where they are going, and how fast. The major activities are:

- determining the mineral equilibrium characteristics of water using speciation models;
- displaying the spatial distribution of chemical concentrations in order to understand the hydrogeological controls on the chemical variations within an aquifer or chemical differences between aquifers;
- identifying the controlling chemical reactions;
- interpreting water analyses in terms of hydrological, geological, and anthropogenic controls;
- chemical reaction modelling to predict chemical changes in the aquifer during exploitation.

Groundwater quality data may be used to independently validate the hypotheses put forward to explain the physical functioning of the groundwater system. Hypotheses about residence times, flow paths and rates can be checked using hydrogeochemical data in combination with data about isotopes. Chemical, mineralogical and lithological data about the aquifer is of great help during interpretation of groundwater quality data. This information must be collected during drilling of the wells, piezometers, etc.

2.4.2 Characterisation of groundwater quality

A first evaluation of the groundwater quality is whether the water is fresh, brackish or salt.

Nowadays the amount of dissolved salts is most commonly expressed in one of the following ways:

- The electrical conductivity (EC) of water expressed in S/cm or in µS/cm (10^{-6} S/cm) for fresh water, or in mS/cm (10^{-3} S/cm) for salt water; where S stands for Siemens (formerly called mho), the inverse of the resistance expressed in ohm (Ω);
- Salinity, where total dissolved solids is expressed as parts per thousand of unit weight of water, for example, normal seawater has a salinity of about 35‰;
- Total dissolved solids (TDS) which is the residue on evaporation at 105°C or 180°F. A classification of water based on TDS is given in Table 2.2.

Table 2.2 Nomenclature for water

Classification	*TDS (mg/l)*
Fresh water[1]	<1,000 to 2,000
Brackish water	1,000 to 20,000
Saline water	20,000 to 50,000
Brine	**>50,000**

1. Limits vary from one country to another, depending on the salinity of available water.

The next step is to evaluate the individual constituents. Only about a dozen of the somewhat more than one hundred chemical elements are normally analysed in water samples.

Over the years, hydrogeochemists have been guided in their selection of significant constituents by the relative concentration of a particular element and its physiological, agricultural, industrial, or geochemical significance. In groundwater (Table 2.3) the concentrations are

Table 2.3 Normal range of chemical composition (in mg/l, except pH) of groundwater, seawater and rainwater away from the coast

	Groundwater	*Mean ocean water*[a]	*Rainwater*[b]
pH	5.5 – 9	8.2	4.1 – 5.6
Ca	10 – 200	400	0.1 – 2
Mg	0.1 – 100	1,350	0.05 – 0.2
Na	1 – 300	10,500	0.1 – 1
K	0.1 – 20	380	0.08 – 0.3
NH_4	0 – 5	<0.5	0.1 – 2.3
Fe[c]	0 – 10	0.01	0 – 0.2
Mn[c]	0 – 2	0.002	0 – 0.02
SiO_2	10 – 30	6.4	–
HCO_3	80 – 400	142	0
SO_4	10 – 100	2,700	0.4 – 8
Cl	1 – 150	19,000	0.25 – 2
NO_3	0 – 50	<0.5	0.3 – 4
F	0.1 – 2	1.3	0.03 – 0.1
Br	< 5	65	–
B	< 2	4.6	–

a. Hem (1986); *b.* Rainwater away from the coast (Royal Netherlands Institute of Meteorology; Appelo and Postma, 1993); *c.* The redox status is different for the water types.

intermediate between rainfall and ocean water. Rainwater is not pure water and in some coastal areas or desert areas the chemical load of the atmospheric precipitation can be a significant contribution to the chemical character of groundwater.

The chemistry of rainwater depends on such factors as the windward distance from the coast or salty lakes and soils, wind intensity, the period of rainfall within the storm, distance from cities and industrial centres.

The chemical composition of mean ocean water (Table 2.3) is typical of the open ocean not concentrated by evaporation nor diluted by fresh water from river flow or groundwater discharge. In coastal aquifers, seawater intrusion can dramatically alter the chemical composition of the groundwater and result in chemical processes within the saline water and zone of dispersion. A classification based on salinity of water from such environments (Table 2.2) demonstrates the rather arbitrary limits for fresh, brackish, and saline water and brines.

In addition to rainfall and seawater, the sources of ions are from the dissolution of minerals, sorption reactions, and degradation of organic material in the unsaturated and saturated zones. Tables 2.4, 2.5 and 2.6 present a brief summary of the sources of various chemical species in groundwater. Speciation calculations should be performed to determine the distribution of the total concentration of a compound among free species and aqueous complexes. It is the chemically active concentration of the relevant aqueous species that is calculated.

Table 2.4 Sources of major constituents

Major constituents > 5 mg/l	*Source*
Calcium, Ca	• primarily from carbonates, gypsum, feldspars
Magnesium, Mg	• feldspars, olivine, pyroxene, amphiboles, mica, Mg-calcite
Sodium, Na	• feldspars, evaporites, cation exchange, seawater, industrial waste
Potassium, K	• feldspar, fertiliser, K-evaporites, glauconite
Silicic acid, H_4SiO_4, SiO_2	• silicates
Ammonia, NH_4	• pollution, degradation of organic matter, reduced NO_3, cation-exchange
Sulphate, SO_4	• dissolution of gypsum and anhydrite, oxidation of pyrite, seawater, windborne fertiliser salts
Chloride, Cl	• windborne rainwater, seawater and brines, evaporite deposits, pollution
Nitrate, NO_3	• atmospheric deposition, decay of nitrogen-fixing plants, oxidation of ammonia or organic nitrogen, contamination
Carbonate, CO_2, HCO_3, CO_3	• soil and atmospheric CO_2, carbonate rocks, oxidation of organic material, volcanic gases
Oxygen, O_2	• soil gas and atmosphere

Table 2.5 Sources of minor constituents

Minor constituents, 0.01 to 10 mg/l	*Source*
Aluminium, Al	• clays, feldspars, amphiboles, micas
Boron, BO_3	• tourmaline, evaporites, sewage, seawater, volcanic emissions
Fluoride, F	• fluorite, some silicates, volcanic emissions
Sulphide, H_2S	• pyrite, reduced sulphate, oil field gas, volcanic emissions
Phosphate, PO_4	• apatite, fertiliser, sewage sludge, degradation of organic matter
Iron, Fe	• oxides, sulphides, carbonates and clays
Manganese, Mn	• oxides and hydroxides
Strontium, Sr	• carbonates
Organic acids, mostly humic and fulvic	• organic matter decomposition
Argon, Ar	• air
Methane, CH_4	• organic matter degradation under intense reducing conditions
Nitrogen, N_2	• air, nitrate reduction

Table 2.6 Sources of trace elements

Trace constituents, generally <0.1 mg/l and commonly below detection level	*Source*
Arsenic, As	• arsenic insecticides, pyrite oxidation
Bromide, Br	• evaporites, seawater, rainwater
Chromium, Cr	• contamination
Iodide, I	• marine vegetation, evaporites
Lithium, Li	• silicate weathering
Uranium, U	• uranium dispersed minerals, mill tailings
Vanadium, V	• not well known
Zinc, Zn, cadmium, Cd	• sulfides, industrial waste, PO_4-fertiliser salts, sewage sludge

Determination of the distribution of species and their active concentrations is an iterative process, because a set of mass-balance equations has to be solved together with a set of thermodynamic mass action equations. After the active concentrations have been calculated, the saturation state of groundwater with respect to minerals is calculated.

Conventionally, the saturation index is provided, which is defined as the log value of the ratio of the ion activity product to the equilibrium constant of the mineral. For example, the saturation index with respect to calcite is calculated as:

$$SI_{cc} = \log \{[Ca^{2+}][CO_3^{2-}]\} / K_{cc}$$

where SI_{cc} is the saturation index with respect to calcite, K_{cc} is the thermodynamic solubility product of calcite, and $[Ca^{2+}]$ and $[CO_3^{2-}]$ is the active concentration of Ca^{2+} and CO_3^{2-}, respectively. Unsaturation for a mineral (SI < 0) indicates that the mineral will dissolve if present. Saturation (SI = 0) indicates that thermodynamic equilibrium exists and the mineral equilibrium may control the concentration of the constituents present in that mineral. Supersaturation (SI > 0) increases the possibility of precipitation; non-equilibrium is the result of slow kinetics, non-ideal behaviour or inhibition of precipitation.

2.4.3 Compiling groundwater quality data

It is generally convenient to use one or more of the many classification schemes and geochemical diagrams available, in order to identify and depict the significant chemical characteristics of water in various parts of the groundwater regime or the chemical differences between aquifer systems (Schoeller, 1962; Hem, 1989; Custodio and Llamas, 1976; Back and Freeze, 1983; see also Chapter 4). Chemical information gained from such diagrams is commonly plotted on hydrogeological maps and cross-sections to illustrate the areal and stratigraphic distribution of various water types and the change over time in the chemical character of groundwater. In order to prepare hydrochemical diagrams, it is frequently necessary to convert analyses from one set of units to other units; these conversion factors are listed in the Appendix A-1, Conversion factors.

Concentration ratios are also very helpful when interpreting groundwater analyses in terms of hydrogeochemical processes. Table 2.7 lists species ratios that are of general use, together with the information that these ratios may provide. Note that these rules are based on many groundwater quality studies; however, there are extreme environments that do not behave according to these general insights. Molar ratios or equivalent-concentration ratios are often more convenient than weight-concentration ratios to identify hydrochemical ratios. To be able to accept or reject hypotheses about the ongoing processes, the information derived from the ratios needs to be combined.

2.5 Process interpretation and modelling

2.5.1 Modelling

The next step after the groundwater analysis has been interpreted in terms of the origin of groundwater and chemical evolution, is to quantitative model the processes proposed. In chemical reaction modelling, the data and calculations are used to determine:

- which chemical reactions have occurred,
- the extent to which the reactions have proceeded,
- the hydrogeological conditions under which the reactions have occurred,
- which changes may occur over time,
- the speed of such changes.

Here the calculations of mineral equilibria determine the saturation state of the water and identify which minerals will tend to dissolve or precipitate. In addition to the normal chemical

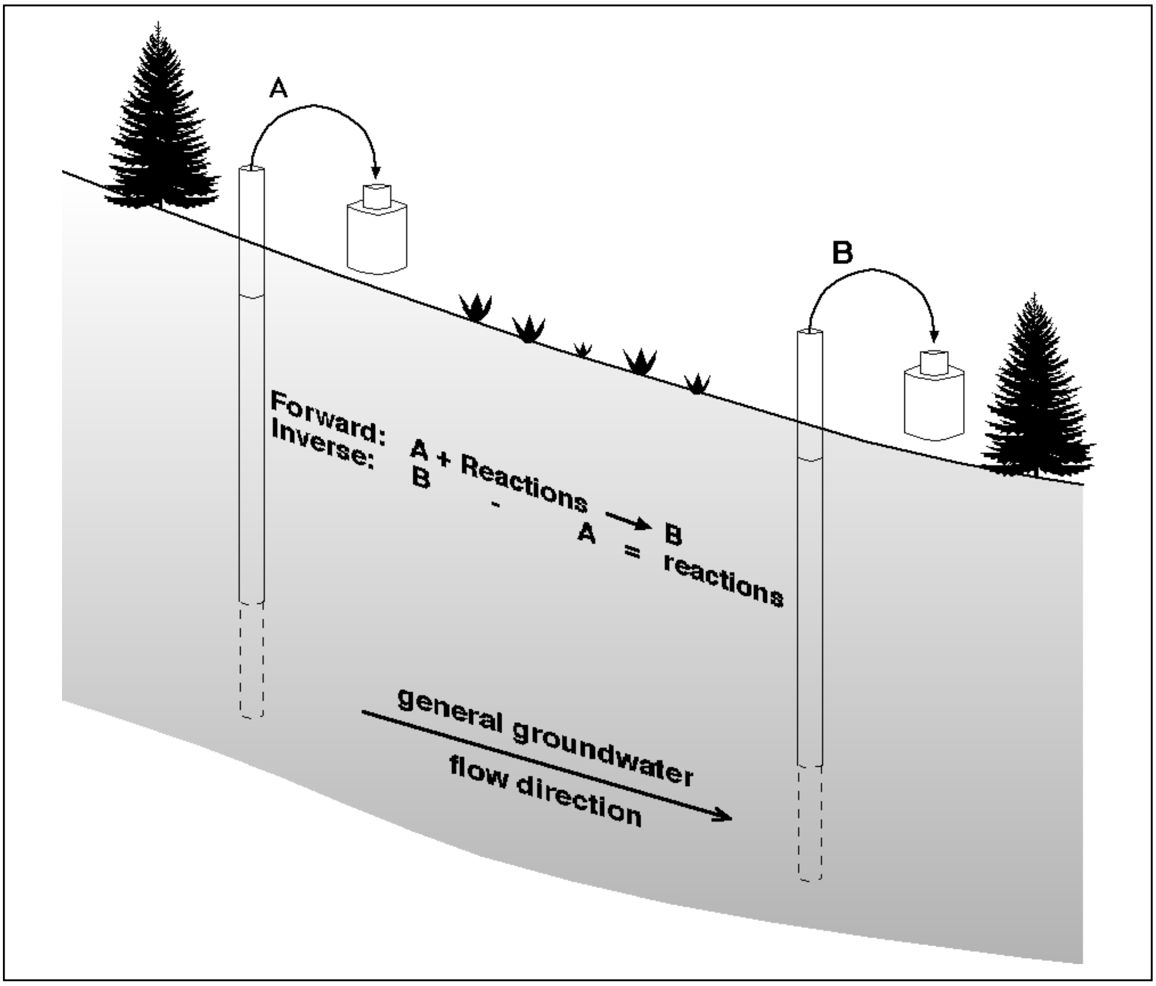
Table 2.7 Concentration ratios and their meaning in terms of process control

Ratio	*Application*
SO_4/Cl	Seawater has a typical weight ratio of 0.14. When both constituents originate from seawater smaller ratios suggest SO_4-reduction. Higher ratios suggest additional sources of SO_4 and/or additional processes such as dissolution of gypsum or oxidation of pyrite.
Na/Cl	The seawater weight ratio is 0.55. Higher values imply Na-desorption during freshening, if paleohydrological studies have revealed that marine water was once present. Higher ratios can also be indicative of silicate weathering.
Mg/Cl	The seawater weight ratio is 0.071. Deviations from this value together with deviations for Na/Cl suggest that there has been cation-exchange for the appropriate hydrogeological environment. Silicate weathering also gives rise to high ratios.
Na/HCO_3	If seawater/fresh water displacements can be excluded, high ratios indicate substantial weathering of Na-feldspar or other Na-silicates.
Ca/HCO_3	To identify Ca-carbonate dissolution (check with saturation index with respect to calcite and other Ca-carbonates) or Ca-silicate weathering.
Ca/SO_4	To identify gypsum dissolution.
Ca/Mg	To identify dissolution or precipitation of dolomite, calcite, etc. (check with saturation index with respect to dolomite, calcite, etc.). Deviations from the seawater ratio occur in coastal regions due to cation-exchange, albeit combined with carbonate reactions.
(Ca+Mg)/(Na+K)	To identify aquifer lithology: importance of carbonate reactions versus silicate weathering.
Br/Cl	To differentiate between brines and seawater.
Fe/SO_4	To identify oxidation of Fe-sulphide.
HCO_3/sum anions	To distinguish between weathering reactions and input of dissolved species from the surface in the recharge area.
NO_3/sum anions	To identify the extent of anthropogenic contamination due to e.g. agricultural activities.
(H+Al)/sum cations	To indicate the extent of acidification (use equivalent concentrations).
F/Mg	To identify biotite dissolution as the source of F.

constituents, isotopes have proved extremely useful in groundwater quality studies. Isotopes are atoms of an element with the same number of protons but different numbers of neutrons in their nucleus. The distribution of isotopic species in water provides additional information on sources of groundwater, on flow paths and mixing, and on chemical reactions and sources of ions. Isotopes measured in water include: ^{18}O, ^{2}H (or deuterium, D), ^{3}H (or tritium, T), ^{3}He, ^{13}C, ^{14}C, ^{34}S, ^{15}N, ^{37}Cl and ^{234}U. See Chapter 9 for a discussion of isotope hydrology. The ultimate objective of a groundwater quality study, including isotope chemistry, is to provide chemical input for the mass transport equation in order to predict concentrations of organic and inorganic constituents at some point in space and time. Several hydrogeochemical software models are freely available via internet, e.g., the U.S. Geological Survey offers a series of models at: http://water.usgs.gov/software/geochemical.html.

Two types of approaches are distinguished in hydrogeochemical modelling and in other types of hydrological modelling as well (Fig. 2.2): forward modelling and inverse modelling. In *forward modelling* we define an initial state of the system studied and we impose a series of processes on that system. The resulting state is calculated by mathematical quantification of the processes imposed. The result is thus one unique state of the system. In *inverse modelling* we consider two (or more) states of the system studied and from the difference between these two (or more) statuses we quantify the extent of processes that are assumed to happen. The different states of the system must of course be linked in some way, e.g., the second state of the system is a groundwater that lies along a flow line downstream from the first state of the system.

Figure 2.2 Illustration of the principles of forward and inverse modelling



Both forward and inverse hydrogeochemical modelling require a proper understanding of the ongoing hydrogeochemical processes. A conceptual model of the system and the likely reactions in that system is thus necessary before any mathematical modelling can be performed. The quality of a mathematical model is ultimately linked to the quality of the modeller's model concept: any ill-defined model concept will lead to a wrong mathematical model and consequently to wrong model results. For groundwater systems we have to deal with:

- thermodynamic constraints;
- mineralogical constraints (including non-mineralogical reacting solids);
- kinetic limitations;
- hydrological transport limitations.

The mineralogical composition of the system studied determines the type of reactions that may happen, and the thermodynamic constraints will determine the direction of the chemical processes together with the maximum amount of mass transfer of chemical species from one phase to another. Kinetic limitations will determine the actual extent to which the potentially occurring reactions have happened within the time span that is available for the system studied and whether or not a metastable thermodynamic state of the system is likely to be reached. The latter is indicated by the Ostwald's kinetic step rule: 'a system follows the kinetically most favourable path instead of the thermodynamically most feasible path'. Kinetic limitations thus determine the actual amount of mass transfer. For groundwater, the kinetic limitations have to be compared with the travel times obtained from the hydrogeological study.

Forward hydrogeochemical modelling is the as prediction of an aquatic system from an initial solution combined with a series of postulated reactions. Thermodynamic principles are used in hydrogeochemical software models and forward hydrogeochemical modelling is thus implicitly constrained by thermodynamic considerations. However, it is possible to violate thermodynamic constraints if the modeller incorporates kinetics, e.g. by defining zero-order reaction steps in the model. To what extent this may happen depends on the features of the model used. Thermodynamic constraints can also be deliberately violated in order to consider some non-equilibrium or irreversible reactions. Thermodynamic databases may be deliberately changed into databases that need to be considered as non-equilibrium databases. One may, for example, decouple the thermodynamic relationships for N-species (NH_4, NO_3, NO_2 and N_2), because oxidation of ammonium to nitrate and reduction of nitrate to nitrogen gas are usually kinetically controlled by irreversible reactions. Generally, redox reactions may have to be modelled as irreversible thermodynamic processes instead as reversible thermodynamic processes. Here, we should remember that kinetically controlled irreversible processes must always be thermodynamically feasible, even those that are microbially controlled. Bacteria may catalyse geochemical reactions but they cannot change the direction of geochemical reactions!

Inverse hydrogeochemical modelling uses existing data about aqueous systems and calculates the net mass transfer for a series of reactions proposed. Mass balance constraints are the only mathematical constraints that are explicitly included in inverse hydrogeochemical modelling. Kinetic, thermodynamic and mineralogical considerations are implicitly included via the modeller's concept of which reactions are involved. The input data will at least be chemical data about the aqueous compositions, but may include isotope data. The inverse problem is not constrained by thermodynamic considerations: the model's output is the net transfer of a master species into or out of the aqueous phase. This net mass transfer can be the difference between the dissolved amount of one mineral and the precipitated amount of another mineral. The modeller is thus responsible for controlling the thermodynamic feasibility. One way to do this is to perform a *post hoc* forward modelling for the initial composition, including thermodynamic constraints with respect to mineral equilibria, etc. The outcome of an inverse modelling can be a variety of possible reaction pathways. A major shortcoming of inverse models is that surface reactions are not adequately included in the mathematical model: the extent to which sorption reactions occur between two observation points is controlled by the total sorption capacity between these points and the selectivity coefficients, whereas the extent to which precipitation/dissolution can occur is independent of the amount of solid present as long as more is present than can be dissolved (see Section 2.2). Forward hydrogeochemical transport modelling circumvents this problem and as soon as sorption reactions are involved is thus actually superior to inverse hydrogeochemical modelling

2.5.2 Coupling of hydrochemical reactions

For groundwater chemistry it holds true that the behaviour of minor solutes is partly determined by the behaviour of major solutes. Two concepts deserve attention in how they interlink different types and reactions, and couple the behaviour of minor species and that of major species. They are the *common-ion effect* and *sequential equilibrium control*. The common-ion effect refers to two reactions in which a similar species is involved. For example, carbonate equilibrium and cation-exchange:

$$CaCO_3 + H \leftrightarrow Ca^{2+} + HCO_3^-$$

and

$$Ca{-}X_2 + 2\,Na^+ \leftrightarrow 2\,Na{-}X + Ca^{2+}$$

Calcium is the common ion for these two reactions and the two reactions are coupled to each other by means of the activity of Ca^{2+}. Combining these two reactions results in salinisation of an aquifer in which Ca-carbonate precipitates and freshening of the aquifer if Ca-carbonate dissolves.

Sequential equilibrium control is comparable to the common-ion effect. Now, a major species dictates the behaviour of a minor species in aqueous concentration by means of equilibria for two solid phases that have an identical co-ion. An example is the parallel behaviour of the weak carbonate and phosphate acids when equilibrium is attained for siderite and vivianite:

$$FeCO_3 + H \leftrightarrow Fe^{2+} + HCO_3^-$$

and

$$Fe_3(PO_4)_2 \cdot 2H_2O + 4\,H^+ \leftrightarrow 3\,Fe^{2+} + 2\,H_2PO_4^- + 2\,H_2O$$

The thermodynamic dissociation constants can be combined after eliminating the activity of Fe^{2+}, which is the identical co-ion. Combination results in:

$$\frac{[HCO_3^-]^3}{[H_2PO_4^-]^2} = \frac{K_{si}}{K_{viv}} \frac{[H_2O]^2}{[H^+]}$$

The behaviour of the carbonate species thus determines the phosphate concentration for such a system. The essential difference between the common-ion effect and sequential equilibrium control is that for the latter, major processes dictate minor processes and, associatedly, major solutes dictate minor or trace solutes, whereas for the former the processes balance each other.

2.5.3 Redox zoning

The redox status of an aquifer system deserves special attention for a combination of reasons:

- the types of minerals present are related to the prevailing redox conditions,
- degradation rates for microorganics such as pesticides often depend on the redox status, because these rates depend on the oxidants involved,
- aqueous speciation of redox-sensitive species such as arsene is related to the redox status.

The major redox-sensitive elements in groundwater systems are oxygen, hydrogen, nitrogen, manganese, iron, sulphur and carbon. The redox half-reactions for these elements in aqueous systems are:

$$O_2 + 4\,H^+ + 4e^- \leftrightarrow\ H_2O$$
$$NO_3^- + 6\,H^+ + 5e^- \leftrightarrow ½\,N_2 + 3\,H_2O$$
$$MnO_2 + 4\,H^+ + 2e^- \leftrightarrow Mn^{2+} + 2\,H_2O$$
$$Fe(OH)_3 + 3\,H^+ + e^- \leftrightarrow Fe^{2+} + 3\,H_2O$$
$$SO_4^{2-} + 10\,H^+ + 8e^- \leftrightarrow H_2S + 4\,H_2O$$
$$H_2CO_3 + 8\,H^+ + 8e^- \leftrightarrow CH_4 + 3\,H_2O$$
$$2\,H^+ + 2e^- \leftrightarrow H_2$$

in which e^- refers to an electron, which does not occur as free aqueous species in aqueous systems. Note that all reduction reactions consume protons and, consequently, all oxidation reactions produce protons. Also note that the oxidised forms of Fe and Mn occur in the solid state, which reflects the insoluble nature of these oxides at neutral pH.

Infiltrating groundwater at the water table is usually oxic and redox potential decreases in downstream direction due to the consumption of oxidants present by reductants present in the subsurface. When the supply of reductant is not extremely high the oxidants are consumed in order of decreasing energy yield. A characteristic sequence for the disappearance of dissolved oxidants and the appearance of dissolved reductants therefore develops along a flow line.

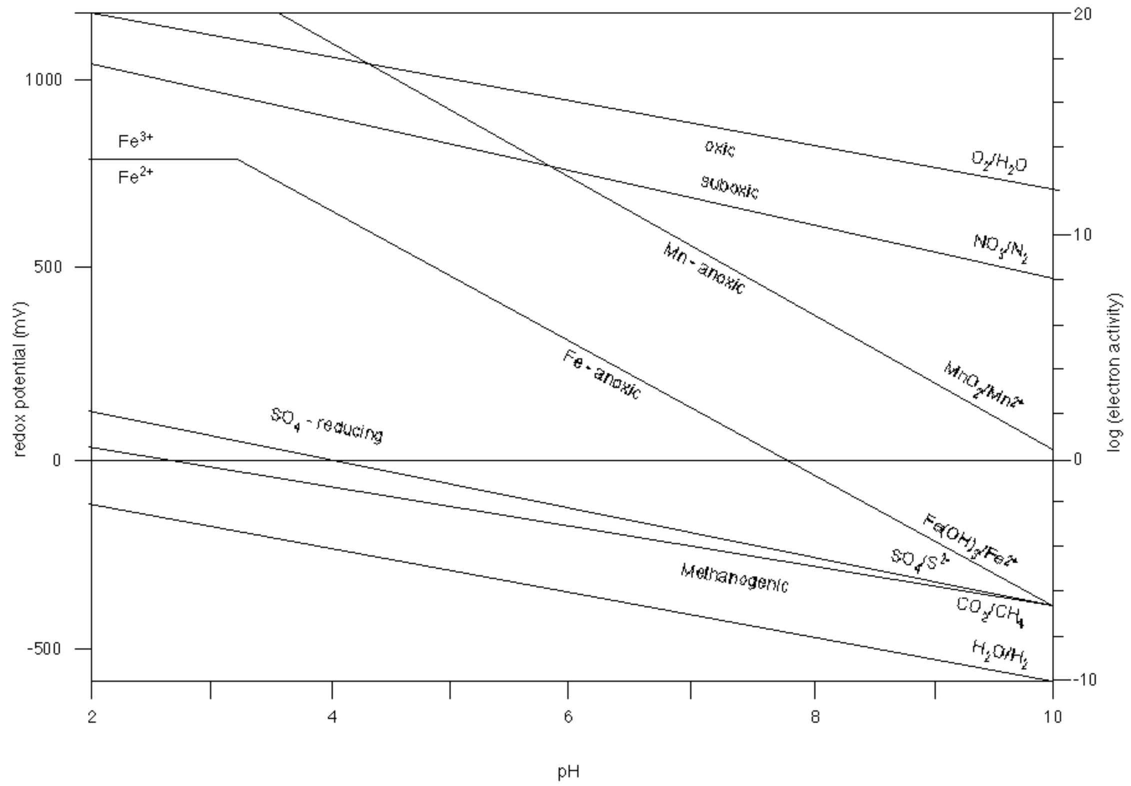
Figure 2.3 presents the redox clines for the above redox half-reactions in a pH-pe diagram. We can classify groundwater on its redox status based on the dissolved redox-sensitive species. Table 2.8 present a classification scheme that is valid for pH 7. When using this scheme note that deviations may occur at low pH when the Fe-redox cline and the N-redox cline lie close to each other. The Fe-redox cline is also sensitive to the type of Fe-oxyhydroxide present. Crystalline Fe-oxyhydroxides are much less soluble than amorphous ones. The redox cline shifts in response to lower redox potential for crystalline oxyhydroxides. Finally, the redox clines for S and C lie close to each other. Methanogenesis may happen before SO_4 reduction or parallel to it, depending on whether the reaction products are removed by secondary precipitation reactions (e.g. Fe-sulphide precipitation). This lowers the active concentrations of the products and keeps the forward redox reaction thermodynamically favourable.

Table 2.8. Redox classification of groundwater (x refers to a concentration above the detection limit, which usually lies around 1–10 µg/l)

Redox status	*O_2*	*NO_3*	*SO_4*	*Mn*	*Fe*	*H_2S*	*CH_4*
Oxic	x	x	x				
Suboxic		x	x				
Mn-anoxic1		x	x	x			
Fe-anoxic			x	x	x		
SO_4-reducing			x	x	x	x	
Methanogenic				x	x	x	x

1. Due to kinetic limitations NO_3 and Mn_{2+} are often jointly observed in the absence of both dissolved O_2 and Fe.

Figure 2.3 PH/redox potential diagram with redox clines for the major redox couples in groundwater



2.6 Groundwater supply and health

Groundwater is abstracted from aquifers for several reasons: drinking water for man or livestock, process water in food-, beverage- and table luxuries industries, irrigation water for agriculture, process water in industries, cooling water and other low-quality uses. In the long term technical problems may arise during the withdrawal of groundwater. The problems are a consequence of specific, reactive species in groundwater together with improper installation of the technical facilities. Some characteristics of malfunction are readily observable and guidelines for their possible causes and standard remediation are given in Table 2.9.

Groundwater is used as drinking water in many areas in the world. However, some groundwaters contain such high concentrations of species that their intake by man is undesirable. Toxic concentration levels present in groundwater not only interfere with human health directly via consumption of drinking water, but also indirectly via consumption of food (Fig. 2.4). Irrigation water is one of the factors which, together with geochemical and soil factors, determine the trace element composition of crops. Here, a particular problem is that some elements are more toxic to humans than to plants: in healthy plants these elements may accumulate to levels that threaten human health.

Mercury, lead and cadmium are particularly toxic to humans (mammalian toxic), whereas copper, nickel and cobalt are more toxic to plants (phytotoxic). The trace element composition of leguminous crops, pulses and cruciferous crops fluctuates more widely than that of cereal grains. The variability for a specific crop also varies for the different elements. No simple,

Table 2.9 Guidelines for remediation of groundwater withdrawal systems

	Observation	*Possible causes*	*Standard remedial practice*
i.	Rusty stains in clothes and/or plumbing; Water is red or brownish when tap is opened or after chlorination.	High concentration of dissolved iron	Water aeration and pH correction followed by settling and filtration
	Black stains	High concentration of dissolved manganese	Mn is more difficult to remove than Fe
ii.	Hard, white or brown encrustations in pipes and fittings	If soluble in HCl, caused by calcium carbonate or iron carbonate	Water softening, but may be expensive. Acidification is effective but may cause corrosion. Avoid the loss of dissolved gases, rapid cross-section reductions, excessive well drawdown. Periodic cleaning advisable. High water velocity and constant temperature delays the carbonate deposition. Removal of iron may be effective
	As above	If not soluble in HCl, combination of iron carbonate and silicates. High silica content (gypsum encrustations are rare)	Difficult to prevent. Avoid evaporation, temperature changes and sediments that can accelerate the deposition
iii.	Loose sediments and foreign matter in pipes, elbow bends and tanks	Particulate matter	Improve the well or correct. the screen and tube. Well completion or well development may have been inappropriate. Screen slots may be too wide or the gravel pack may not filter properly. Avoid intermittent operation as much as possible
	Clay-like matter	Clay	Improve the well or eliminate some portions of the screen. Prevent the entry of water that produces clay deflocculation
	Yellow to red flocculant or black filamentous organic matter	Iron hydroxide	Treat for iron as in part *(i)*
iv.	Metallic corrosion in pipes, fittings, and tanks	Acidic water, high sulphide content, high free CO_2	Correct the pH with lime

Table 2.9 (contd.)

Observation	Possible causes	Standard remedial practice
v. Corrosion and alteration of concrete, cement and rocks	Acidic water	Correct the pH with lime
	High salinity	Avoid evaporation and protect the surface from being alternately wet and dry
vi. Water turbidity	Clay	Correct the well. See item *(iii)*.
	High silica	Difficult to treat.
	Gas bubbles	Purge the pipes, avoid air suction
vii. Water taste:		
- Salty	Excessive salinity	Change water sources
- Rusty	Excessive iron and/or manganese	See item *(i)*
- Astringent	Excessive sulphate	Change water sources
- Earthy	High hardness	Softening
- Tasteless	Not fresh. Lack of dissolved oxygen	Aeration
- Tingling	High CO_s content	Aeration and/or pH correction.
- Disagreeable	Hydrocarbons, chlorinated, hydrocarbons, chlorophenols which may appear after chlorination	Adsorption on charcoal
viii. Water odour, most noticeable in low, unventilated areas:		
- Rotten egg	Sulphides	Aeration
- Petrol	Hydrocarbons	Adsorption on charcoal
ix. Water colour:		
- Reddish	Iron	Treat for iron as in item *(i)*. Avoid air. Purge the pipes
- Blackish	Manganese	Same as above
- Yellow	Hexavalent chromium in toxic quantities (greater than 0.5 mg/l)	Difficult to treat. Reducing agents and filtration
- White	Small air or gas bubbles	Store the water in a quiescent state
x. Warm water	Deep well and/or volcanic activity	Cooling tower
xi. Retarded plant growth, brown leaves	High salinity	Change water source
	Excess sodium	Add lime or gypsum
	Excess boron	Change water source
	Dissolved methane (rare)	Aerate

straightforward relationships are known between the composition of irrigation water and the trace element composition of crops.

The toxic concentrations in groundwater are the result of either natural processes or of anthropogenic contamination. Groundwater with natural above-normal concentrations typically occurs in areas with hydrothermal or volcanic activity and in areas with ore deposits.

The opposite of concentration excess is concentration deficiency: drinking water is an important source of various elements for humans, several of which are essential for human health. Essential ions for human health are Na, K, Ca, Mg, Cl, SO_4, HCO_3, NO_3, F, PO_4, Fe, Zn, Cu, Mn, I, Se, Co, Cr and Mo and possibly also Si, Li, B, V, Ni, As, Sr, Nb and Sn.

Shortage of iodine, selenium and fluorine has been reported to cause deficiency-related health problems. These problems arise in rural areas in developing countries, where the inhabitants depend on local food and small-scale drinking water facilities. Iodine deficiency is particularly associated with remote mountainous areas away from the sea.

Out of the two series of essential ions, Se, As, and Cr are also classified as 'characteristically hazardous' by the U.S. Environmental Protection Agency together with Ba, Cd, Pb, Hg and Ag. It should be realised that the toxicity of redox-sensitive elements depends on the redox status of that element. For example, $Cr(VI)O_4^{2-}$ is a human carcinogen, but Cr^{3+} is not. The World Health Organisation (1993) has defined guideline values for drinking water. Here, a distinction is made between species that have an adverse effect on health and species that may give rise to complaints from consumers.

Table 2.10 lists the guide values for inorganic constituents. Ideally, microorganic constituents should not be present.

Figure 2.4 Exposure route in groundwater of dissolved species harmful to man

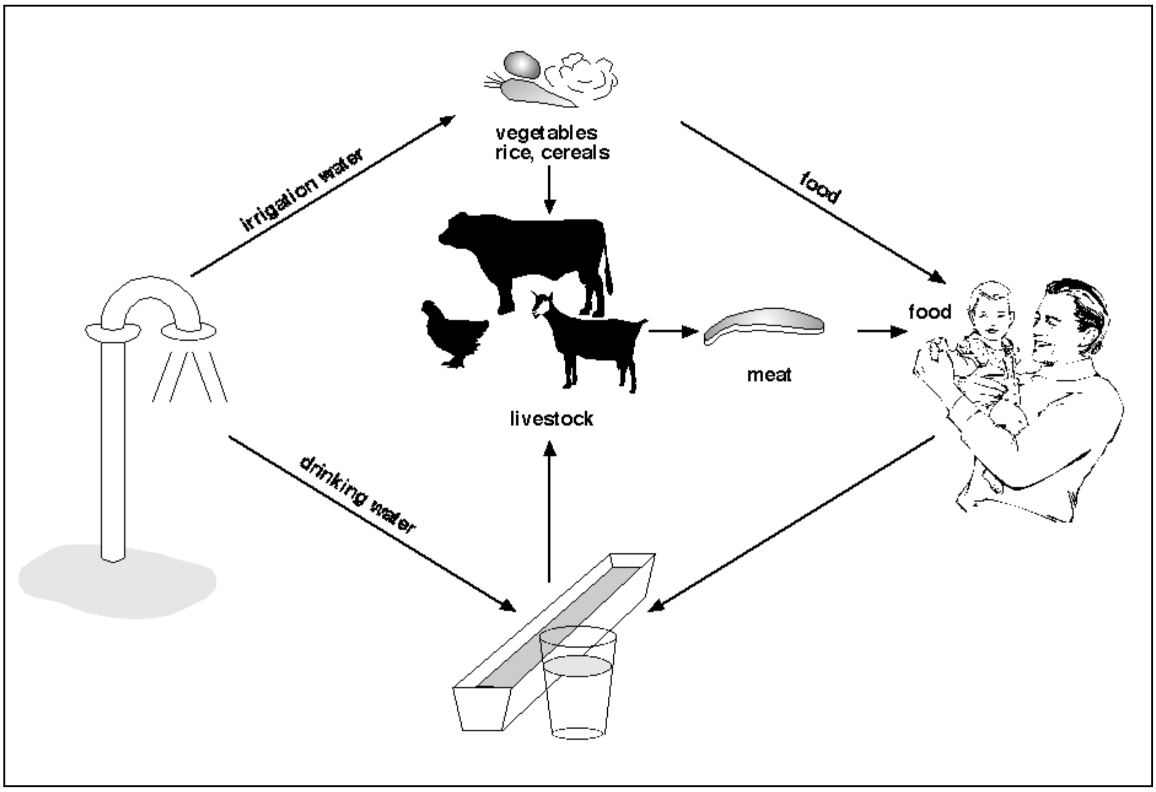


Table 2.10 WHO guideline values (in mg/l) for species in drinking water (1993)

Species	*Health significance*	*Species*	*Consumer complaints*
Antimony (Sb)	0.005 P[1]	Aluminium (Al)	0.2
Arsenic (As)	0.01 P	Ammonium (NH_4)	1.5
Barium (Ba)	0.7	Chloride (Cl)	250
Beryllium (Be)	NAD[2]	Iron (Fe)	0.3
Boron (B)	0.3	Sodium (Na)	200
Cadmium (Cd)	0.003	Sulphate (SO_4)	250
Chromium (Cr)	0.05 P	Sulphide (H_2S)	0.05
Copper (Cu)	2 P	Zinc (Zn)	3
Fluoride (F)	1.5	TDS	1,000
Lead (Pb)	0.01		
Manganese (Mn)	0.5 P		
Mercury (Hg) - total	0.001		
Molybdenum (Mo)	0.07		
Nickel (Ni)	0.02		
Nitrate (NO_3)	50		
Nitrite (NO_2)	3 P		
Selenium (Se)	0.01		
Uranium (U)	NAD		

1. P is provisional value.
2. NAD means no adequate data to permit recommendation of a health-based guideline value.

2.7 References and additional reading

ALLEY, W. M., (ed.), 1993. *Regional Ground-Water Quality*. Van Nostrand Reinhold, New York, 634 pp.

APHA-AWWA-WPCF. *Standard Methods for the Examination of Water and Waste Water.* American Public Health Association, Washington, D.C., (publication office).

APPELO, C. A. J.; POSTMA, D., 1993. *Geochemistry, Groundwater and Pollution*. A. A. Balkema, Rotterdam, 536 pp.

APPLETON, J. D.; FUGE, R.; MCCALL, G. J. H., (eds.), 1996. *Environmental Geochemistry and Health*. Geol. Soc. Spec. Publ. (113).

BACK, W.; BAEDECKER, M. J.; WOOD, W. W., 1993. Scales in Chemical Hydrogeology: A Historical Perspective; In: W. M. Alley (ed.), R*egional Ground-Water Quality.* Van Nostrand Reinhold, New York, pp. 111–29.

BACK, W.; FREEZE, R. A., (eds.), 1983. *Benchmark Papers in Chemical Hydrogeology*. Hutchinson and Ross, Stroudsburg, Pa, 416 pp.

CHAPELLE, F. H., 1993. *Ground-Water Microbiology and Geochemistry*. John Wiley and Sons, 424 pp.

CUSTODIO, E.; LLAMAS, M. R., 1976. *Hidrología subterránea*. Ediciones Omega, Barcelona, 2350 pp.

DEUTSCH, W. J., 1997. *Groundwater Geochemistry. Fundamentals and Applications to Contamination*. Lewis Publ., Boca Raton, 221 pp.

DOMENICO, P. A.; SCHWARTZ, F. W., 1990. *Physical and Chemical Hydrogeology*. John Wiley and Sons, 824 pp.

DREVER, J. I., 1997. *The Geochemistry of Natural Waters. Surface and Groundwater Environments.* Prentice Hall, Upper Saddle River, NJ, 436 pp.

FETTER, C. W., JR., 1994. *Applied Hydrogeology*. Prentice Hall, Upper Saddle River, NJ, third edition, 691 pp.

——, 1993. *Contaminant Hydrogeology*. Macmillan Publishing Co., New York, 458 pp.

FREEZE, R. A.; CHERRY, J. A., 1979. *Groundwater*. Prentice Hall, 604 pp.

HEM, J. D., 1985. *Study and Interpretation of the Chemical Characteristics of Natural Water*. U.S. Geol. Surv. Water Supply Paper 2254, third edition, 263 pp.

LANGMUIR, D., 1997. *Aqueous Environmental Geochemistry*. Prentice Hall, Upper Saddle River, NJ, 600 pp.

LICHTNER, P. C., STEEFEL, C. I.; OELKERS, E. H., 1996. Reactive Transport in Porous Media. R*eviews in Mineralogy*, Vol. 34. Miner. Soc. of America, Washington, D.C., 438 pp.

MATTHESS, G., 1982. *The Properties of Groundwater* (translated from the original German by John C. Harvey), Wiley-Interscience, 406 pp.

Parkhurst, D. L.; Plummer, L. N., 1993. Geochemical Models. In: W. M. Alley (ed.), *Regional Ground-Water Quality.* Van Nostrand Reinhold, New York, pp. 255–94.

PLUMLEE, G. S.; LOGSDON, M. J., (1999). The Environmental Geochemistry of Mineral Deposits. Part A: Processes, Techniques, and Health Issues. *Reviews in Economic Geology,* Vol. 6A. Soc. of Econ. Geologists, Littleton, CO, 371 pp.

SCHOELLER, H., 1962. *Les eaux souterraines*. Mason et Cie, Paris, France.

THURMAN, E. M., 1985. *Organic Geochemistry of Natural Water*. W. Junk Publishers, 497 pp.

WORLD HEALTH ORGANISATION, 1993. *Guidelines for Drinking-Water Quality. Volume 1. Recommendations*. WHO, Geneva, second edition, 188 pp.

3 Groundwater networks and observation methods

3.1 Introduction

A groundwater monitoring network is an organised system for the continuous or frequent measurement and observation of the actual, dynamic state of the underground environment, often used for warning and control (definition adapted from UNESCO, 1992). Monitoring of groundwater is essential for the characterisation of groundwater systems and a key activity in defining sustainable environmental development of such systems. Hydrological processes are highly variable in space and time making data collection over all scales difficult and expensive, but groundwater systems can only be identified if ample hydrological data in space and time have been collected.

To understand the prevailing system in a region both physical and chemical data need to be systematically collected.

There should be no substantial use of groundwater resources (e.g. for drinking water, irrigated agriculture, industry) unless its exploitation is based on a sound groundwater management plan, which is supported by a continuous effort of collecting hydrologic data from observation networks. Then, possible detrimental effect of exploitation, such as the drying up of springs or ecologically relevant wetlands, as well as the impact of human activity on the groundwater resources (e.g. pollution or recharge reduction) can be recognised at an early stage of the groundwater development.

Groundwater monitoring networks have been extensively developed in the humid-temperate, mid-latitude, industrialised countries, although methods for the optimal design of networks are still under development. Furthermore, more sophisticated observation instruments are being developed to reduce monitoring costs or to collect more specific data. On the other hand, in many developing countries data from groundwater monitoring networks are woefully lacking, despite the great need for these data to manage the sometimes scarce groundwater resources and the high water demand resulting from population increases. This is why UNESCO promotes the elaboration of groundwater monitoring procedures, including the design and analysis of networks and the description of observation methods, in the framework of the International Hydrological Program (IHP-UNESCO, 1990).

The first operational procedures for hydrological data collection in networks were proposed by Langbein (1954). Over the years many good reviews of groundwater monitoring networks and observation methods have been published. Brown et al. (1972), Bachmat (1989) and WMO (1994) give detailed guidelines on the design of groundwater monitoring networks and the observation methods to be used. Other relevant information is provided by Moss (1982; 1986), Nielsen (1991) and OPEA (1993). It is not the intention of this chapter to cover in full all the basic aspects that are well documented in the publications mentioned above. In this chapter a summary and an update is presented, which is mainly based upon a recent UNESCO document (Van Lanen, 1998).

3.2 Place of monitoring networks in groundwater management

Groundwater management should rely on hydrological data of the actual state of the groundwater system. In this context two different types of groundwater monitoring activities should be distinguished, namely background monitoring and specific monitoring.

3.2.1 Background monitoring networks

Background monitoring, or primary, networks are introduced because of the need to start monitoring before significant human interference occurs. Background monitoring is applied to large areas without significant human interference (low technology). Its objective is to identify the actual state of the aquifer system. It establishes the initial conditions prior to significant groundwater development and gives background data for future discussions on possible overexploitation.

Before any groundwater monitoring can start, the objectives to be derived from the national land and water policy must be clear. In the upper block of Figure 3.1 the main steps of the pre-monitoring investigations are presented.

Before a monitoring system can be designed , the groundwater flow system in the area must be understood and described as far as existing data allow. These activities start with the collection of time-independent data, e.g. about geological framework, hydraulic properties, boundary conditions and hydrochemistry resulting in a definition of the prevailing groundwater flow systems. It is likely that at the start of the background monitoring phase knowledge is still insufficient for a full understanding of groundwater systems, because an objective of this monitoring is to improve the knowledge about the prevailing groundwater systems, but similar regions may have been investigated, which might allow transfer of knowledge to the region to be monitored (Fig. 3.1).

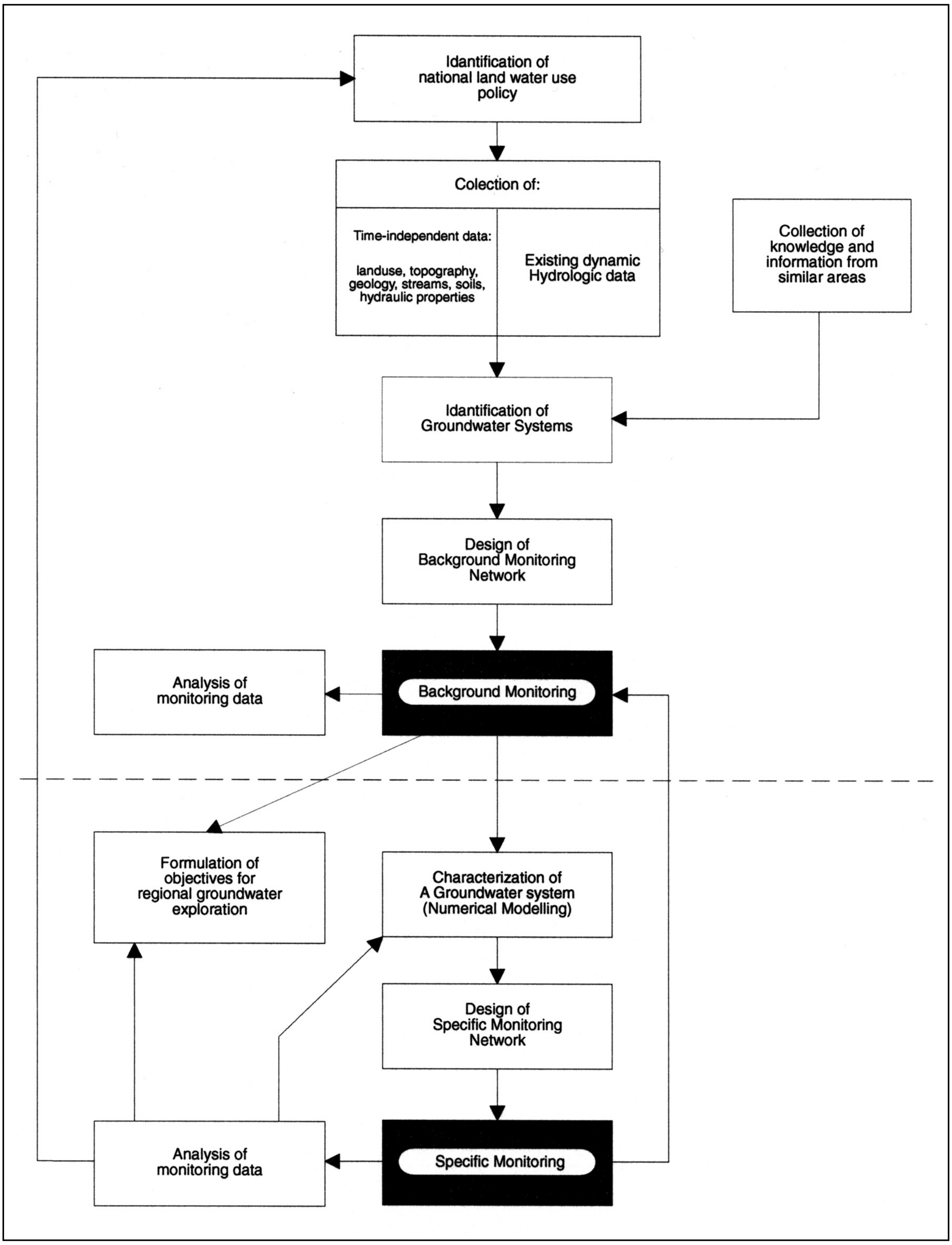
After the system identification phase, a first version of the groundwater monitoring network is designed (middle block of Figure 3.1). Systematic measurements of heads and chemical composition of groundwater in existing and abandoned wells should start. The groundwater monitoring should be supported by the collection of meteorological data, data on the vadose zone, and spring flow and stream flow data in similar networks. After the first or second year of systematic data collection, the initial version of the background monitoring network should be thoroughly analysed. The earlier defined version of the groundwater system and associated conceptual model should be refined. Statistics and geostatistics can help to improve the sampling density and frequency. It is likely that the analysis based on the improved knowledge of local hydrological phenomena of the region itself will lead to the first version of the monitoring network being refined to provide better representation of the specific characteristics of the groundwater systems in the area of interest (Fig. 3.1).

The refinement of the background monitoring network should be a continuous process of analysing incoming data, refinement of the description of the groundwater system and subsequent modification of the network. It is likely that the network will need to be modified more than once, especially in regions with extreme meteorological conditions, such as the arid and semi-arid or arctic areas. Background monitoring in these areas is a long-term effort.

3.2.2 Specific monitoring networks

Specific or secondary monitoring networks follow what happens in the underground environment when it is substantially exploited for particular purposes or when ecologically relevant wetlands, springs or streams are expected to be affected or polluted by human activities. Specific monitoring characterises the transient state of the aquifer and acts as an early warning for over-

Figure 3.1 General layout of the groundwater monitoring procedure (Van Lanen, 1998)

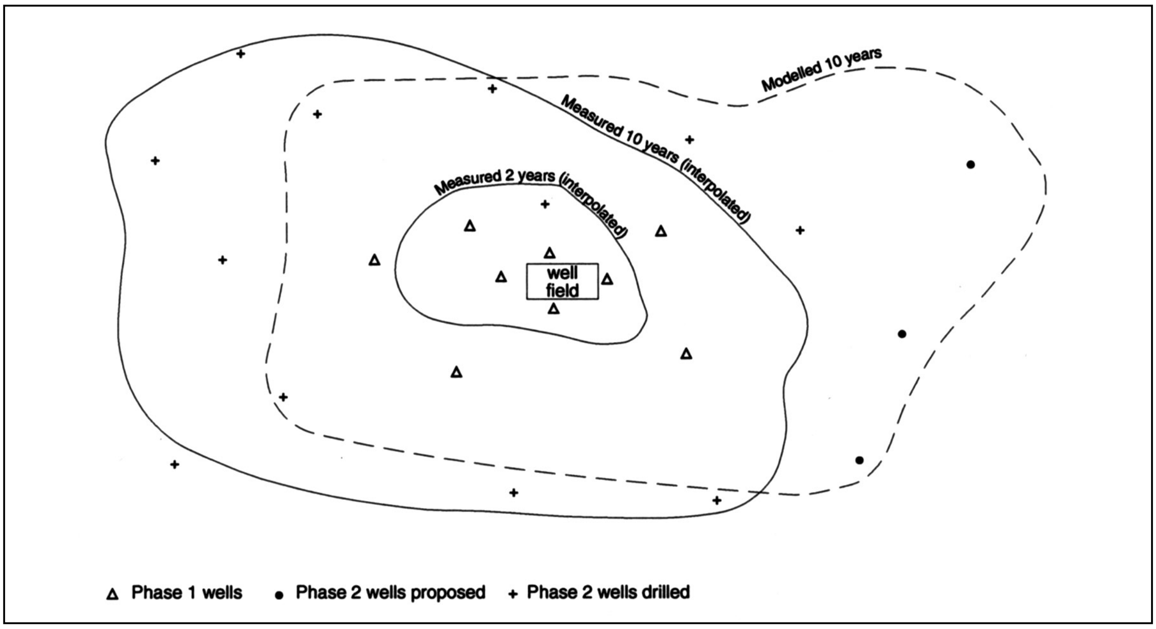


exploitation or pollution. The monitoring is restricted to those areas where effects are expected; for example, because of significant abstraction and a more accessible aquifer. Potential problems are the reduction of spring flow, falling groundwater heads, falling well yields, deterioration of water quality (including sea water intrusion or upconing of brackish or salt water) and land subsidence.

A specific monitoring network should be set up after it has been decided to develop groundwater resources in a particular region or when contamination of the water resources is expected. The specific network should be designed on the basis of a comprehensive analysis of background monitoring data. The conceptual model of the groundwater flow system is replaced by a numerical one (lower block of Figure 3.1). This groundwater simulation model must specify the consequences of different abstraction scenarios or remediation measures in terms of groundwater heads, groundwater flow lines, residence times, changes in recharge conditions (e.g. induced recharge), chemical composition, and spring flow and streamflow. Subsequently, the specific monitoring aspects can be formulated, e.g. the boundary of the affected area, type of hydrological variables to be monitored, and sampling density and frequency. A specific groundwater monitoring network is likely to require modification if the collected data show that the response of the groundwater flow system differs from the simulated one. An example is presented in Figure 3.2. Similar to background monitoring, specific monitoring needs continuous efforts in terms of the collection of data, data analysis and refinement or redefinition of the monitoring network.

The specific monitoring should be integrated into the background monitoring efforts (Fig. 3.1). Eventually the ideal situation is to have a background monitoring network that covers the whole country and specific monitoring networks in regions where groundwater resources are significantly exploited or where pollution occurs (nested monitoring networks).

Figure 3.2 Adaptation of the first version of a specific monitoring network after differences were observed between monitored and modelled effects (Lloyd, 1998)



3.3 Hydrological variables

The essence of a groundwater monitoring network is the measurement and observation of the state variables of the groundwater body itself. The dynamic state of the groundwater system can be monitored through observing:

- depth to water table (x, y, t)[1];
- chemical composition of phreatic groundwater (x, y, z, t);
- piezometric heads (x, y, t) of each aquifer, or
- heads (x, y, z, t) of each aquifer unit if vertical differences occur;
- chemical composition of deep groundwater (x, y, z, t) in each aquifer;
- spring flow (x, y, t);
- well yields (x, y, t); and
- chemical composition of abstracted groundwater (x, y, t).

A first analysis of the chemical composition should preferably be carried out in the field. Standard measurements that characterise the physical-chemical composition of groundwater, e.g. temperature, dissolved oxygen (DO), redox potential (Eh), hydrogen concentration (pH) and electrical conductivity (EC).

Furthermore, in a multiple aquifer system attention should also be paid to vertical differences, both in aquifers and aquitards. Carrillo-Rivera (1998) showed that in Mexico City subsidence caused by groundwater abstraction cannot be adequately understood unless the vertical head differences in the upper aquifer are monitored.

Groundwater monitoring networks need to be complemented by other networks monitoring streamflow and hydrological variables for the determination of groundwater recharge (Fig. 3.3). Unless this is done, the state of a groundwater system cannot be understood adequately. WMO (1994) systematically presents acquisition and analysis techniques for precipitation, evapotranspiration and soil moisture to determine groundwater recharge. Lerner et al. (1990) and Simmers (1997) give an extensive review of recharge estimating techniques and associated collection of hydrological data for different types of recharge (e.g. precipitation recharge, river recharge). A review of methods to measure streamflow is presented by Boiten (1993) and WMO (1994). In Figure 3.4 the average results of integrated long-term monitoring and subsequent modelling are presented for three different catchments in West and Central Europe. These results were used to identify the prevailing groundwater systems.

The Hupsel catchment in the Netherlands has a small groundwater storage and a quick response of the streamflow to excess precipitation. A pronounced summer drought occurs. The Gulp basin consists of chalk with a large groundwater storage; the streamflow shows a smoothed and delayed response on excess precipitation. The Cerná Desná catchment, which is located in the Jizera Mountains in the Czech Republic, has a small groundwater storage (weathered granite overlying fractured granite) but a relatively constant discharge because of the evenly distributed precipitation. The high peak in May-June is from snow melting.

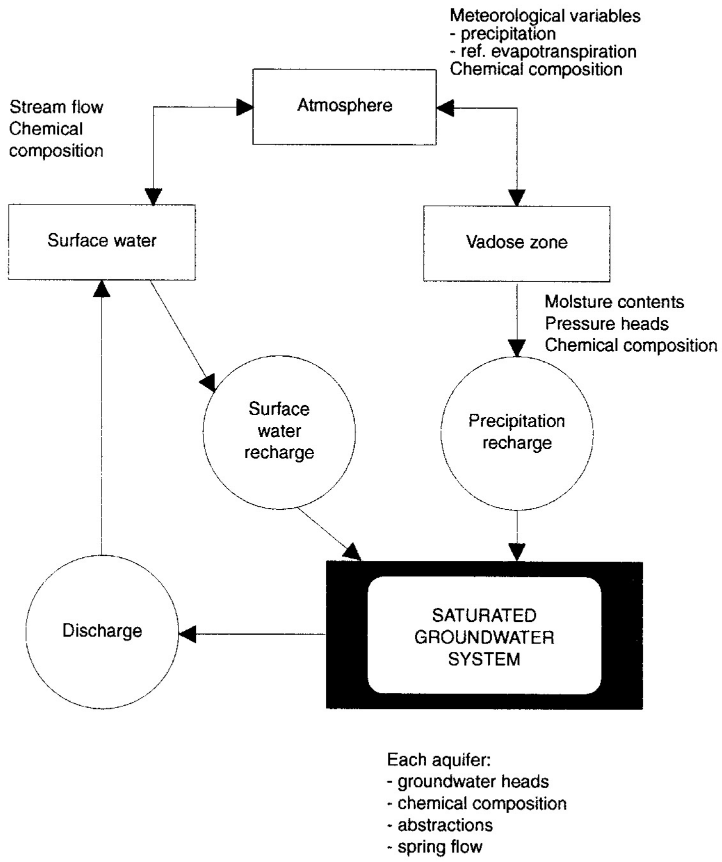
The meteorological, vadose zone, groundwater and streamflow monitoring networks should be integrated from the outset (Moss, 1986). Commonly, more than one organisation is responsible for the acquisition of the data so coordination must be good. For example, locations, monitoring frequency, accuracy, data processing, data transfer and publication, in the different networks need to be harmonised.

3.4 Network design

Ideally the design of network density and sampling frequency would be based on an optimisation of the cost of monitoring and of the accuracy of collected and derived data related to the objectives of the network. Without a thorough understanding of the hydrogeological setting of a region, there is little chance that a network would produce adequate information. At the start of monitoring in a region, however, a classic problem is insufficient hydrogeological knowledge

1. x, y, z, t denotes position in space and time.

Figure 3.3 Hydrological variables to be monitored in a groundwater network and variables of other networks which are complementary but necessary



and therefore an unknown spatial and temporal variability for each variable to be monitored, although we know that groundwater heads and chemical components are spatially and temporally correlated.

Occasionally, lack of prior hydrogeological knowledge on how to interpolate (in space and time) between measurement points hampers even the beginning of monitoring. If hardly any data are available, background monitoring should start anyway, by designing a network based on the few existing data and on expert knowledge from similar regions. After some years of data collection, statistical and geostatistical techniques can be applied to explore the spatio-temporal structure of each hydrological variable in the region of interest (Fig. 3.5). Eventually, optimisation theory and socio-economic analysis can be used in decision-making procedures to propose optimal networks to the policy makers (WMO, 1994). The following sections on network density and sampling frequency assume that sufficient data are available for statistical analysis.

3.4.1 Network density

The effectiveness of a groundwater network in terms of network density is often defined as the accuracy of the spatial interpolation, i.e. the standard deviation of the spatial interpolation error.

Figure 3.4 Measured and simulated hydrological variables. Average measured precipitation and streamflow, and simulated real evapotranspiration and streamflow (BILAN; Kašpárek and Krejcová, 1994) for a Dutch, a Belgium-Dutch and a Czech catchment

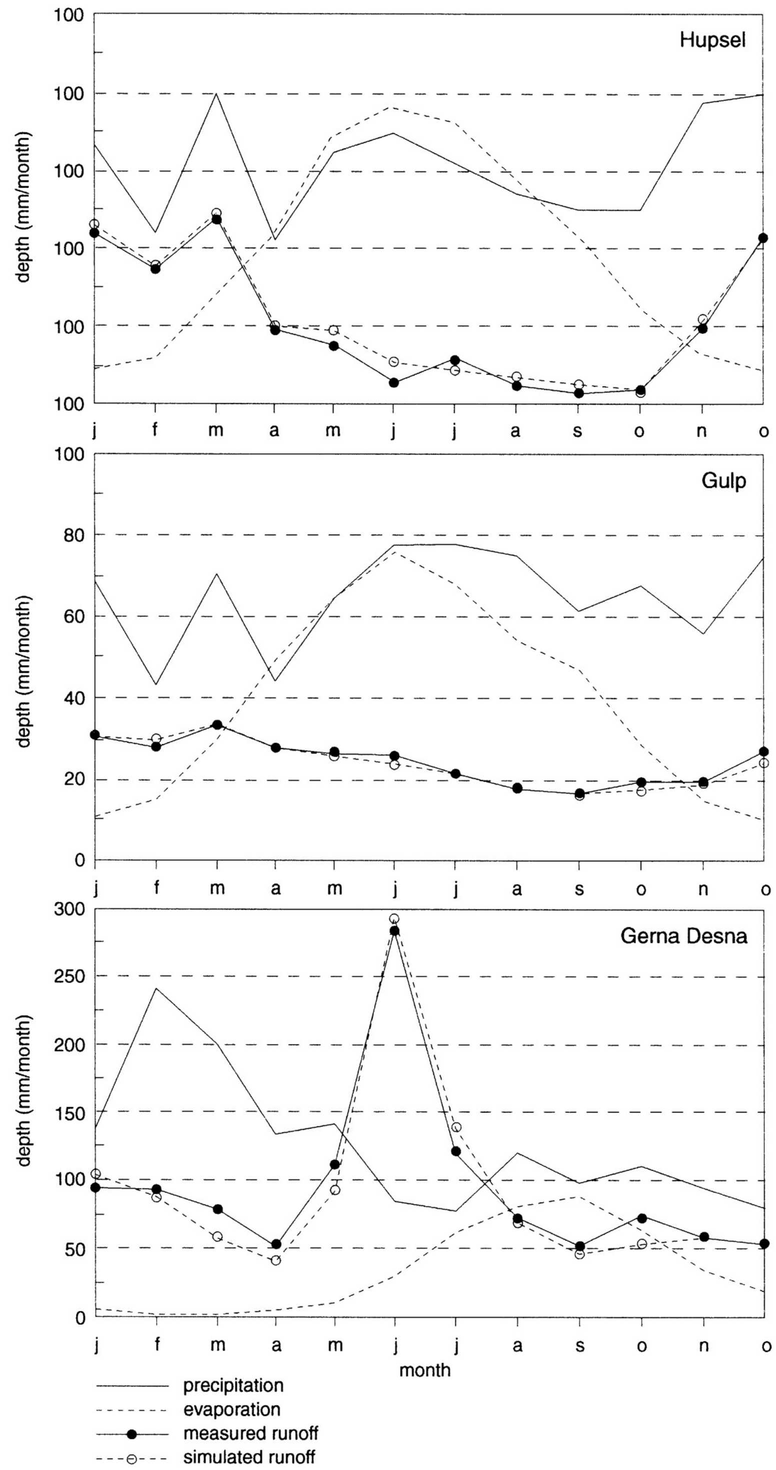
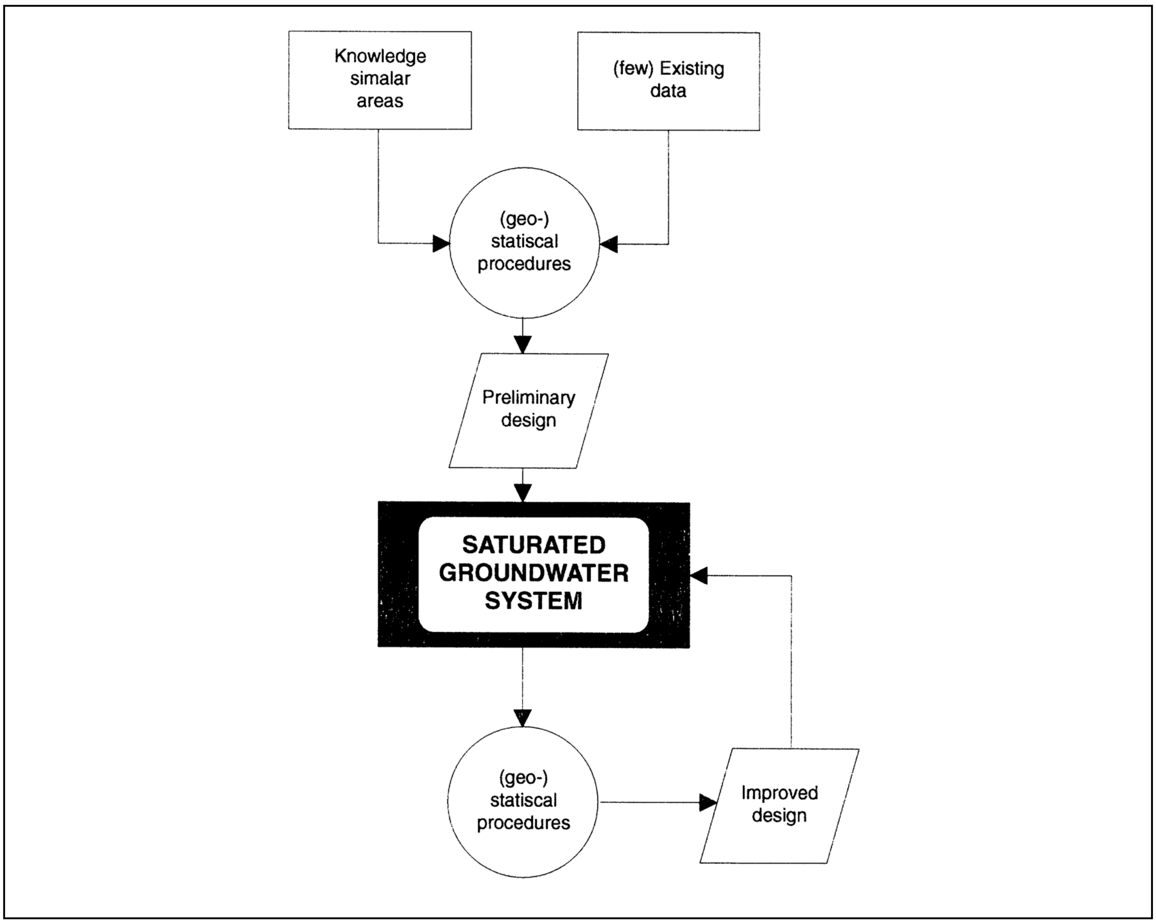


Figure 3.5 Procedure for the determination of network density and sampling frequency



Therefore a spatial interpolation technique is required that not only estimates the groundwater variables, but also provides the standard deviation of the estimation error (e.g. Zhou, 1992a). Kriging is a well known and suitable technique for such a purpose (e.g. Marsily, 1986).

Kriging is a method for estimating the value of a regionalised random variable (e.g. groundwater head or chemical component) at any point that has not been measured from a set of measurements at different locations. The semi-variogram plays a key role in the kriging procedure. It describes the spatial correlation structure of the regionalised variable, i.e. it shows that observations closer to each other are likely to be more similar than observations further away.

The semi-variogram is determined as follows (e.g. Zhou, 1992a):

- Several groups (m) of distances between measurement locations with an average distance d are defined. For example, d_1 is the group where the distances between the locations are small, whereas d_m is the group with the largest distances;
- For each distance group d_k the possible pairs of measurements locations i and j are identified (n_k: all possible pairs);

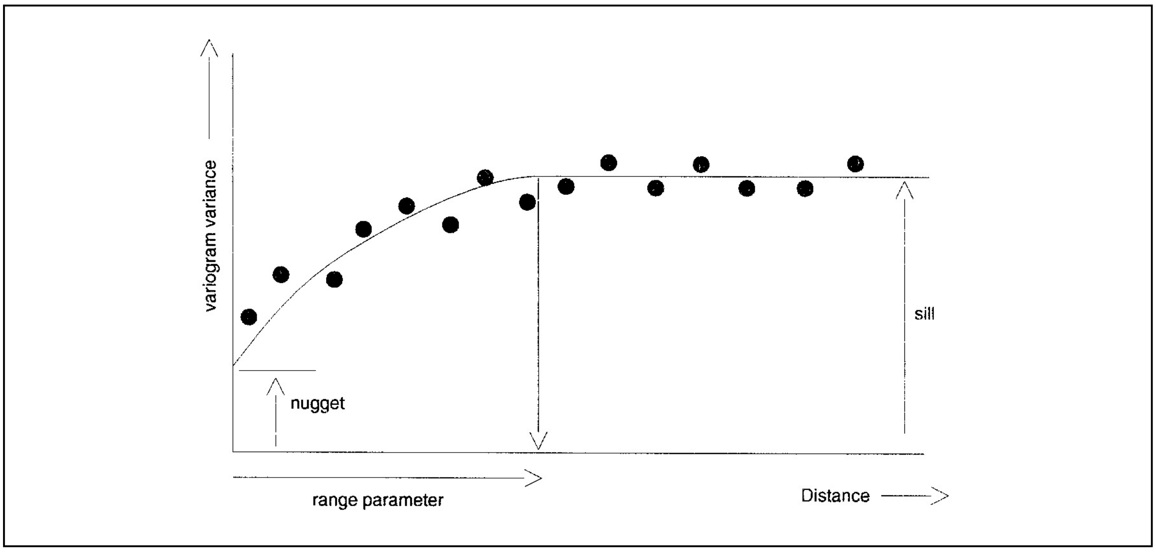
For each distance group d_k the variogram value $\gamma'(d_k)$ is calculated using the following equation:

$$\gamma\,(d_k) = \frac{1}{2\,n_k}\sum_{1}^{k}\,(Z_i - Z_j)$$

where Z_i and Z_j the measured variables are at location i and j.

The semi-variogram is obtained by plotting the $\gamma'(d_k)$ values against the distances of the various groups. This semi-variogram is called the experimental one (dots in Fig. 3.6);

Figure 3.6 Experimental (dots) and theoretical semi-variogram (solid line)



Finally a theoretical semi-variogram is fitted through the experimental one (solid line in Fig. 3.6) by using common models such as an exponential, Gaussian or spherical model with three parameters, i.e. the nugget variance c_0, the range parameter b and the sill A_0. The nugget variance describes the non-spatial variance, i.e. the variance of the measurement error, observer error or micro-distance variability. The range parameter or correlation length and the sill describe the spatial variation. The variation increases with increasing distance (Fig. 3.6) between locations in space. The value of the range parameter is the variability that eventually occurs (A_0) when the distances between the observations increase.

In a heterogeneous region (e.g. geology, surface water system) it might be useful to find out if a stratification of the region leads to better quality semi-variograms. Stratification implies dividing the region into disjoint subareas (strata) with more homogeneous characteristics. To model the spatial variability for each stratified area, a within-strata semi-variogram needs to be determined. For a within-strata semi-variogram, only pairs of points belonging to the same stratum contribute to the semi-variogram. If the hydrological variable is related to the stratification, the within-strata semi-variogram will take lower values than the ordinary semi-variogram, because the within-strata variability is less than the overall variability. However, if this relation is not present, it will show higher variability, and hence stratification is useless.

If the semi-variogram is known, it is only a simple routine to compute the hydrological variable on the nodes of a specified network, including the standard deviation of the interpolation error (*SDIE*), from the locations with measurements. In the subsequent network density design the *SDIE* plays a key role. For example, the following procedure can be applied (Van Bracht and Romijn, 1985):

- divide the region into homogeneous subareas (e.g. geology, water table depth etc.) also considering available observation wells per subarea. Regional hydrogeological knowledge is a prerequisite for defining an adequate stratification;
- determine within-strata semi-variogram for each Subaru for a particular date and check persistence for other dates (e.g. Stein, 1998);

- Calculate the network density graphs for each Subaru using the within-strata semi-variogram (Fig.3.7). The graph specifies the relation between the *SDIE* and the number of observation wells per km^2 and is derived by calculating the hydrological variable on the nodes of a specified network from a decreasing number of locations in the Subaru for which measurements are available. The Subaru's sensitivity to decreasing the density of sampling points and the SDIE depend on the characteristics of the Subaru's. For instance, in a clay polder area with controlled water levels (subarea B) the *SDIE* and the sensitivity are significantly lower than in a sandy area with natural drainage and with a shallow impermeable base (subarea A).
- For each subarea determine the allowable *SDIE* that can be related to the priority of the subarea in terms of groundwater management. Van Bracht and Romijn (1985) derive this priority from planned groundwater exploitation and the sensitivity of the subarea to lowering of the water table. Some fictitious results of this step are illustrated in Figure 3.8. First the priority of each area is converted in an allowable *SDIE*; the allowable *SDIE* increases from 40 to 80 cm for subareas 5 to 2 (black bar in Fig. 3.8) corresponding to a decreased priority. Then theoretically, the network density could be derived from the network density graphs (Fig. 3.7). However, in some subareas the allowable *SDIE* is even lower than the one belonging to the highest density in the network density graph (subarea 1), whereas in others (subarea 4) the allowable *SDIE* is still higher than the one belonging to the lowest density. Therefore a maximum and minimum network density are introduced; in this example, one well per 2 and 10 km^2, respectively.

  The *SDIE* belonging to these densities is derived from Figure 3.7 and plotted in Figure 3.8 as the maximum and minimum (first and second bars). Now the network density can be determined for each subarea. In subareas 2 and 4 the minimum density can be used, i.e. one well per 10 km^2 (1/10), whereas in subarea 1 the maximum density

Figure 3.7 Network density graphs; relationship between standard deviation of the interpolation error (SDIE) and density of observation wells

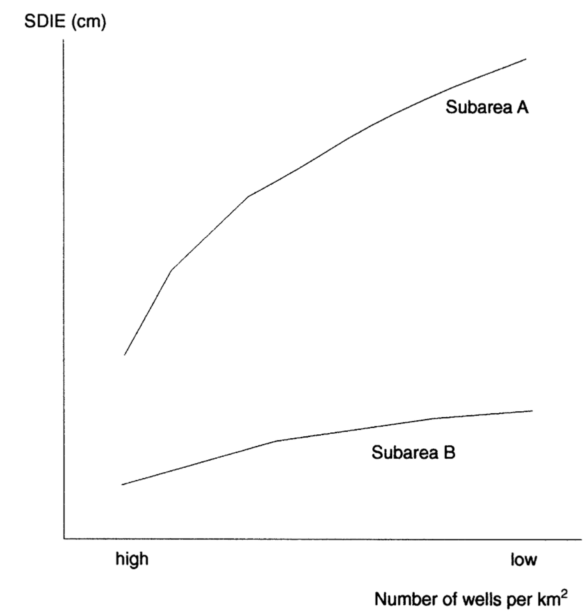
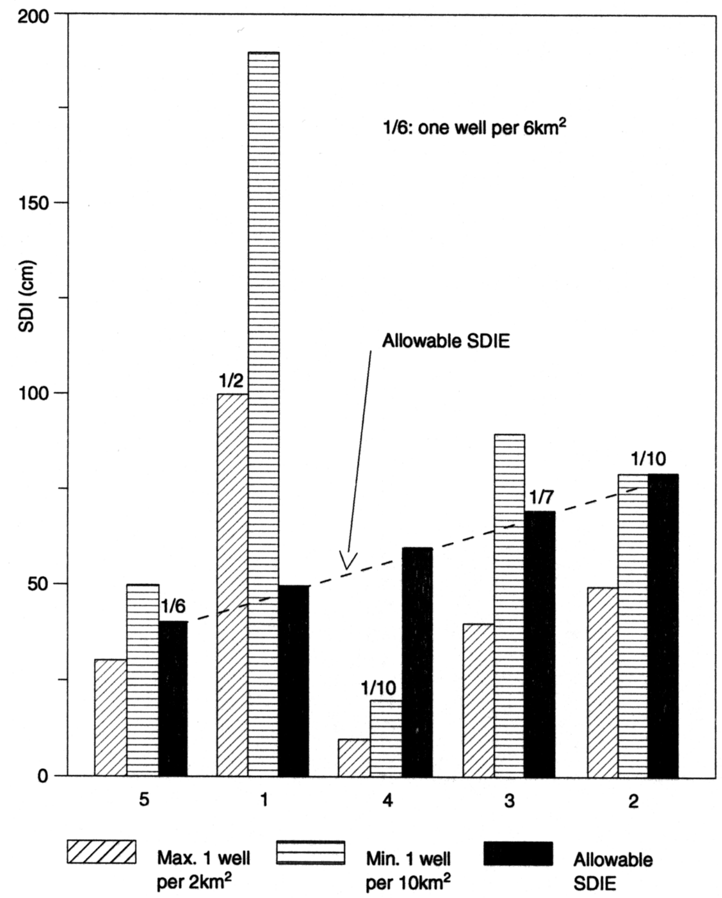


Figure 3.8 Network density derived from allowable standard deviation of the interpolation error (SDIE), minimum and maximum density for different subareas



is used (1/2), although the allowable *SDIE* is exceeded. In the subareas 3 and 5 the network density is between the prescribed minimum and maximum density, namely 1/7 and 1/6, respectively.

- Perform a sensitivity analysis (monitoring alternatives); i.e. specify other values for allowable *SDIE*, and for minimum and maximum network density, and determine network density for each subarea and total network density. Van Bracht and Romijn (1985) show that in their example reducing the maximum allowable *SDIE* from 65 to 60 cm entails increasing the number of observation wells by 33%. A 10 cm decrease from 65 to 55 cm implies a 77% increase in wells.
- Compare cost and benefits of the different monitoring alternatives and eventually select optimal network density.

Stein (1998) also gives an example of a redesign of a groundwater network using kriging. He shows that in an area of 635 km^2, where approximately 500 wells occur, the uncertainty can even decrease if 50 existing wells are removed and 5 new ones are introduced at strategic locations.

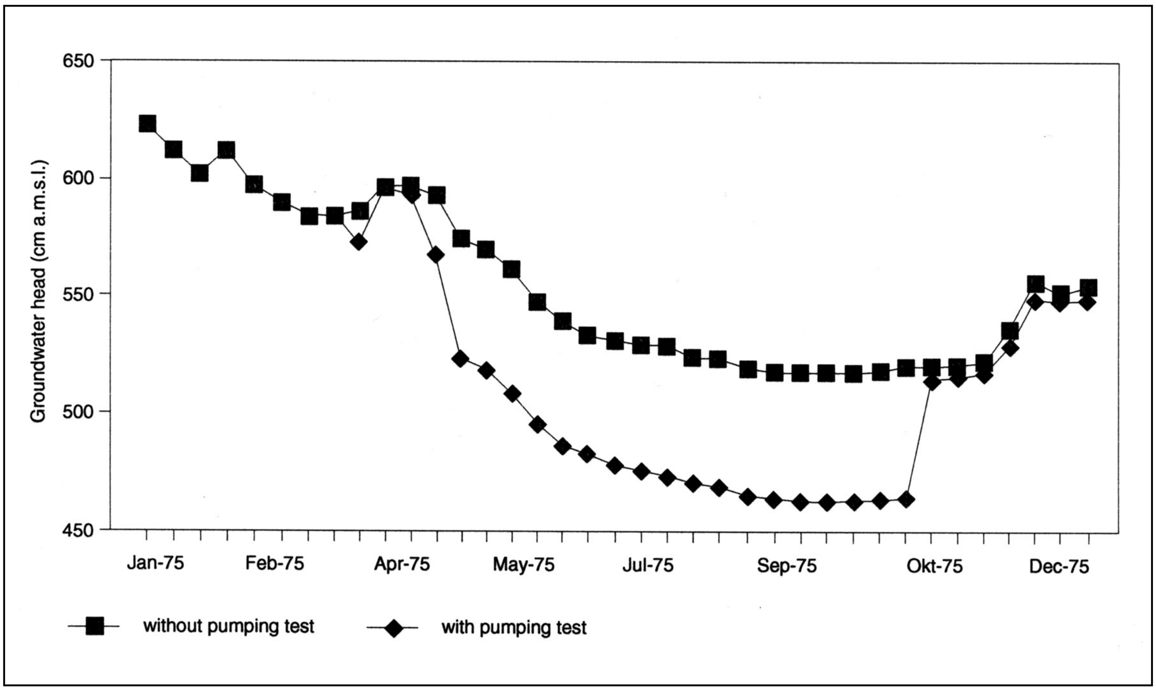
3.4.2 Sampling frequency

Besides exhibiting spatial variability, groundwater heads and chemical components show a temporal variability, which introduces the question of how often a variable has to be monitored. Temporal variability can be analysed with time series analysis procedures (e.g. Box and Jenkins, 1976; Chatfield, 1989; WMO, 1994), which give information about trends, periodic fluctuations and the mean. Detection of trends and periodic fluctuations is relevant for monitoring the effects of groundwater abstraction, seasonal climatic effects, or the deterioration of the groundwater quality. Sometimes, use of time series analysis is hampered by incomplete data series, irregularities in sampling intervals, and shortness of the time series.

The analysis of groundwater time series is confronted with some special properties of the groundwater system, e.g. the groundwater head or chemical composition at time *t* is dependent on previous values at time *t–1* (autocorrelation), and non-stationarity due to trends and periodic fluctuations. If these features are recognised the appropriate advanced time series analysis techniques can be applied to evaluate the time series and subsequently to design the sampling frequency (e.g. Zhou, 1992a; Zhou, 1992b). The following steps can be recognised:

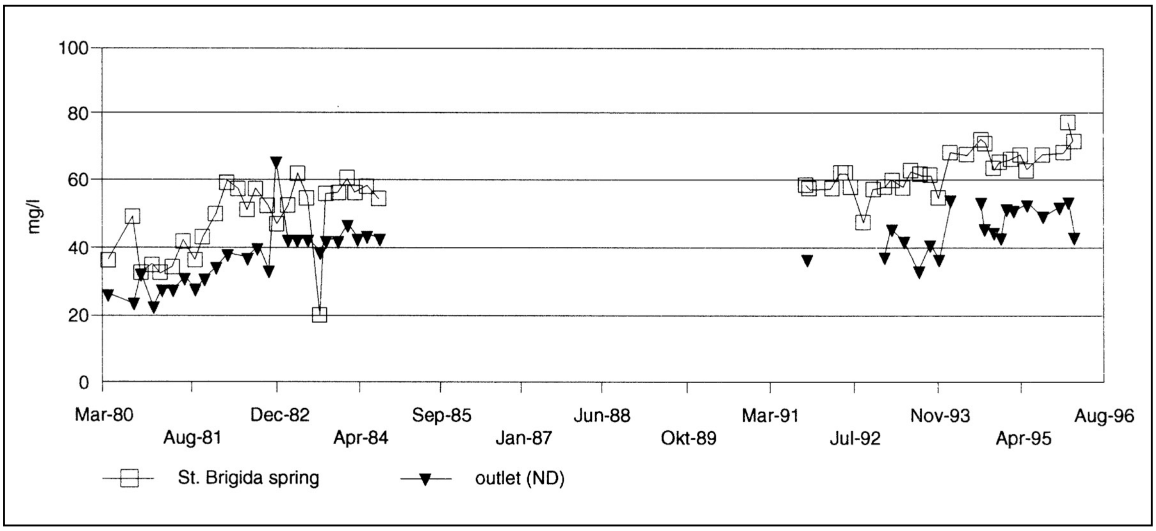
- Analysis of characteristics of groundwater time series. First possible *trends* in time series of groundwater data need to be investigated. The probability of trend detection depends on trend magnitude, correlation structure of the time series, the observation period, and the required reliability of the statistical test. A plot of the data is likely to show the presence of an obvious trend and if so, which type of trend prevails (step, linear, or exponential trend).
- Figure 3.9 shows an example of step trends of about 50 cm in groundwater heads of a semi-confined aquifer, due to the seasonal groundwater abstraction. Step trends are easiest to recognise by comparing the means of groundwater heads or other com-

Figure 3.9 Step trends in the groundwater heads because of a seasonal groundwater abstraction in a semi-confined aquifer



ponents from the periods before and after a known or expected change in the groundwater system. The detection of a step trend is simply a statistical test, where the hypothesis that the means of the time series are equal or unequal either is accepted or rejected. Student's t distribution is often used. Nitrate concentrations in the groundwater discharge of the Noor catchment (Belgium-The Netherlands) show a linear trend (Fig. 3.10). Linear trends are commonly evaluated by using a classical linear regression model. The detection of the linear trend is a statistical test against the slope of the trend (Zhou, 1992a).

Figure 3.10 Nitrate concentrations in the groundwater outflow of the Noor catchment showing a linear trend

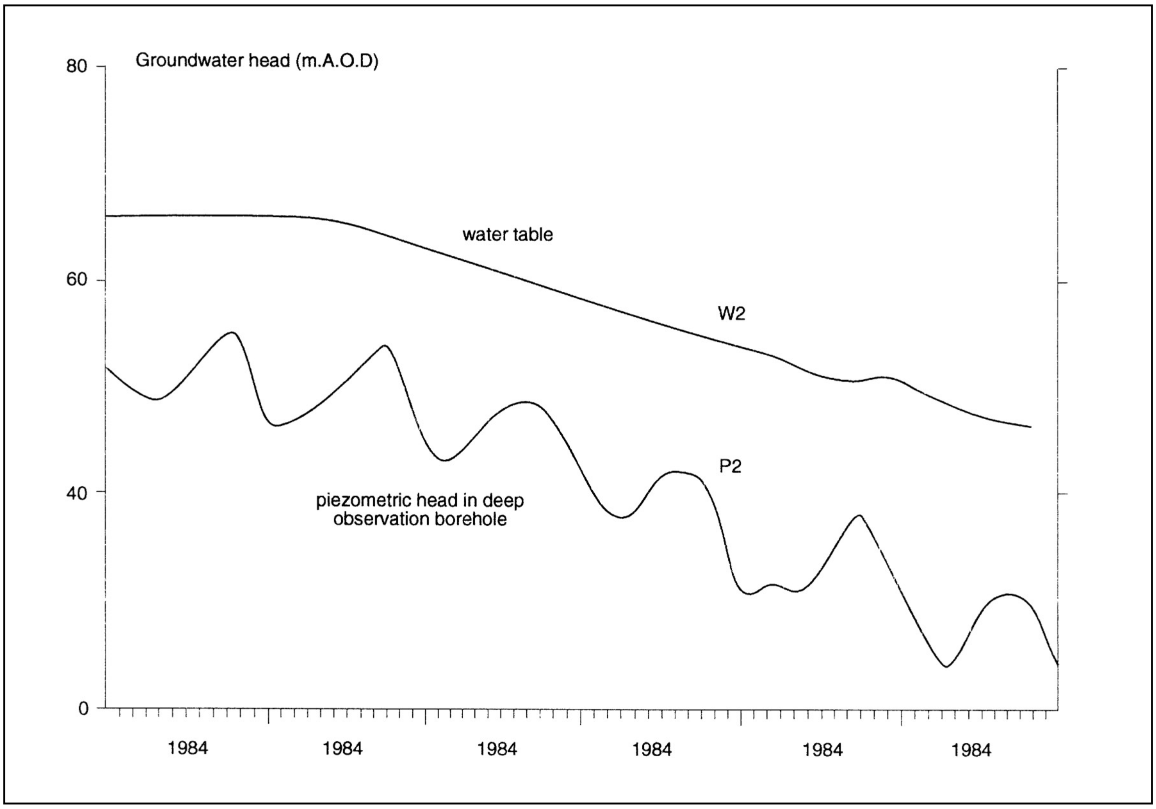


- After trend analysis, the time series should be evaluated on its *periodicity* due to seasonally varying rainfall, evapotranspiration or abstractions. A clear example of periodic fluctuations and a linear trend in the piezometric heads of deep groundwater in India due to abstraction for irrigation is presented in Figure 3.11. Spectral analysis is applied to recognise periodic fluctuations. The periodic properties of groundwater time series may be modelled by harmonic series. The linear least squares method is applied to derive the constant and the harmonic coefficients (Zhou, 1992a).
- Finally the *stationary* or *stochastic component* of the time series can be evaluated. This component is determined by subtracting the trend and periodic components from the original time series, which results in a groundwater time series of residuals. This time series is usually stationary and autocorrelated. Zhou (1992a) proposes using the mean of the residual time series to estimate the sampling frequency associated with the stationary component of the time series.
- Determination of the sampling frequency (Zhou, 1992a). The sampling frequency f equals the maximum value of three different frequencies, if all three are relevant:

$$f = \max(f_T, f_P, f_M)$$

f_T : the sampling frequency for trend detection;
f_P : the sampling frequency for periodicity detection;
f_M : the sampling frequency associated with the mean of the stationary component.

Figure 3.11 Periodic fluctuations and a linear trend in the groundwater heads in Gujarat, Western India, due to the over-exploitation of an alluvial aquifer (Rushton, 1998)



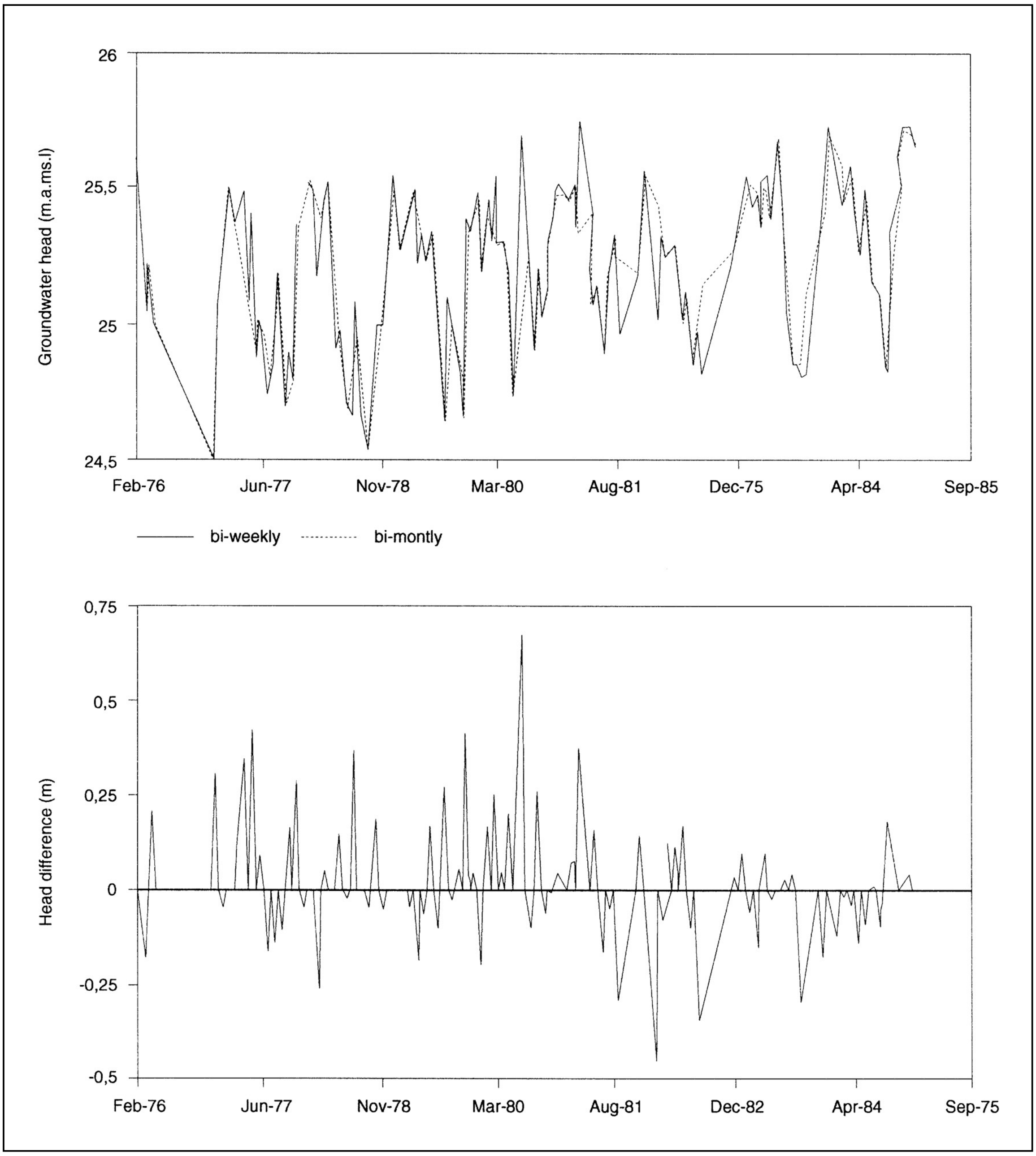
The relationship between accuracy of the estimate of the mean and the sampling frequency (f_M) is introduced by using the concept of standardised half width of the confidence interval of the mean. Eventually a family of curves is obtained with the standard half width of the mean versus the sampling frequency for various correlation coefficients, sampling periods, and autoregressive model structures. These curves clearly show that increasing the sampling frequency (f_M) beyond a certain critical value results in only a marginal increase of accuracy. The critical value is the key factor in determining f_M. The critical value decreases with an increase of the correlation coefficient. Therefore, for highly autocorrelated time-series, f_M is lower than for slightly autocorrelated series.

The analysis of periodic fluctuations using harmonic series reveals the highest significant periodic fluctuation in the real time series f_H. The sampling frequency to determine the trend f_P needs to be more than twice f_H. For example, when the significant periodic fluctuations in groundwater time series are monthly and annual, f_P should be at least twice a month. Furthermore the variances of the estimation of the harmonic coefficients can also be used as a criterion for determining the sampling frequency f_P (Zhou, 1992a).

Significant trends are explored by using Student's t distribution. If these trends prevail in the time series, Zhou (1992a) shows that for given standard trend magnitude and length of observation period, the power of the trend detection (power=0 for no trend recognition, and power=1 for complete recognition) will only be a function of the sampling frequency for a particular confidence level. Obviously, for a given trend magnitude the power of the trend detection increases with an increase of the sampling frequency. Furthermore the power of trend detection decreases with a smaller trend magnitude. In this way a family of curves is obtained with the power of the trend detection versus the sampling frequency for various trend magnitudes, observation periods and confidence limits. The sampling frequency f_T can be derived by evaluating the shape of these curves.

The proposed sampling frequency can be verified by comparing the characteristics of the original time series with the series obtained using the recommended frequency (Fig. 3.12). The statistical characteristics (see step 1) of the original time series based on, for example, two-weekly data should be nearly identical with the characteristics of the derived series based on the recommended lower frequency (e.g. monthly data).

Figure 3.12 Difference between groundwater heads measured with different sampling frequencies in a observation well in the Hupselse Beek catchment (The Netherlands)



Zhou (1992a) and Zhou and Li (1992) present comprehensive elaborations of the above-mentioned stochastic procedure for the redesign of the sampling frequency in two cases: groundwater level monitoring in the region around a Dutch well field for drinking water supply, and the monitoring of shallow groundwater levels in the Central Yellow Plain in China.

The above-mentioned stochastic procedure for determination of the sampling frequency is based on historical time series. Therefore the recommended frequency applies to a situation with no interference, e.g. in the case of the redesign of the background or specific monitoring network, after some years of monitoring without a significant change in the groundwater system (Fig. 3.12). However, the sampling frequency should be adapted if human interference starts and is expected to result in a trend and possibly other periodic fluctuations. The new sampling frequency can be determined by a combined stochastic-deterministic approach. The stochastically modelled time series are converted into a new time series using deterministic groundwater models, such as MODFLOW (McDonald and Harbaugh, 1988).

Adaptation or introduction of a trend and periodic fluctuations can be based upon the changes in groundwater heads simulated by these models. In this way the first sampling frequency of the specific monitoring network can be determined (Fig. 3.12). After some years the recorded time series can be used with the stochastic approach to refine the sampling frequency.

3.4.3 Simultaneous design of network density and sampling frequency

Under some conditions, a trade-off between sampling frequency and network density occurs for a particular hydrological variable (Van Geer and Van der Kloet, 1986).

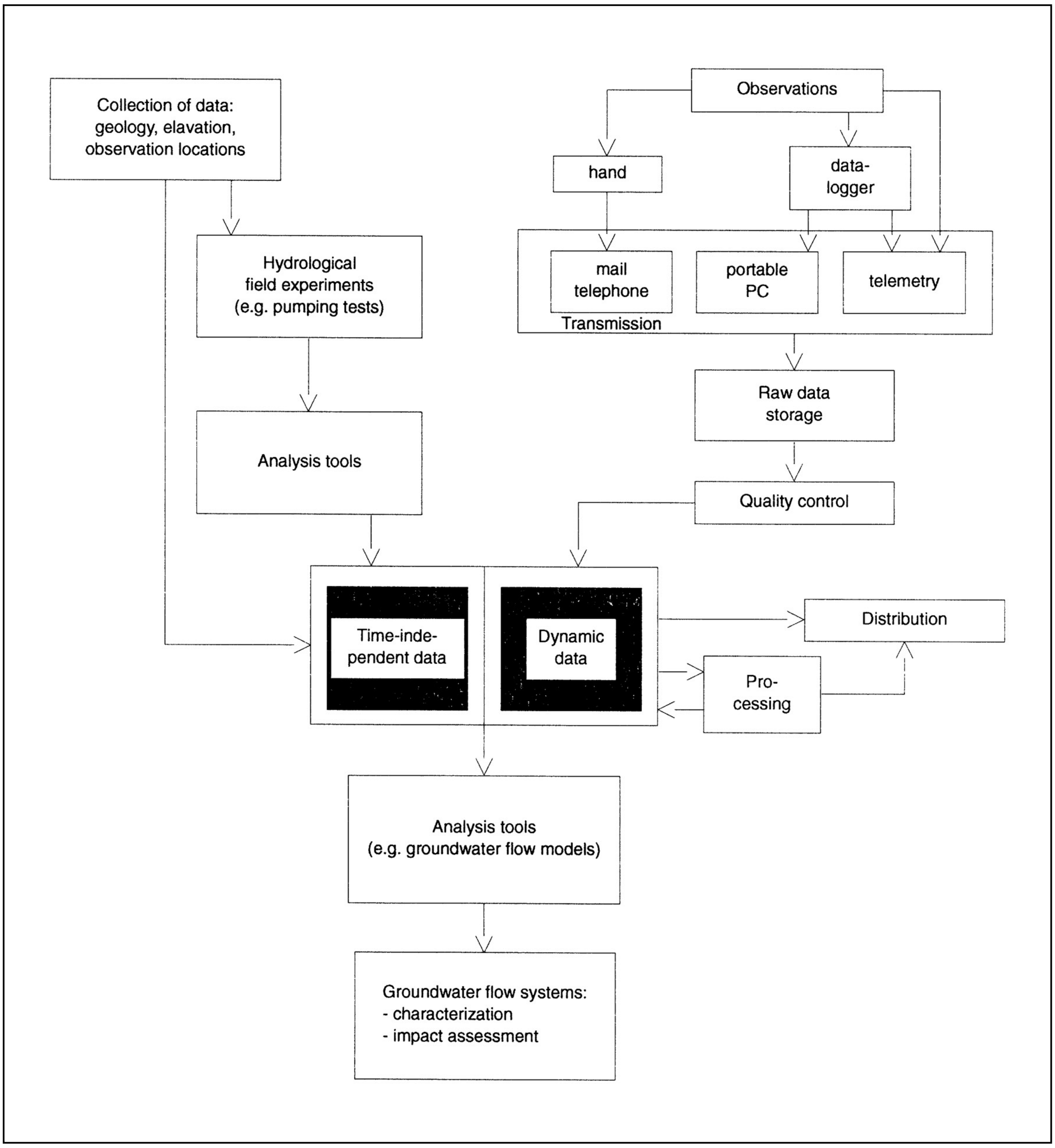
Stein (1998) introduces the dynamic semi-variogram in this context. The parameters of a hypothetical dynamic semi-variogram are allowed to change in time. The procedure to be followed is essentially the same as already explained for the network density. With the dynamic semi-variogram procedure, a value of a hydrological variable can be interpolated towards an arbitrary, unmonitored site and time, based on monitored data. The specified prediction error can serve as a criterion to find out whether the network and sampling density are already adequate.

Van Geer and De Stroet (1990) and Zhou (1992a) propose applying a stochastic-deterministic approach, i.e. a Kalman filtering model that consists of the Kalman filtering algorithm and a deterministic groundwater model. In this way observation-based information and physically-based information are incorporated in the network design. Usually the network design is obtained by minimising the network density and sampling frequency under the constraint of a prescribed threshold value of the standard deviation of the estimation error (*SDEE*). A trial and error procedure is followed, which implies that for a given network alternative a contour plot of the *SDEE* is computed with the Kalman model, and compared with the threshold value. In areas where the *SDEE* is lower than the threshold, the network density can be decreased, whereas the opposite occurs in areas where the *SDEE* is higher than the threshold. Alternatively, a closer fit of the *SDEE* can be achieved by adjusting the sampling frequency.

3.5 Groundwater information system

The collection of data from a groundwater monitoring network is useless, unless an information system is available that organises the data flow from observation or measurement to dissemination Although most steps are trivial, many mistakes are made in this context. The following steps can be distinguished (Fig. 3.13; readers are also referred to WMO, 1994):

Figure 3.13 Outline of a groundwater information system



- *Recording*. At specified times the value of a hydrological variable is recorded either in a field notebook or is stored automatically in a data logger. Recording by a skilled observer offers the possibility for an initial quality control. To maintaining good quality observations it is essential that the observation station itself is regularly inspected on possible malfunctioning of the observation devices.
- *Transmission*. At specified times the observed data go to the data-processing centre. Copies of field notebooks are mailed, or the observer calls the centre. Automatically stored data are retrieved on site with a portable PC. In the field a first quality check can be done, by plotting the observed data in a graph, and performing simple statistics

(e.g. determination of minimum and maximum). Fully automated stations can transmit the recorded data instantaneously or at regular intervals to the centre. This relative expensive option is attractive for remote areas, but skilled staff are needed. If the groundwater systems involved are significantly developed, these fully automated systems at far-off locations also can operate as an 'early warning system'; this implies that the system transmits a message if a specified level is exceeded (e.g. groundwater head below a certain value or Cl^- concentration above a critical level). In general, the recording and transmission is done by several observers. Therefore it should be adequately organised and well supervised, i.e. clear, concise notebooks with obvious instructions on coding, required accuracy and initial quality control, and a proper manual for each observation instrument, data logger and PC used for data retrieval.

- *Central data storage*. After receiving the data in various forms and storage media, the processing can start. For instance, correction of data for zero-shift, detection of missing values and replacement with an appropriate code, conversion of stage levels into discharges and transfer of water tables or piezometric levels to metres above datum.
- *Quality control*. A quality check must be carried out. Several methods are available, e.g. visual control by plotting in a graph, detection of outliers, comparison with similar time series, comparison with data of an allied hydrological variable (Delft Hydraulics, 1992). Information associated with the data element, such as the date and station code, need to be checked as well. Data can be stored in databooks and quality control can be done by hand, but usually it is more efficient to store the data in a computer system. Then, quality control is more feasible because it can be done automatically.
- *Processing*. After the data have been checked, descriptive statistics can be applied, such as the calculation of totals over different periods, mean, median, minimum and maximum and variation. Probability distributions can be computed, giving the probability of occurrence of certain events or return periods. At this stage, missing data also can be replaced by a predicted value by using statistical techniques, such as regression.
- *Dissemination.* People dealing with groundwater must have access to monitored data. If the data are not stored on a computer system, data books should be published. If not all basic data can be published, then at least some processed data are provided, and basic data can be supplied on request. If the data are available in a computer system, an on-line connection with the database is preferred. Such a connection makes it possible to retrieve and to analyse only those data relevant for the user. Of course, the database should be well organised, described and protected if people outside the data processing centre have access to it.

In a groundwater information system not only dynamic data are stored. Actually it should comprise two parts (Fig. 3.13), namely:

- a part in which time-independent data are stored, and
- a part with the dynamic data.

In the time-independent domain of the information system, for example, locations and descriptions of drillings are stored, as well as permeabilities, thickness of aquifers, hydraulic resistances, storativities. Observation locations and type of observation tools are also stored. It also should contain a logbook with comments on each monitoring site, such as malfunctioning or failure of the monitoring device, and checks with other, or more accurate tools.

Some groundwater information systems not only contain data, but also include analysis techniques. An example is REGIS (Kuipers, 1998). It comprises techniques to derive hydraulic properties (permeabilities, hydraulic resistances) of the subsurface from pumping test data (see also Chapter 8). Furthermore, different types of groundwater simulation models are linked to the information system. Dynamic and time-independent data available in the database can be tailored in different ways, so that they can serve as input data for the groundwater models.

3.6 Observation methods

3.6.1 Observation wells and piezometers

There have been extensive reviews of methods to monitor the groundwater variables (e.g. WMO, 1994; Otto, 1996). Below only techniques that monitor dynamic variables of groundwater in the field and the laboratory, such as elevation of water levels and water chemistry will be discussed. In this section a summary is presented, and only some of the recent developments are explained in more detail.

The data on groundwater heads and the chemical composition are obtained at observation wells or piezometers. Observation wells and piezometers should be installed carefully, taking account of rock type (consolidated, unconsolidated), and tested (WMO, 1994). An observation well is a non-pumping well used to observe the height of the water table. It is generally of larger diameter than a piezometer and may be screened throughout the thickness of the aquifer. The observation well is a length of fully slotted tubing that is lowered into the bore hole and backfilled with sand or soil around the side of the tube. Water can freely enter the tube along its entire length. Comparisons between individual piezometers and an open well have shown that the open well may act as a vertical conduit for water to flow, thereby distorting the groundwater conditions in the vicinity of the measuring location.

A piezometer is a non-pumping well, generally small in diameter, that is used to measure the groundwater head. It is an open-ended pipe, placed in a borehole that has been drilled to the desired depth in the ground. The bottom tip of the piezometer is fitted with a perforated or slotted screen (length about 1 m), to allow the inflow of water. The annular space around the screen should be filled with a gravel pack. The remaining annular space around the pipe can be filled with any material, except where the presence of aquitards requires a seal of bentonite clay or cement grouting to prevent leakage along the pipe. The water levels measured in piezometers represent the average hydraulic head at the screen of the piezometer. Rapid and accurate measurements can best be made in small diameter wells. If their diameter is large, the volume of water contained in the pipe may cause a time lag in water level changes.

Multi-piezometer installations are required for measuring the vertical hydraulic head distribution in anisotropic aquifer systems. Well-known methods are:

- Clusters of small-diameter piezometers that are placed in a single borehole at different depths and insulated by impervious material. The bottom part of each piezometer is screened. Leakage along the pipes from one piezometer to another has to be avoided.
- Clusters of piezometers in different boreholes in one place. The piezometers are installed at different depths. This is technically less demanding, but usually more expensive.

Before an observation well or piezometer is used, it should be developed. This entails injecting and abstracting water into and from the well, which induces groundwater flow alternately from and to the well. This procedure removes clogging from the screens and fine materials from the bottom of the well and the pack around the well. Subsequently a performance test is needed. An adequately installed well or piezometer should quickly follow head changes in the aquifer. A simple test is carried out by observing the fall of the water level in the well after the recharge of a known volume of water injected in the well. If the decline is too slow, the well should be further developed. Occasionally the tests of the wells or piezometers should be repeated, e.g. every two or three years.

3.6.2 Methods for monitoring groundwater quantity

Methods to monitor groundwater heads are described below. Furthermore some information is provided about the monitoring of groundwater well discharge. Methods to monitor groundwater recharge and discharge are not dealt with, so interested readers should consult the relevant publications (e.g. Lerner et al., 1990; Boiten, 1993; WMO, 1994; Simmers, 1997).

Both manual-operated and automated-recording instruments are available to measure groundwater heads in observations wells or piezometers (Table 3.1).

Table 3.1 Summary of instruments commonly used to measure groundwater heads (From: Driscoll, 1986; Nielsen, 1991; WMO, 1994; Otto, 1998)

Method	*Readout device*	*Advantages*	*Disadvantages*	*Costs, skills*
Manual				
Wetted-tape or flexible steel	Tape markings, sometimes with steel ruler	Accurate if depth is not too large	Several measurements needed to find approximate depth	Low price and easy to produce and to use
Dipper	Tape markings, sometimes with steel ruler	Accuracywithin 0.01 m, fast	Not-applicable in noisy environments	Low price and easy to produce and to use
Inertial devices	Tape markings	Accuracy within 0.01 m, fast and simple, to use in polluted groundwater	Calibration	Moderately priced, easy to operate
Two-electrode devices	Tape markings	Fast and simple, accuracy decreases with depth	Calibration, regular maintenance, batteries	Moderately to high priced dependent on cable length, easy to operate
Automatic recording				
Mechanical float recorder system	Drum chart or data logger	Widely applied	Float lag, mechanical failure, large well diameter	High priced, regular maintenance and checking
Pressure transducer	Data logger	Less components than float systems	Temperature effect, connection with the open air, calibration	High priced, regular checking
Ultrasonic sensors	Data logger	Less components than float systems	Temperature and humidity effects; for under-water types effects of pressure, solute concentrations and air bubbles	High priced, regular checking

(i) Manual groundwater head monitoring

The most common manual method is to suspend a weighted plastic-coated tape or flexible steel cable from the well head to a point below the water level. The water level is determined by the difference between the length of the wound-out cable and the length of that part that has been submerged. The length can either be read out directly from the markers, or by using a steel rule and the nearest marker, if fewer markers are adjusted on the tape or cable. Sometimes chalk or pastes that change colour are used to simplify the determination of the part submerged. Depths of 50 to 100 m are measured with ease. At greater depth a thin steel cable or lightweight plastic-coated tape is recommended. A skilled observer can achieve an accuracy of a few millimetres, although usually the error increases with depth.

The dipper is a cylindrical probe with a hollow space at the end, which is connected to a plastic-coated tape or flexible steel cable. If the dipper reaches the water table an audible signal is produced. Usually the depth is determined after raising and lowering the probe a number of times over a distance of a few centimetres. The depth of the water table equals the length of the tape or cable, usually measured at the well head.

An inertial device is a portable instrument that is designed so that a weight attached to a cable moves downwards at a constant velocity. If the weight reaches the water table, a braking mechanism prevents further downward movement. A counter gives the depth of the water table relative to the position of the portable instrument, usually the well head. With this equipment depths greater than 100 m can be measured. Accuracy is high if the cable has negligible stretch. An advantage is the high resistance, for example, to oil-polluted water and corrosive waters.

The two-electrode equipment consists of a portable reel with plastic-coated tape or cable connected to a probe at the end of the cable. The probe with a length of about 10 to 20 cm has two small adjacent electrodes. The circuit between the two exposed electrodes is closed when the probe reaches the water table, and a visible and/or audible signal is produced by a lamp or buzzer built-into the reel. The depth of the water table is read from the cable and is related to the position of the well head. With this equipment large depths to over 300 m are measured. The maximum depth is dependent on the length of the electrical cable, the design of the electrical circuitry and the acceptable weight of the equipment. The accuracy is comparable to the inertial device, although at great depth (in the order of 500 m) errors of 0.15 m are reported. However, if differences in water level have to be measured at this great depth (e.g. pumping test) and the cable is left suspended, an accuracy of millimetres is achieved.

Other techniques use other physical properties, such as the resistance or capacitance. The electrochemical effect of two different metals is also used. Then no batteries are needed. Measurable current flow can be produced in most groundwaters by immersing two electrodes (e.g. magnesium and brass). There is even a single electrode version: a magnesium electrode in the probe and a steel earth pin at the surface.

The above-mentioned instruments can also be used for free-flowing groundwater, as long as the groundwater head and the piezometer are no more than approximately 1.5 to 1.8 m above the soil surface. As an alternative the piezometer can be connected to a transparent tube just above the soil surface. The head is read from the marked tube, or a steel rule is used. If the head of the artesian aquifer is more than 1.8 m above the soil surface a pressure device needs to be applied (see below).

(ii) Automatic recording of groundwater heads

Automatic recording of groundwater heads is sometimes necessary, i.e. to investigate fast changes (e.g. pumping test, tidal effects, scientific research) or to monitor at remote places.

The mechanical float recorder is widely applied. This device is based on a float that is linked to a counterweight by a cable that runs over a pulley. The device is above the well, and the float and the counterweight are in the observation well if the diameter permits. In wells or piezometers with small diameters a vertical pipe has to be installed in the ground for the counterweight next to the well with the float. Float and float-line friction against the well casing can be a problem in wells with deep water levels and should be avoided. A large-diameter float is recommended because of its greater sensitivity to water level changes. The cable should be long enough to account for the groundwater fluctuation between the dates that the site is visited. The turn of the pulley is converted into a vertical movement of a pen on a drum chart (analogue recorder), which is driven by a spring or an electrical clock. Usually the speed at which the drum rotates can be adjusted, e.g. drum charts from one day to one month are available. Instead of recording on a drum chart, the turn of the pulley also can be converted into digital information (electronic recorder) that can be stored on a data logger. Then the water level is not recorded continuously as with the drum chart, but it is stored only at prescribed intervals, for example every minute, 15 minutes or hour. A proper data storage is necessary; the data logger in the field should have a backup memory. Then the data can retrieved again, if something goes wrong with the data exchange between the data logger, the portable PC in the field and the central computer in the office (Fig. 3.13). Careful maintenance and checking of the float recorder is a prerequisite to avoid malfunctioning. Every time the recorder is visited the recorded level should be checked against manual observations. In reliability, versatility, maintenance efforts and need for skilled staff, the float type method is considered to be a good compromise compared to other automatic recording methods. Other methods based upon the pressure transducer and ultrasonic sensors are being used increasingly (Table 3.1).

Table 3.2 Accuracy of various methods to measure groundwater heads (After Boiten et al., 1995)

Method	*Accuracy (m)*
Floaters	
• analogue recorder	
- diameter 0.10 m	0.003–0.005
- diameter 0.20 m	0.001–0.003
• electronic recorder	
- diameter > 0.08 m	0.001–0.003
Pressure transducers	
• cheap types	0.010–0.050[1]
• expensive types	0.002–0.010[1]
Ultrasonic sensors	0.005–0.020

1. Depends on measuring range, etc.

In a pressure transducer the water pressure is converted to an electrical signal. The transducers are designed so that the water pressure is linearly related to the electrical signal. A calibration of the pressure transducer is recommended. The small pressure transducer is lowered into the observation well or piezometer below the deepest possible water level and stays there. The transducer is connected to the open air via a cable (atmospheric pressure is the reference) and to a logger which periodically stores the water level reading. Temperature affects both the electronic and the mechanical part of the transducer; at high or low temperatures the transducer may not operate adequately. Nowadays there are probes that combine a pressure transducer and a data logger. The 15–20 cm long probes are lowered into the observation well and stay there for one month, recording pressure data every 15 minutes. This device is a good alternative for locations exposed to vandalism.

An ultrasonic sensor transmits ultrasonic pulses. These pulses are reflected at the water surface and received again by the sensor. The time between transmission and receipt is linearly related to the distance between the sensor and the water surface, and consequently to the water level. Ultrasonic sensors can be placed either above or below the water surface. Below the water surface the propagation velocity is lower than in the air. Ultrasonic sensors are sensitive to temperature and humidity. Furthermore the performance of underwater types is affected by variations in water pressure, solute concentration and occurrence of air bubbles. The ultrasonic sensor is connected to a data logger.

An indication of the accuracy of the various methods is given in Table 3.2. Errors from the conversion of readings related to the well head (local reference) to datum (e.g. m a.m.s.l.) are not included.

(iii) Abstraction monitoring

Mechanical *flow meters* record the total water abstraction and the data is read from the meter at periodic intervals. Electronic totalling flow meters are used if the data can be stored on a logger. For proper groundwater management, each abstraction well within a well field should be instrumented to monitor abstraction amounts and well performance.

3.6.3 Methods for monitoring groundwater quality

Methods for monitoring the chemical composition of groundwater are essential for the identification of groundwater flow systems (background monitoring). Furthermore the monitoring is necessary to investigate the development of the chemical composition during the exploitation of the groundwater resources (specific monitoring). This section will describe the sampling of aquifers, reasons why the chemical composition of the sample may differ from the composition of the aquifer itself, sampling devices and analysis techniques. For a more comprehensive description, readers are referred to WMO (1994) and Otto (1998).

(i) Sampling of aquifers

The objective of groundwater quality monitoring programmes is to obtain samples of water that represent the groundwater within an aquifer, or within particular depth intervals in an aquifer. For some purposes hydrochemical monitoring should be discrete in its location within the aquifer system, which implies that additional hydrochemical monitoring boreholes have to be constructed. Where possible installed observation wells or piezometers for head measurements should be used, provided that the design is suitable for efficient sampling.

When sampling aquifers for monitoring groundwater quality a distinction should be made between non-point and point sampling. Non-point samples obtained by pumping from open boreholes or fully screened observation wells provide information on the overall changes in groundwater quality; the samples might be a mixture of different groundwater types. Such a sampling programme monitors broad changes in the chemical composition of groundwater in an aquifer at relatively low cost. Non-point sampling, however, is inadequate for monitoring groundwater quality changes at a local and three-dimensional scale. Therefore a different sampling design to the non-point method is required to monitor site-specific groundwater changes such as salt water intrusion or plumes from waste disposal sites. Important considerations in the design include the need for close vertical interval point sampling and sampling locations that take the groundwater flow patterns into account. Using piezometers for water quality sampling, the screen length of the borehole casing or piezometer has to be considered. There is little difference if hydraulic head is measured in a small or large volume of the aquifer, because the vertical hydraulic gradient in the permeable unit is usually small. However, the result can be dramatically different for water quality sampling.

The most reliable method for multilevel sampling can be achieved by piezometers in different boreholes at various depths at one site. The second best method is to bundle piezometers at different depths in one borehole and install samplers at discrete sampling points. Sometimes it is possible to sample continuously in an uncased borehole or fully screened standpipe by using a multi-layer sampling probe (MLS). The MLS collects undisturbed water samples at small depth intervals (e.g. 3 cm). The MLS consists of a rod with tens of cylindrical cells on top of each other with a small volume (e.g. 15 cm^3). The MLS is lowered into the well to the required depth. The cells are separated by rubber seals or inflated packers that prevent vertical mixing of groundwater in the borehole. The sampling is based on the dialysis-cell method, using the process of diffusion through a nylon membrane. For large ions the diffusion process reaches an equilibrium after a maximum of 5 days. After the equilibrium stage has been reached the MLS can be pulled out of the well, and the composition of the water in each cell can be analysed (Krajenbrink et al., 1989; Ronen et al., 1986). The results obtained with the MLS are only reliable if cross-migration of fluids in the well can be prevented.

(ii) Factors changing the chemical composition of groundwater samples

To obtain a representative sample of groundwater the well casing volume of a well has to be purged by bailing or pumping at least three bore volumes. The adequacy of the purging can be checked in the field by examining readily measured physical-chemical components (e.g. temperature, pH, EC) of the purged groundwater that should approach a constant value. A small diameter well would also mitigate purging problems for chemical sampling.

Three problems may seriously affect the chemical composition of a groundwater sample, namely:

- effects of well construction and contamination with drilling fluid. For example, the use of cement rings in dug wells (higher pH) or bentonite as seal in a piezometer can affect the water chemistry. Also, drilling fluid that has infiltrated into the surrounding aquifer material is often difficult to remove.
- sample deterioration. The chemistry of samples can change due to variations in temperature and gas pressure. Cool and dark storage of samples is necessary. In situ analysis, or prompt transportation to the laboratory can preserve data quality;
- careless field and laboratory practices. Sample contamination caused by improper bottle washing and filtering is a main concern. It is recommended to wash the bottle with the groundwater to be analysed, prior to the collection of the sample. Organic compounds can be absorbed by plastic containers, also the loss of volatile components can be a problem. The quality of laboratory analyses should be checked by submitting blanks and duplicates.

(iii) Sampling devices

The initial consideration in selecting a sampling device is whether the well will accommodate the device. The smaller the diameter of the well, the more limited, complex and expensive the available samplers become. Depending on the analytical requirements of a sampling programme, the device should not affect the chemical and physical composition of the sample:

- the material of the device should not sorb or leach contaminants;
- pH, Eh of the sample should not alter;
- no volatile should be released from the device;
- the device should not introduce air or non-inert gas into the sample.

Also, the depth of sampling has to be considered. The deeper the sampling interval, the more head a pump must overcome to deliver a sample to the surface. Ease of the sampling procedure, transportation, cleaning and maintenance of the device should be considered. Otto (1998) lists the characteristics (e.g. well and device diameter, sampling depth, sample volume, chemical alteration, relative costs) of some standard sampling devices, such as the cheap bailers and the expensive submersible pumps.

Free-flowing or artesian wells are easy to sample by making a small hole in the observation well about 15 to 20 cm above the soil surface, which is normally closed to allow the observer to measure the correct groundwater head. The water is allowed to flow for some time to purge it, before taking samples.

(iv) Analysis techniques

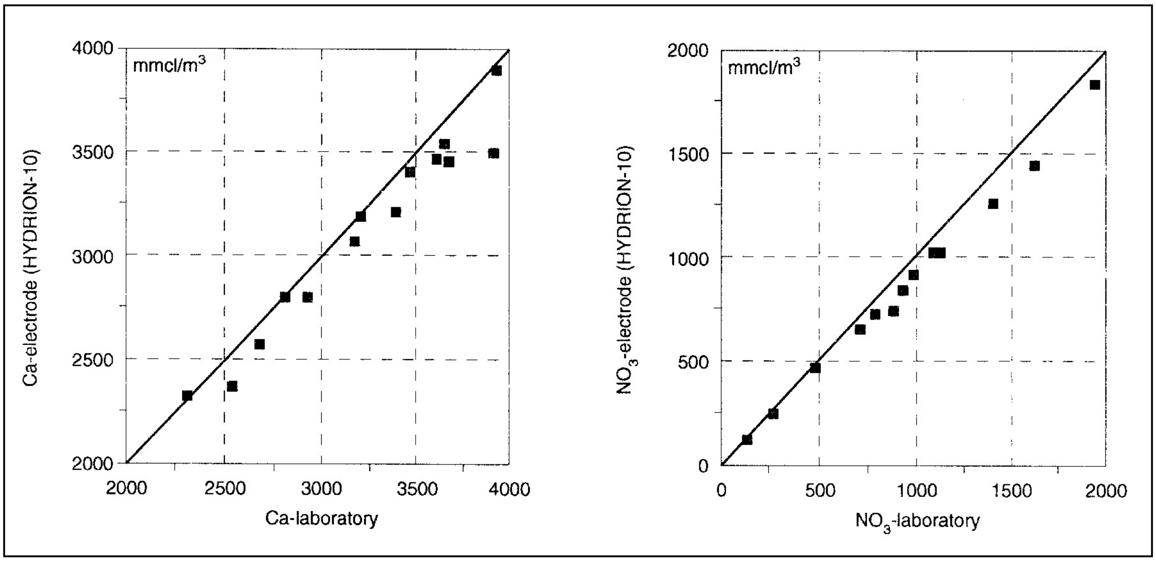
Analysis techniques can be subdivided into the conventional analytical chemical techniques predominantly used in the laboratory and sensors that measure one or more physical-chemical components either in the laboratory or in the field. Excellent handbooks are available that describe the former group (e.g. Clesceri et al., 1989; Velthorst, 1993). In this section only the sensor-based approach that has produced encouraging results in recent years, is described.

For many years, sensors have been used to measure the physical-chemical properties (pH, Eh, T, EC, turbidity) of groundwater in a well. There are many commercial sensors available. Currently a few monitoring devices are on the market that can measure chemical components (major anion and cations, dissolved oxygen) either in the laboratory or directly in the field. These portable in-situ monitors are expensive, but when many samples have to be analysed the costs per sample may be below those of the conventional analytical techniques. They require repeated and careful calibration but are easy to operate. CSIRO[2] has developed an in-situ water quality monitor that can measure a range of physical and chemical water parameters. Another promising device, HYDRION-10[3], analyses samples with a set of integrated ion-selective or ion-related electrodes (ISEs). It consists of a sensor unit and a measurement unit. The sensor unit is made of stainless steel and contains the electrodes. This unit fits into an ordinary 500 ml glass beaker. A water sample of 250 ml is sufficient for the analysis. The sensor unit can also be put directly into the water, e.g. in a free-flowing well or in a surface water stream.

The measurement unit takes care of the operation and the data storage. The operational features include the simultaneous measurement of EC, pH, K^+, Ca^{2+}, Mg^{2+}, NH_4^+, Cl^-, NO_3^-, and HCO_3^-; the specially-developed software automatically corrects for (mutual) effects affecting the measurement of an individual component. No additives are needed, but filtering is necessary for turbid water. About 10–15 samples can be measured per hour. For about two years the outcome of the HYDRION-10 has been compared with laboratory results using conventional analytical procedures. Some results are given in Figure 3.14. Results correspond reasonably well, both for weathering ions and for anthropogenic ions.

2. Dr G. Skying and Mr I. Jones, CSIRO Division of Water Resources, P.O. Box 1666, Canberra, Australia.
3. HYDRION, P.O. Box 522, 6700 AM Wageningen, The Netherlands.

Figure 3.14 Comparison of some measured ion concentrations with conventional analytical techniques and a with set of integrated ion-selective electrodes HYDRION-10



3.7 Concluding remarks

Background monitoring networks should be installed if groundwater exploitation is expected or planned. Monitoring should also be started when other future human interference can lead to deterioration of the underground environment, ecologically relevant wetlands, springs or streams due to changed groundwater conditions. Specific monitoring networks should be introduced when a particular human interference starts.

- Groundwater systems and the impact of changes can only be identified adequately if both physical and chemical groundwater data are collected. Furthermore vertical differences in groundwater components should not be neglected.
- Groundwater monitoring networks should be supported by other networks collecting data on groundwater recharge and stream flow.
- Network density and sampling frequency are sometimes hard to define for background monitoring. Knowledge obtained in similar areas should be used. In subsequent phases statistical and geostatistical techniques can be applied successfully. Kriging has proved to be successful for the determination of network density, whereas time-series analysis has been favourably applied to define sampling frequency. Operational procedures are available.
- Groundwater information systems, which organise the flow of monitoring data from observation to dissemination, are a prerequisite for an adequate monitoring network.
- Several methods are available to monitor groundwater quantity data. Required frequency (continuous or intermittent), accuracy and skills, and available budget determine the best selection. Similar criteria apply to methods for quality monitoring. Additionally, diameters of wells and devices, required sample volume, and changes in chemical parameters should be considered.
- Groundwater quality monitoring should consider the frequent occurrence of vertical concentration gradients. For some purposes it is necessary to sample discrete parts of the underground environment instead of taking a mixed sample over the entire depth is a necessity. Special hydrochemical piezometers or sampling devices (e.g. multi-layer sample probes) are needed.

- Besides conventional analytical analysis techniques used in the laboratory, sets of integrated ion-specific electrodes produce encouraging results. The latter analysis technique can also be applied in the field; this offers maximum flexibility and a minimum of sample deterioration due to transport and storage.

3.8 References and additional reading

BACHMAT, Y. 1989. *Management of Groundwater Observation Orogrammes*, World Meteorological Organisation, Operational Hydrology Report No. 31, Geneva, 96 pp.

BOITEN, W. 1993. Flow-Measuring Structures. In: Sanderson et al. (eds.), *Open Channel Measurements.* Special issue, Flow Measurements and Instruments, pp. 17–24.

BOITEN, W.; DOMMERHOLT, A.; SOET, M., 1995. *Handbook of Discharge Measurements in Open Water Channels.* Department of Water Resources Report No. 51, Agricultural University, Wageningen, 158 pp. (In Dutch.)

BROWN, R. H.; KONOPLYANTSEV, A. A.; INESON, J.; KOVALEVSKY, V. S. (eds.), 1972. *Groundwater Studies; An International Guide for Research and Practice.* Studies and Reports in Hydrology, No. 4, UNESCO, Paris.

BOX, G. E. P.; JENKINS, G. M., 1976. *Time Series Analysis: Forecasting and Control*. Hilden Day, San Francisco, California.

CARILLO-RIVERA, J. J., 1998. Monitoring Exploited Aquifers Resulted in Subsidence, Example: Mexico City. In: H. A. J. van Lanen (ed.), *Monitoring for Groundwater Management in (Semi-)Arid Regions,* Studies and Reports in Hydrology No. 57, UNESCO, Paris, pp. 191–208.

CHATFIELD, C., 1989. *The Analysis of Time Series.* Fourth edition, Chapman and Hall, London, 241 pp.

CLESCERI, L. S.; GREENBERG, A. E.; TRUSSELL, R. R. (eds.), 1989. *Standard Methods for the Examination of Water and Wastewater*. Seventeenth Edition. American Public Health Association, American Water Works Association and Water Pollution Control Federation.

DELFT HYDRAULICS, 1992. *HYMOS: A Database Management and Processing System for Hydro-Meteorological Data*, Delft.

DE MARSILY, G., 1986. *Quantitative Hydrogeology*. Academic Press.

DRISCOLL, F. G. 1986. *Groundwater and Wells*. Johnson Filtration System Inc., St Paul, Minnesota, 1089 pp.

KAŠPÁREK, L.; KREJCOVÁ, K., 1994. *BILAN, Water Balance Model: User's Guide*. T. M. G. Masaryk Water Research Institute, Prague, 12 pp.

KRAJENBRINK, G. J.; BOUMANS, L. J. M.; MEINARDI, C. R., 1989. Hydro-Chemical Processes in the Top Layer of Groundwater under Pasture-Land. *Proceedings, Symposium on Nitrogen in Organic Wastes, Applied to Soils. Aalborg, Denmark.* Academic Press, London, pp. 317–33.

KUIPERS, T. T., 1998. Geohydrological Information Management Systems. In: H. A. J. van Lanen (ed.), *Monitoring for Groundwater Management in (Semi-)Arid Regions*. Studies and Reports in Hydrology No. 57, UNESCO, Paris, pp. 107–27.

LANGBEIN, W. B., 1954. Stream Gauging Networks. In: *Proc. Assemblée Générale de Rome,* Rome, IAHS Publication No. 38, pp. 293–303.

LERNER, D. N.; ISSAR, A. S.; SIMMERS, I., 1990. *Groundwater Recharge. A Guide to Understand and Estimating Natural Recharge*. IAH, Vol. 8, Verlag Heinz Heise GmbH, Hannover, 345 pp.

LLOYD, J. W., 1996. Specific Monitoring. In: H. A. J. van Lanen (ed.), *Monitoring for Groundwater Management in (Semi-)Arid Regions.* Studies and Reports in Hydrology No. 57, UNESCO, Paris, pp. 47–64.

McDonald, M. G.; Harbaugh, A. W., 1988. *A Modular Three-Dimensional Finite-Difference Groundwater Flow Model.* United States Geological Survey.

Moss, M. E., 1982. *Concepts and Techniques in Hydrological Network Design*. World Meteorological Organisation, Operational Hydrology Report No. 19, Geneva, 30 pp.

Moss, M. E. (ed.), 1986. *Integrated Design of Hydrological Networks,* IAHS Publication No. 158, Wallingford, 415 pp.

Nielsen, D. M., 1991. Practical Handbook of Ground-Water Monitoring. Lewis Publishers, Chelsea, Michigan, 717 pp.

OEPA, 1993. *Technical Guidance Manual for Hydrogeologic Investigations and Groundwater Monitoring Programs*. Ohio Environmental Protection Agency, Division of Drinking and Groundwaters, Columbus, Ohio.

Otto, C. J., 1998. Monitoring Tools and Type of Recording. In: H. A. J. van Lanen (ed.), *Monitoring for Groundwater Management in (Semi-)Arid Regions.* Studies and Reports in Hydrology No. 57, UNESCO, Paris, pp. 65−90.

Ronen, D.; Margaritz, M.; Levy, I., 1986. *A Multi-Layer Sampler for the Study of Detailed Chemical Profiles in Groundwater*. Water Research 23, pp. 311–5.

Rushton, K. R., 1998. Monitoring an Over-Exploited Aquifer in India. In: H. A. J. van Lanen, (ed.), *Monitoring for Groundwater Management in (Semi-)Arid Regions.* Studies and Reports in Hydrology, UNESCO, Paris, pp. 141–50.

Simmers, I. (ed.), 1997. Recharge of Phreatic Aquifers in (Semi-)Arid Regions. *IAH Intern. Contr. Hydrogeol.,* A. A. Balkema, Rotterdam, 277 pp.

Stein, A., 1998. Geostatistical Procedures for Analysing Spatial Variability and Optimising the Collection of Monitoring Data. In: H. A. J. van Lanen (ed.), *Monitoring for Groundwater Management in (Semi-)Arid Regions.* Studies and Reports in Hydrology No. 57, UNESCO, Paris, pp. 91–106.

UNESCO, 1990. International Hydrological Programme. *Hydrology and Water Resources for a Sustainable Development in a Changing Environment*; detailed plan for the fourth phase of the IHP (1990−1995). Paris, 54 pp.

UNESCO, 1992. *International Glossary of Hydrology*. Paris, 413 pp.

Van Bracht, M. J.; Romijn, E., 1985. *Redesign of Groundwater Level Monitoring Networks by Application of Kalman Filtering and Kriging Methods.* Proceedings, Symposium on the Stochastic Approach to Subsurface Flow, Montvillargenne.

Van Geer, F. C.; De Stroet, C. B. M. , 1990. *Kalman Filtering Packages for Program CISKA, a Computer Model for the Interpolation of Heads with Kalman Filtering Algorithm.* Report TNO Institute of Applied Geoscience, Delft.

Van Geer, F. C.; Van der Kloet, P. , 1986. Tradeoff Between Frequency and Density of a Groundwater Measurement Network. In: M. E. Moss (ed.), 1986. *Integrated Design of Hydrological Networks,* IAHS Publication No. 158, Wallingford, pp. 213–29.

Van Lanen, H. A. J., (ed.) 1996. *Monitoring for Groundwater Management in (Semi-)Arid Regions.* Studies and Reports in Hydrology No. 57, UNESCO, Paris, 224 pp.

Velthorst, E. J., 1993. *Manual for Chemical Water Analysis*. Department of Soil Science and Geology, Agricultural University, Wageningen, 109 pp.

WMO, 1994. *Guide to Hydrological Practices. Data Acquisition and Processing, Analysis, Forecasting and other Applications*. Fifth edition, WMO-No. 168, Geneva,735 pp.

Zhou, Y., 1992a. *Design and Analysis of Groundwater Monitoring Networks*. IHE, Delft, 124 pp.

Zhou, Y., 1992b. *Sampling Frequency for Monitoring the Actual State of Groundwater Systems.* Research Report, IHE, Delft.

Zhou, Y.; Li, B., 1992. *Sampling Frequency for Monitoring Shallow Groundwater Levels in the Central Yellow River Plain*. Proceedings of International Workshop on Groundwater and Environment, Beijing.

4 Processing and presentation of data

4.1 Scope

The processing and organised display of hydrogeological parameters are essential components of hydrogeological interpretation and reporting. Processing and display of data are inseparable and follow each other logically. They involve:

- data collation,
- data validation and storage,
- retrieval of data for processing and creation of displays.

Typical use for hydrogeological displays may include:

- identification of anomalies and outliers in data series,
- depiction of trends and patterns,
- comparison between several hydrogeological data sets,
- interfacing hydrogeological data with other data sets,
- spatial incorporation into Geographic Information Systems (GIS),
- enhancement of reports.

These are but a few examples of a vast array of possible processing and display of hydrogeological variables. An example of a previous publication that has partially dealt with the subject, is that of Lloyd and Heathcote (1985).

Since the introduction of personal computers, significant developments in computer software have come about. Within PC environments many displays that could previously only be constructed with great difficulty can nowadays be created with relative ease. This does not rule out the manual processing of data; indeed many of the displays discussed below can equally well be drawn by hand. However, considering the massive amount of data that is generated nowadays, preference should be given to the processing of data by computer,

4.2 Data types and presentation possibilities

For a meaningful discussion on the processing and display of hydrogeological variables, certain data groupings are necessary. The groups discussed below have been selected to represent the various environments in which hydrogeologists operate. There are two main groups: point data and spatial information.

4.2.1 Point data

Point data are the data that are generated at a specific geographical point. In hydrogeological terms, this data may assume many forms. Examples are:

- geological data, such as borehole logs;
- hydrogeological data, such as depths to water and yields;
- construction data, such as casing, piezometer, borehole development and costs;
- type of equipment, such as pumping equipment and recorders;
- geophysical data, including surface and borehole measurements;
- hydraulic properties of the aquifer(s), such as transmissivity and storage;
- water abstraction and groundwater level records;
- hydrochemistry;
- other information, such as topography and surface drainage.

Even though many of these variables have spatial connotations, they refer back to the groundwater abstraction point and are, for that reason, grouped under point data. Examples of variables with spatial connotations in the above list include the geological log, which has a vertical dimension, and surface geophysics, which has both horizontal and vertical dimensions.

4.2.2 Spatial information

The concept 'spatial information' emphasises two important differences from point data. The term 'spatial' suggests that point data are processed or displayed in the geographical context, such as on a map. The term 'information' suggests that the point data have been processed, usually to the extent that the original values for variables can no longer be recognised.

4.2.3 Presentation possibilities

Tools for the presentation of data and information can be grouped under three main headings. These are:

- graphs;
- statistical tables and displays;
- maps and cross sections.

A significant overlap is possible between these categories; for instance, maps may contain graphs or statistical interpretations as insets. Alternatively, many of the commonly used displays, such as specialised hydrochemical diagrams, have statistical undertones. Also, most of the statistical displays that will be discussed below, fall in the category of graphs. Statistical displays differ from ordinary graphs in that they usually contain processed information together with raw data. As the discussion of the various displays evolves, the different categories of displays that will be demonstrated in this chapter, will become clear.

4.3 Graphic processing and presentation of data

In order to present hydrogeological point data in as few concise diagrams as possible, certain logical groupings are necessary. The following is suggested:

Group 1: Borehole information, such as geological, hydrogeological and construction data; progressive borehole yield during drilling; packer testing results; borehole geophysics.

Group 2: Hydraulic properties of the aquifer(s), such as hydraulic conductivity, transmissivity, storage and dispersion.

Group 3: Time-dependent data, such as groundwater abstraction; groundwater levels; water quality.
Group 4: Specialised hydrochemical diagrams for detailed chemical interpretations.
Group 5: Other graphics, such as bar, stack, scatter and line graphs.

Most point data generated during hydrogeological investigations, can be accommodated within these five groups.

Layouts for graphs to depict variables within each of these groups may vary from one presentation to the next. There are, however, three basic rules to be considered when constructing displays. These are:

- the displays must be clear, concise and legible,
- they should contain only relevant data,
- displays should be provided with the necessary headings and labels to ensure that they will be meaningful entities on their own.

With these rules in mind, displays that may be used by the hydrogeologist will now be recommended and discussed.

4.3.1 Borehole information

Vast amounts of valuable data are generated during the drilling of a borehole. Most of these are of the 'one off' type, in contrast to repeated measurements such as water levels for instance. Essential 'one off' variables that should be recorded during drilling are:

- geological description of cores and cuttings,
- depth to water and yield,
- penetration rate during drilling,
- measurements on water (electrical conductivity and pH for instance),
- construction data (hole diameter, depth, casing properties).

Other data that are usually recorded after the hole has been completed, are:

- borehole geophysics,
- packer testing results.

The latter two variables are not truly of the 'once off' type, because they may be measured repeatedly, if so desired.

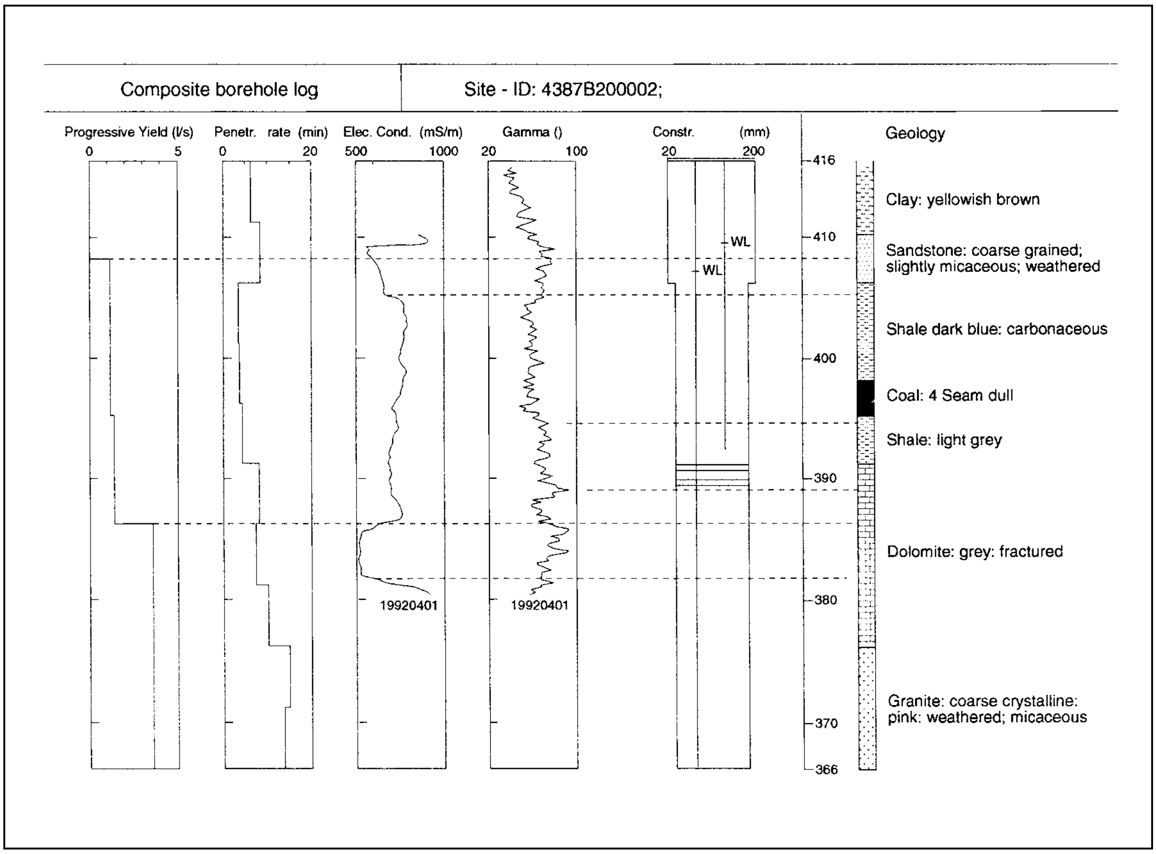
Each of the above variables should be recorded individually for the sake of establishing complete records. Interpretation of the many data sets is, however, very difficult and time-consuming. Ideally, all this information should be summarised onto a single sheet of paper in a format that can easily be included in a report (Figure 4.1).

Figure 4.1 presents one such summary. In this diagram, the geological column has been placed to the right, with abbreviated geological descriptions on the far right. This arrangement allows other variables to be plotted to the left, leaving room for projections and interpretations. This kind of presentation obviously has its limitations, the most important of which is the limited number of vertical profiles that can be included on a single sheet of paper. In practice, it has been found that up to seven vertical profiles can be combined on a normal-sized page. Thereafter, the horizontal scales become so small that variations in values are no longer obvious. If more than seven variables need to be plotted, they should be plotted onto two sheets, repeating essential columns such as the geology on the second sheet.

The vertical scale is unlimited, since logs may be carried over to following pages. For the sake of clarity, a single page that contains the complete log always should be included; certain interesting portions of the log may then be presented, enlarged, on successive pages.

Many other examples of log summaries are found in practice. The exact layout of the logs is not important, as long as the three basic rules for diagram construction, as indicated above, are adhered to.

Figure 4.1 Composite hydrogeological borehole log



4.3.2 Hydraulic properties of the aquifer

Graphical presentation of hydraulic properties of an aquifer may be done on two levels.

The first level is plots of drawdown data for the calculation of the aquifer constants. Such plots and the interpretation methods are discussed in Chapter 8 and have been well documented in literature (Kruseman et al., 1991).

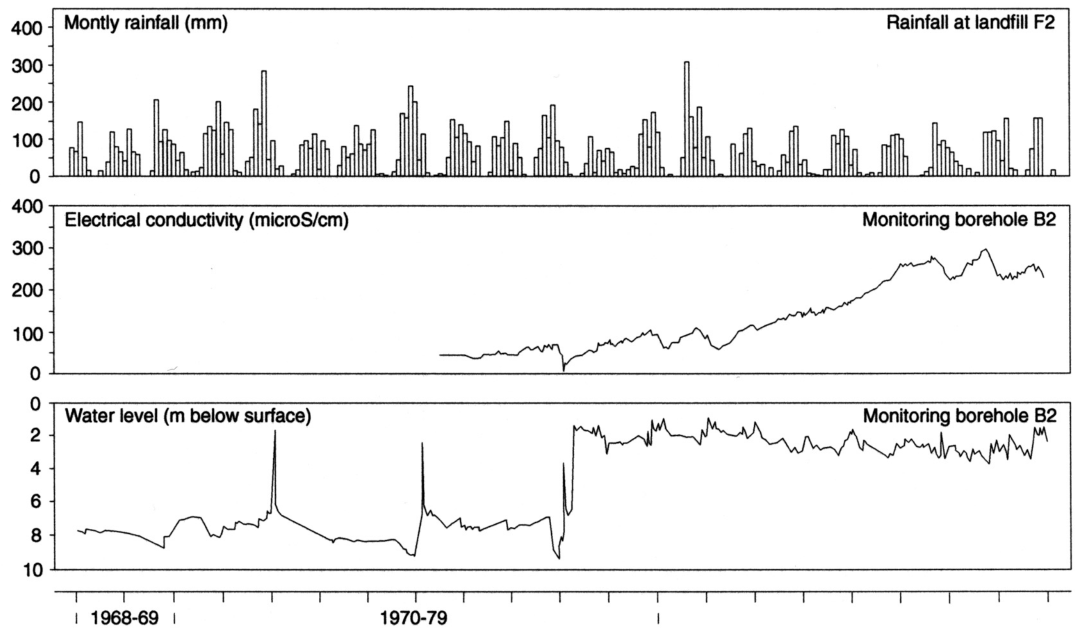
The second level is the presentation of hydraulic properties of aquifers in ways other than time-dependent plots. This usually requires the data to be converted into spatial information. This application will be discussed in detail in Section 4.5.

4.3.3 Time-dependent data

The hydrogeologist measures a variety of parameters that are time-dependent, e.g. water levels, pumping rate and duration, rainfall and water chemistry. Plotting of this data does not normally present a problem. Line or scatter plots and bar charts are used for this purpose. The data sets may be visually compared by combining plots as in Figure 4.2. This type of presentation is effective because of its clear and concise nature.

Processing and interpretation of time-dependent data are usually done through statistical methods. Possibilities include variance within data sets, regression analysis, as well as correlations between the different data sets. These procedures will be discussed under statistical methods in Section 4.4.

Figure 4.2 Time dependent plot of water levels, electrical conductivity and rainfall



4.3.4 Specialised hydrochemical diagrams

Through the years, many special displays that meaningfully present hydrochemical data have been devised. Of these, six displays stand out in terms of clarity and significance.

They are the Piper (Piper, 1944), Durov (Durov, 1948), Expanded Durov (Lloyd, 1965), SAR (Wilcox, 1955; Bower et al., 1968), Schoeller (Schoeller, 1962) and Stiff (Stiff, 1951) diagrams. All of these diagrams are so-called multivariate displays, simultaneously taking up to eight variables into consideration, often projecting these variables to a single point on the diagrams. Examples of the Piper and Expanded Durov diagrams, with their plotting procedures, are included in Figures 4.3 and 4.4.

Advantages of using these diagrams include:

- the plotting of numerous water analyses onto a single diagram,
- the classification of waters according to their chemical characteristics,
- the identification of trends.

Figure 4.3 Plotting procedures for the Piper diagram: Convert mg/l to meq/l by division (Ca/20, Mg/12, Na/23, K/39, T.Alk./50, SO_4/48,Cl/35.5, NO_3/62. Add Na +K and Cl + NO_3. Calculate percentage of cations and anions. Plot cations by scaling off Ca, then Mg. Plot anions by scaling off T.Alk. then SO_4. Project cation and anion points to triangle

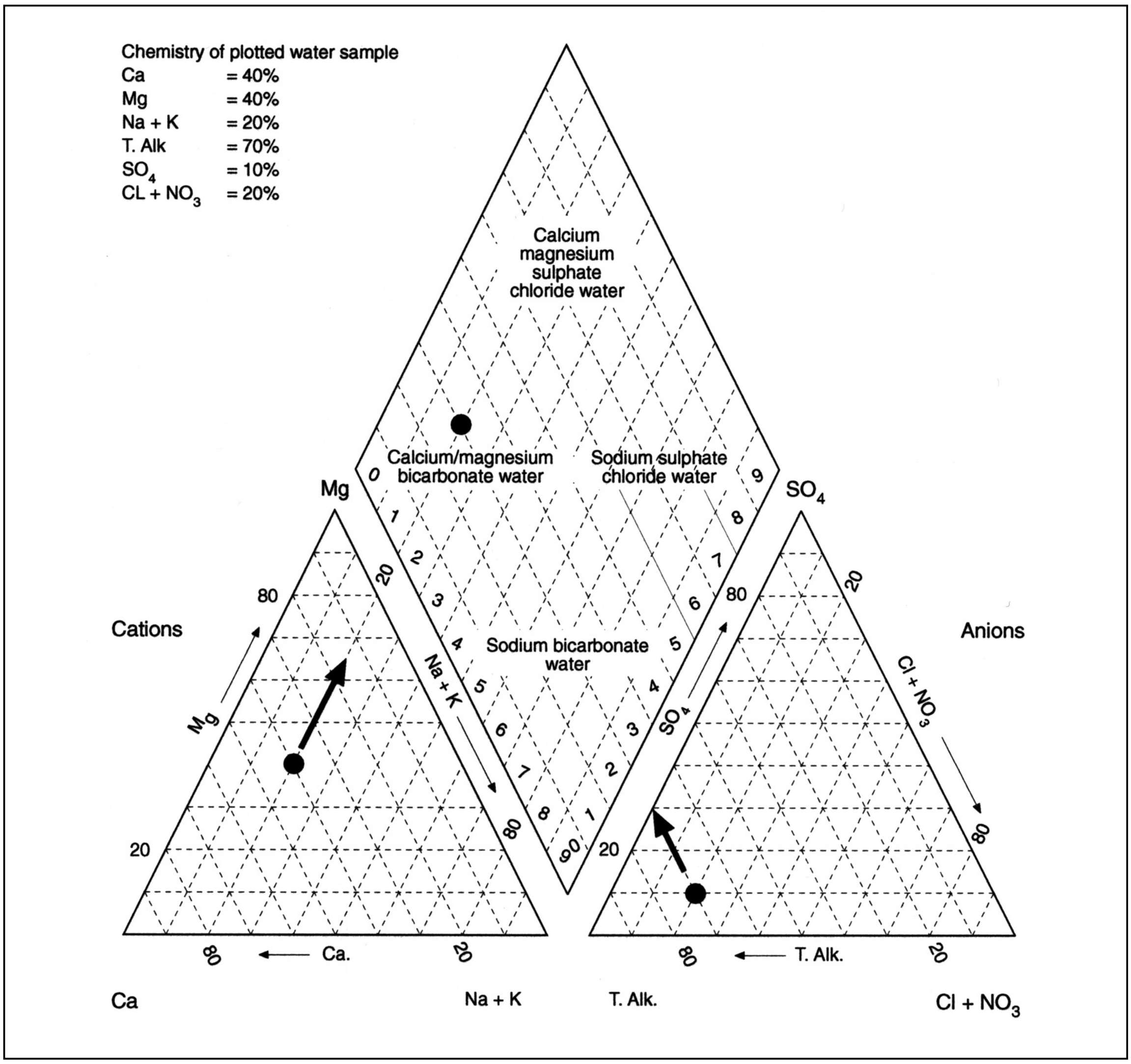
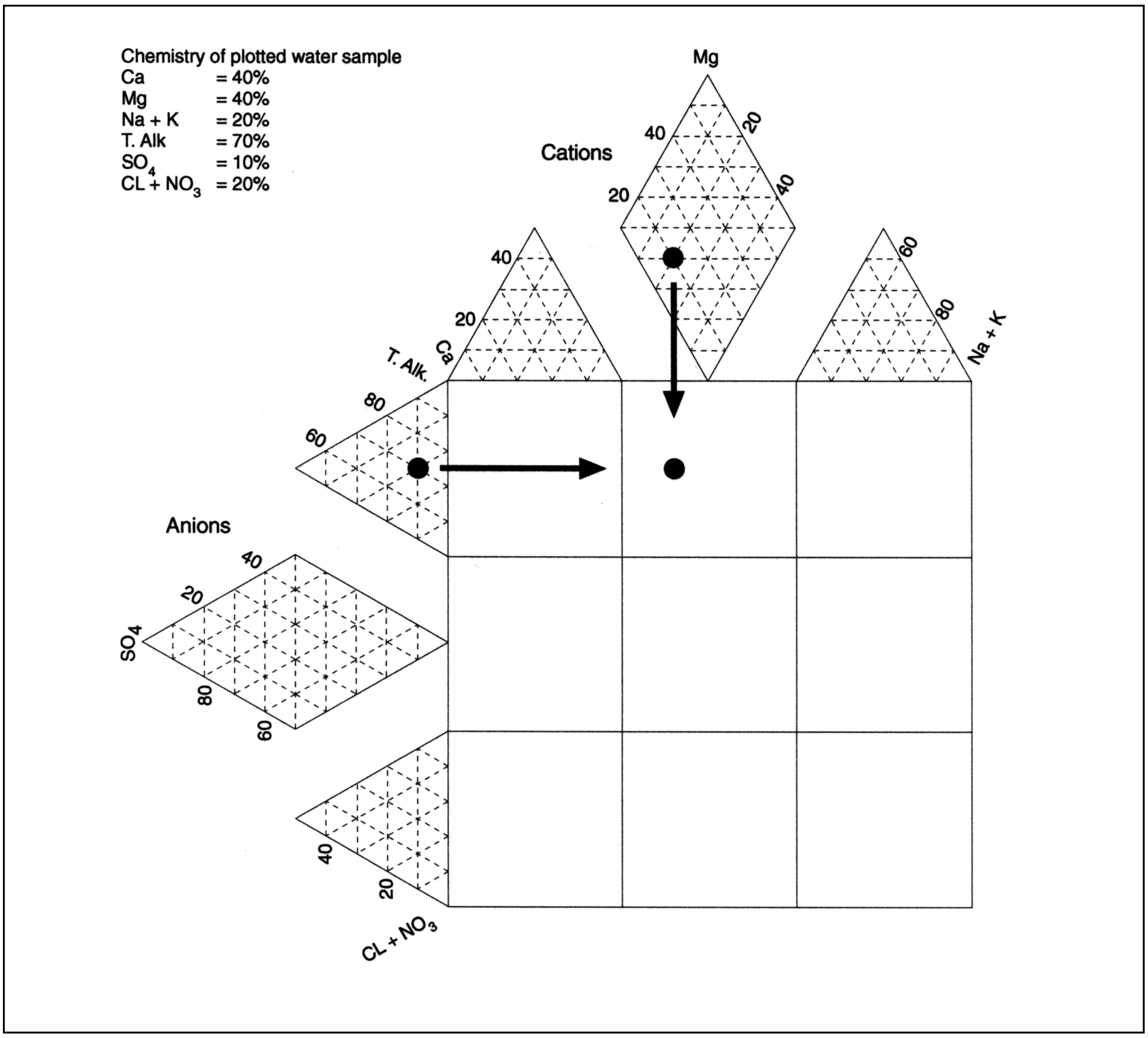


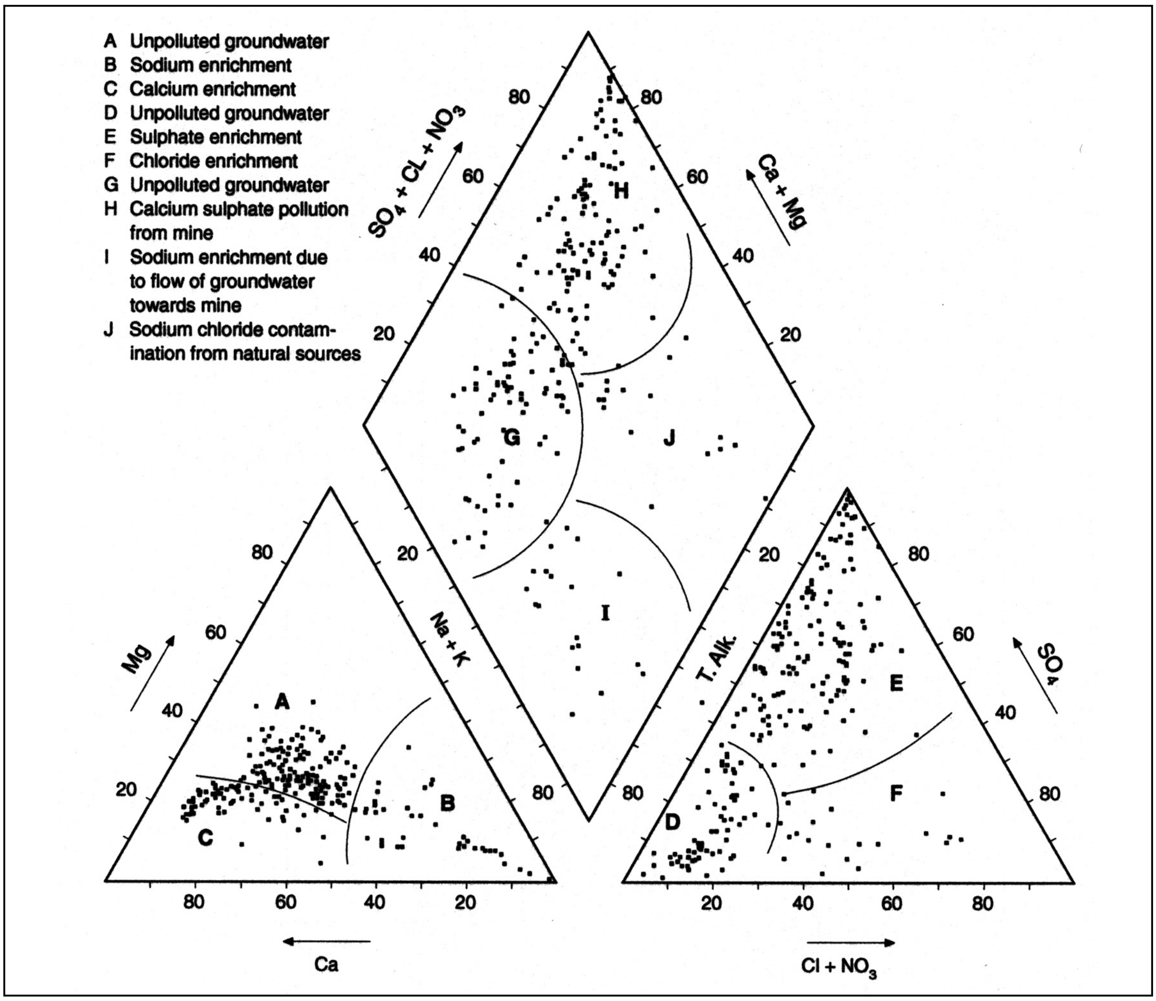
Figure 4.4 Plotting procedures for the Expanded Durov diagram: Convert mg/l to meq/l by division (Ca/20, Mg/12, Na/23, K/39, T.Alk./50, SO_4/48,Cl/35.5, NO_3/62. Add Na +K and Cl + NO_3. Calculate percentage of cations and anions. Plot cations by scaling off Ca, then Mg. Plot anions by scaling off T.Alk. then SO_4. Project cation and anion points to square



As an example of the advantages of these diagrams, a data set from a coal-mining environment has been selected and plotted in Figures 4.5–4.8. To demonstrate the kind of conclusions that may be derived from these plots, the following information is provided:

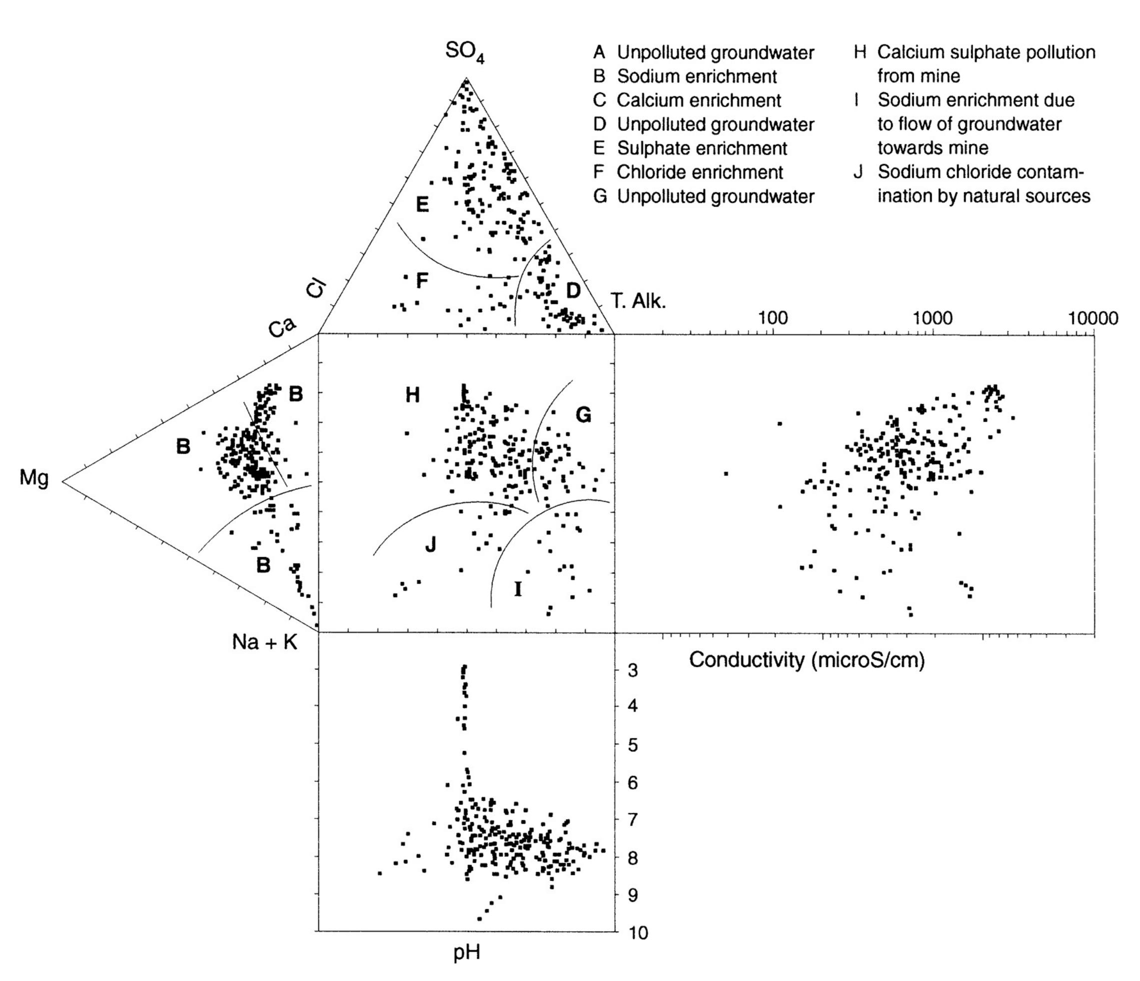
- Only about 30 per cent of the groundwater that has been sampled is unpolluted. The latter is characterised by water of a calcium/magnesium bicarbonate composition.
- The trend towards sodium enrichment, as depicted in the cation triangle, is due to cation exchange, as groundwater flows from surrounding aquifers towards the mine.
- The trend towards sulphate enrichment, as seen in the anion triangle, is due to pyrite oxidation in the coal-mine. This shows the extent to which the groundwater regime has already been polluted by mining.

Figure 4.5 Piper plot of groundwater chemistries from a mining environment



- The Durov diagram is particularly handy because it also includes the electrical conductivity of the water, thus giving a reflection of salt concentrations in the water. It suggests that the increase in the salt load is associated with mining activities.
- The calcium enrichment shown in the cation triangle of the trilinear diagrams originates from the neutralisation of acid water from the mine by calcium carbonate in the ground.
- The natural groundwater and aquifer have significant buffering potential against acidification, since only a few samples plot in the low pH range in the Durov diagram.
- The Expanded Durov diagram categorises waters into nine classes. For this reason, this diagram is for distinguishing between various groundwater populations, rather than for studying trends, as was the case for the diagrams discussed above.
- The SAR diagram suggests that sodium hazard to plants is low. Salinities are in the medium to high range. Crops with some resistance to salinity and soils with a sandy-loam character, may be irrigated successfully with the mine water.
- Nitrate pollution, which is derived from fertiliser application by farmers, is present in some borehole waters. This contamination cannot successfully be demonstrated by

Figure 4.6 Durov plot plot of groundwater chemistries from a mining environment



means of these diagrams, because of the dominance of other constituents. Other diagrams such as bar, line or scatter will have to be used for this purpose.

- The above example demonstrates the usefulness and limitations of these diagrams in the interpretation of hydrochemical data. Clearly, many variations of the above may be used to indicate values or demonstrate trends.

Through statistical methods, similar interpretations are possible. These methods will be discussed in Section 4.4. Statistical evaluations of this kind are usually complex and only understood by those familiar with statistical terminology. Reports for management should therefore always include diagrams, which may be backed up by statistical evidence if considered necessary.

Figure 4.7 SAR plot of groundwater chemistries from a mining environment

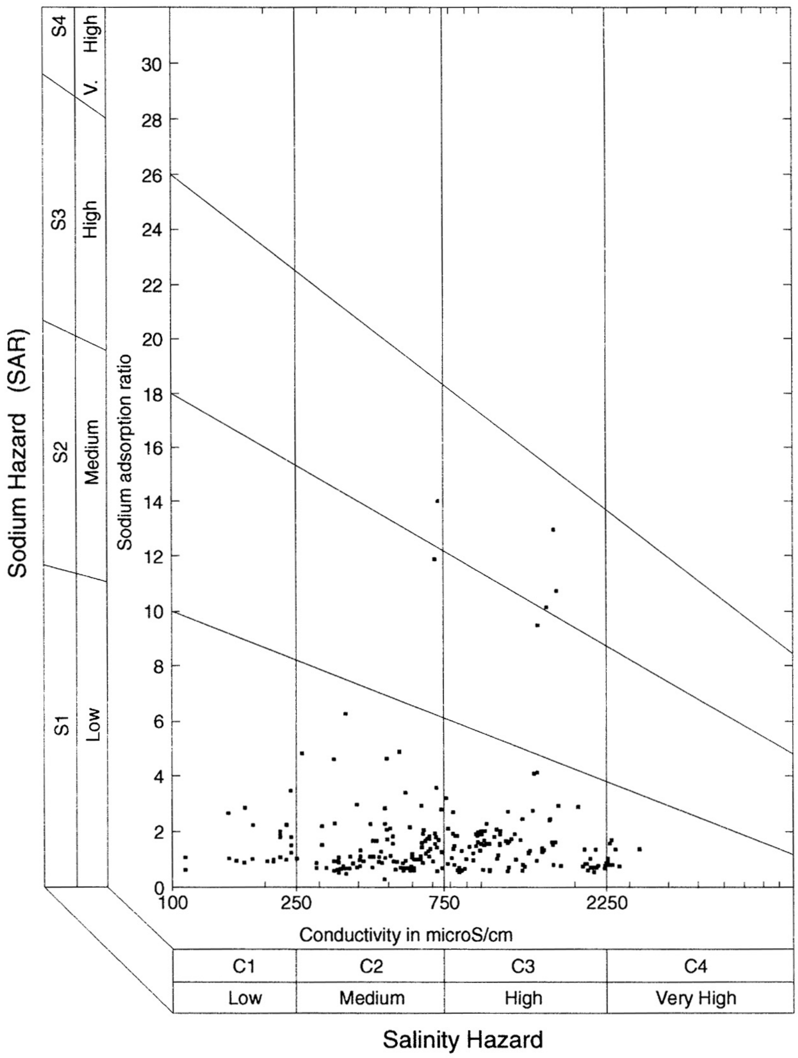
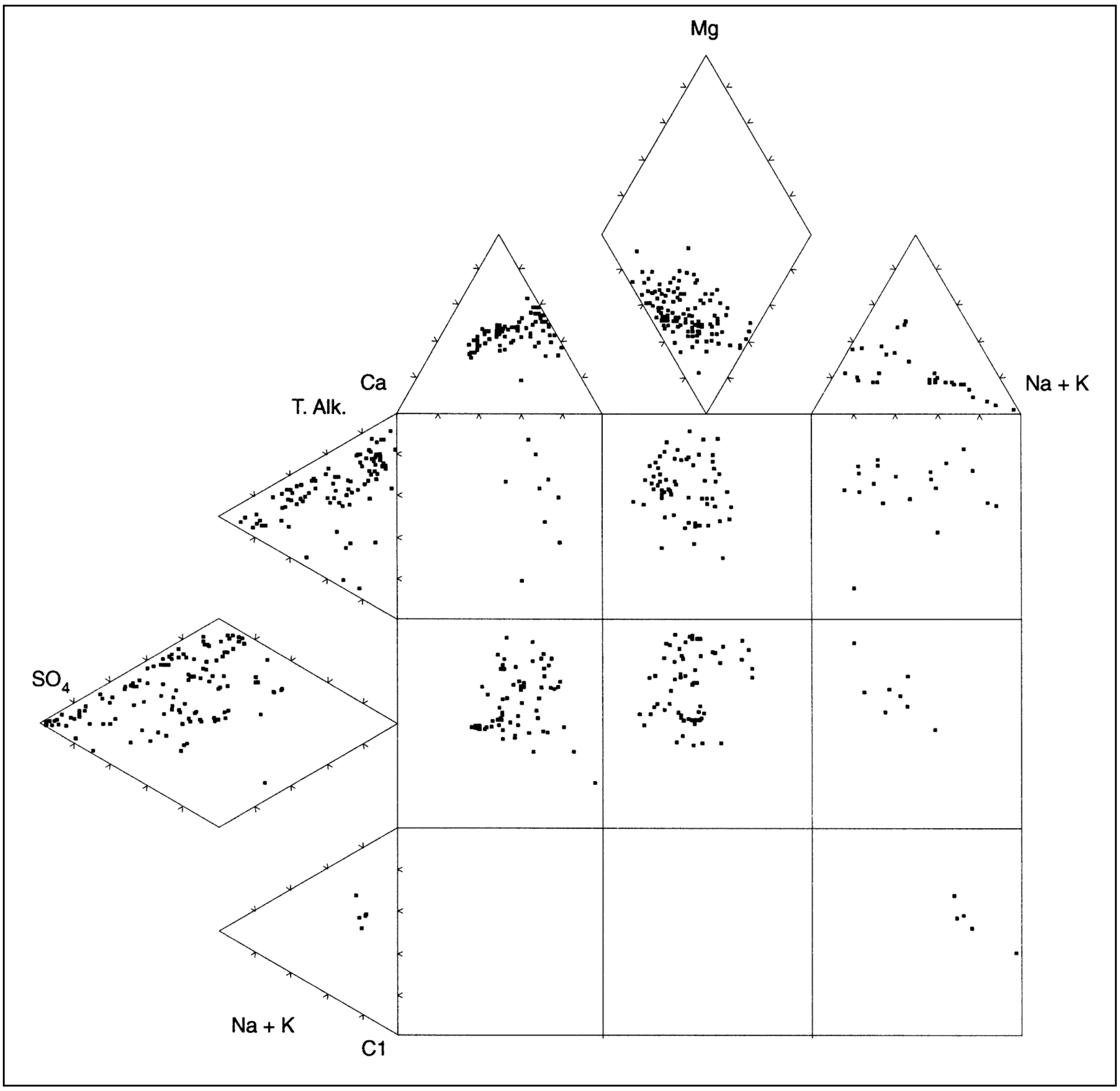


Figure 4.8 Expanded Durov plot of groundwater chemistries from a mining environment



4.3.5 Other graphics

Various possibilities exist to depict water chemistries for individual water samples. Presentations which have been used are the line (including Schoeller), scatter, bar, vector, radial, star, pie and polygon (including Stiff) diagrams (Lloyd and Heathcote, 1985) and Chernoff faces. Examples of some of these plots are presented in Figures 4.9–4.12. It is important to note that in most diagrams the units are meq/l, thus reflecting the true reactive ratios of the constituents. However, in the case of line, scatter and bar diagrams, the units may also be in mg/l. Results of chemical analyses are usually reported by laboratories as mg/l.

Business graphics such as line, scatter and bar plots are often used to depict hydrogeological data. These diagrams allow clear and concise presentations of a variety of aspects. By changing the y-axis of the plot from a linear scale to a logarithmic scale, the range of values that may be depicted becomes almost unlimited.

Figure 4.9 Stiff plots of water chemistries

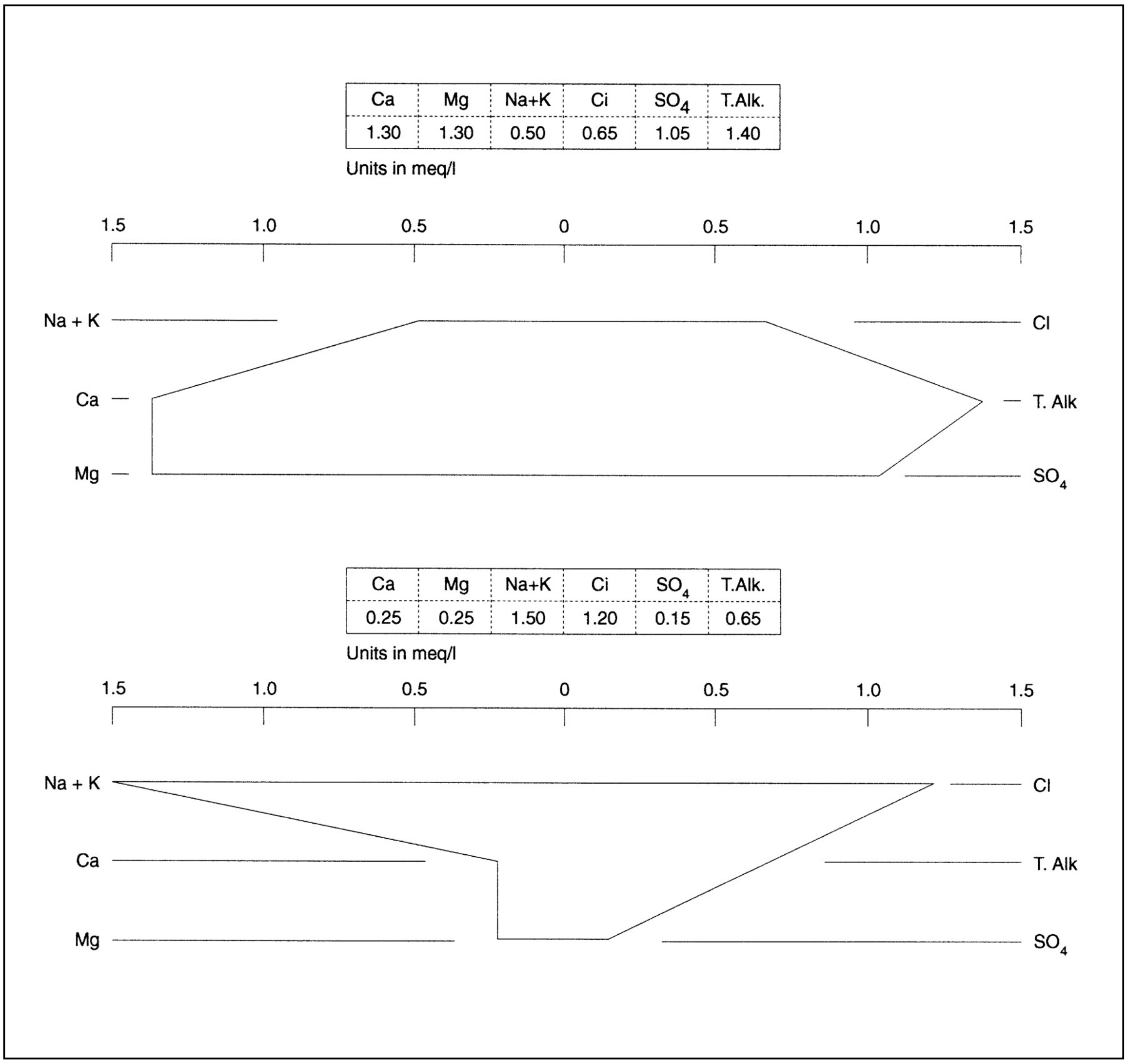


Figure 4.10 Pie diagrams of water chemistries

Ca	Mg	Na+K	Cl	SO_4	T.Alk.
1.30	1.30	0.50	0.65	1.05	1.40

Units in meq/l

Ca	Mg	Na+K	Ci	SO_4	T.Alk.
0.25	0.25	1.50	1.20	0.15	0.65

Units in meq/l

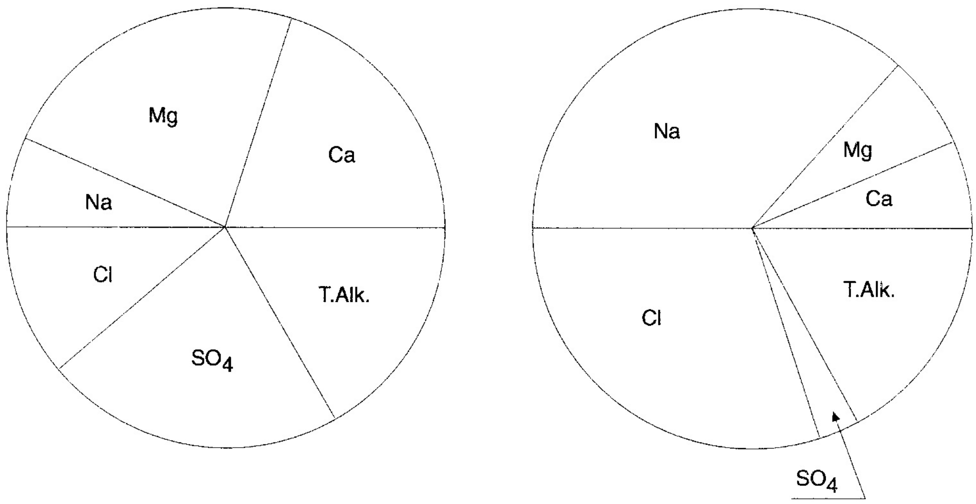


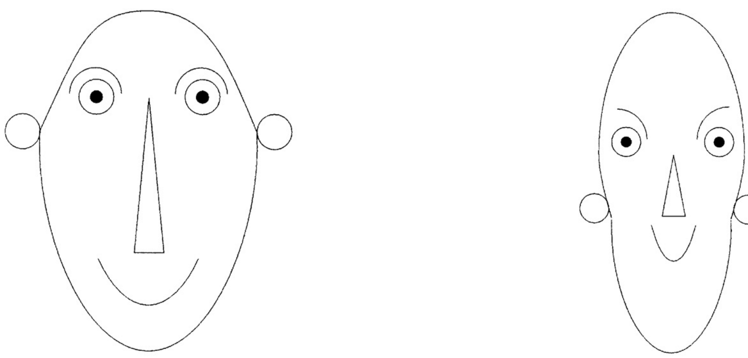
Figure 4.11 Chernoff faces of water chemistries

Ca	Mg	Na+K	Cl	SO_4	T.Alk.
1.30	1.30	0.50	0.65	1.05	1.40

Units in meq/l

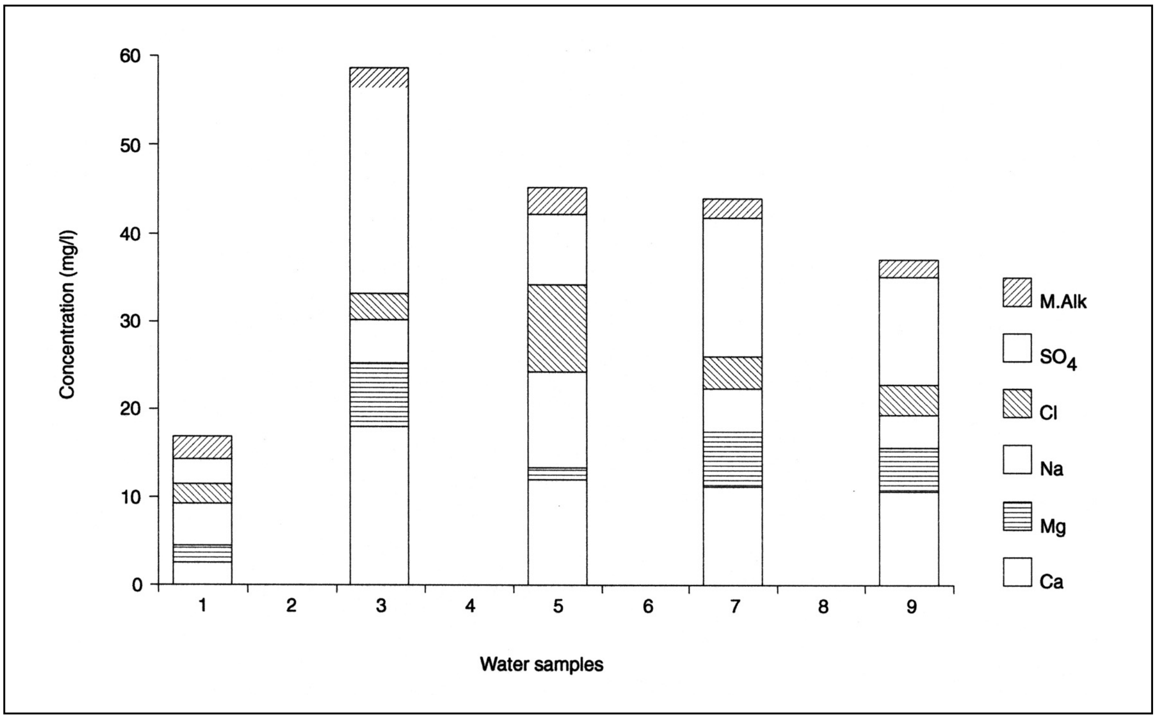
Ca	Mg	Na+K	Ci	SO_4	T.Alk.
0.25	0.25	1.50	1.20	0.15	0.65

Units in meq/l



face/w	= CA	upface/ecc	= CL
ear/lev	= MG	loface/ecc	= SO_4
halfface/h	= NA	nose/l	= T.ALK

Figure 4.12 Stacked bar charts of water chemistries



4.4 Statistical processing and presentation of data

Statistical interpretation of hydrogeological data has not received much attention in the past. The main reason for this probably lies in the fact that groundwater flow and pollution transport are governed by parabolic or elliptical differential equations, in contrast to surface water where stochastic processes play a major role.

4.4.1 Basic statistics

The term 'Basis Statistics' usually includes a variety of statistical interpretations, such as: mean, mode, median, standard deviation, standard error, correlations, probabilities, t-test and analysis of variance. An introduction to basic statistics can be found in handbooks such as Kachigan (1986), and Runyon and Haber (1976). For more advanced discussions on the elementary theory of statistics, the books of Hays (1988) and Kendall and Stuart (1979) may be consulted.

The results of many of these tests can be incorporated into a single diagram, referred to as a 'box and whisker plot'. A box and whisker plot usually depicts five statistical characteristics for each variable (Figures 4.13–4.14). The central dots represent the mean, median or mode of the variables. Around these points, boxes are constructed that show the upper and lower limits of a statistic, such as the 25 and 75 percentiles or the standard errors of the means. The widths of the boxes are arbitrary. Lastly, lines that extend as whiskers above and below boxes typically reflect standard deviations or minima and maxima of variables. Box and whisker plots are generally recognised for their usefulness in visual comparative statistics for frequently measured variables.

Correlations amongst variables may be established by calculating correlation coefficients. This application is particularly useful for studying water chemistries and the interdependence of constituents within waters (Table 4.1).

Figure 4.13 Box and whisker plot of groundwater chemistries from a mining environment

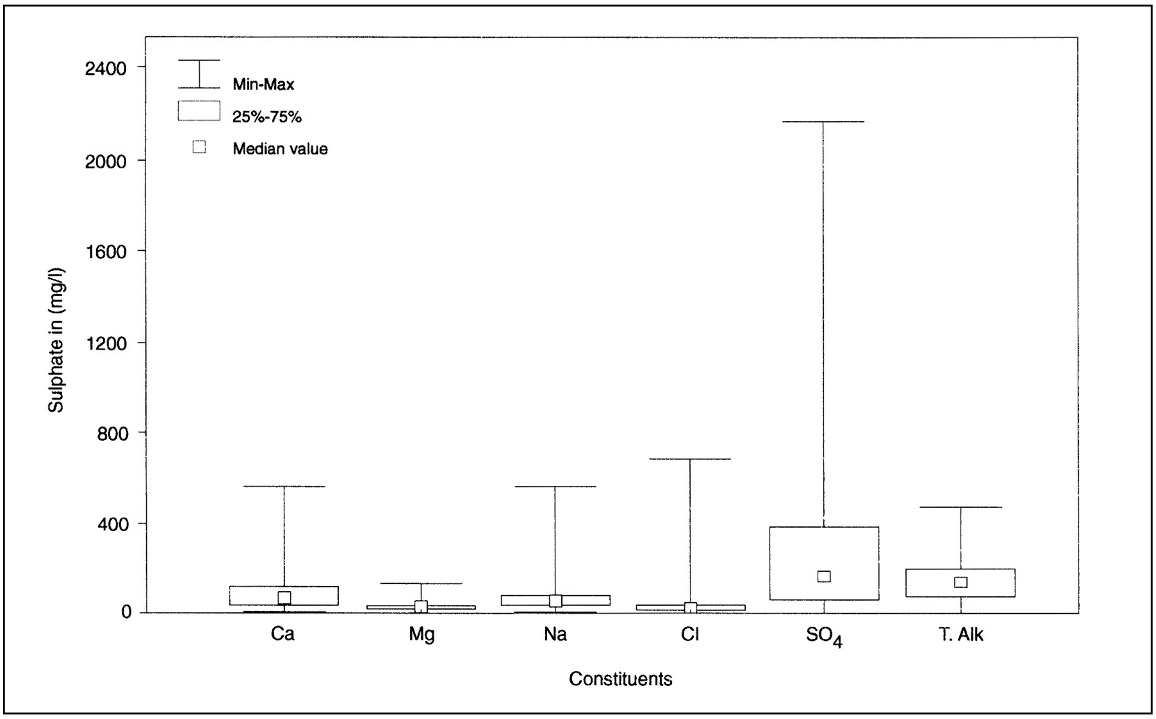


Figure 4.14 Box and whisker plot of groundwater chemistries from a mining environment

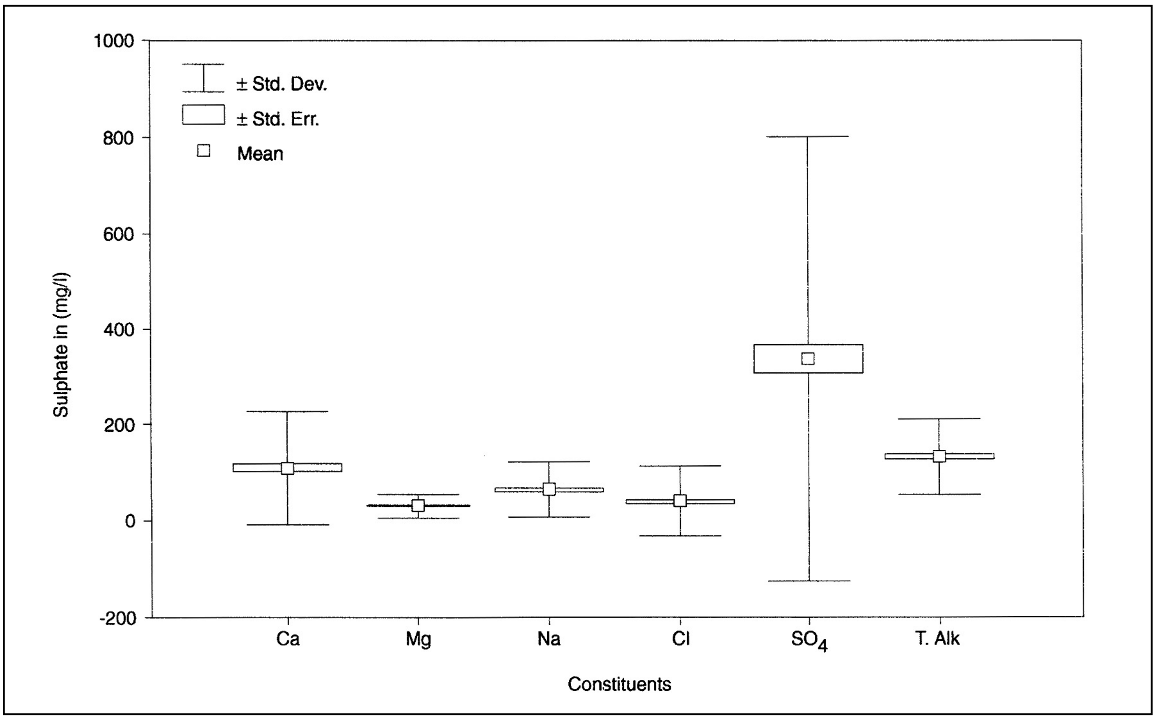


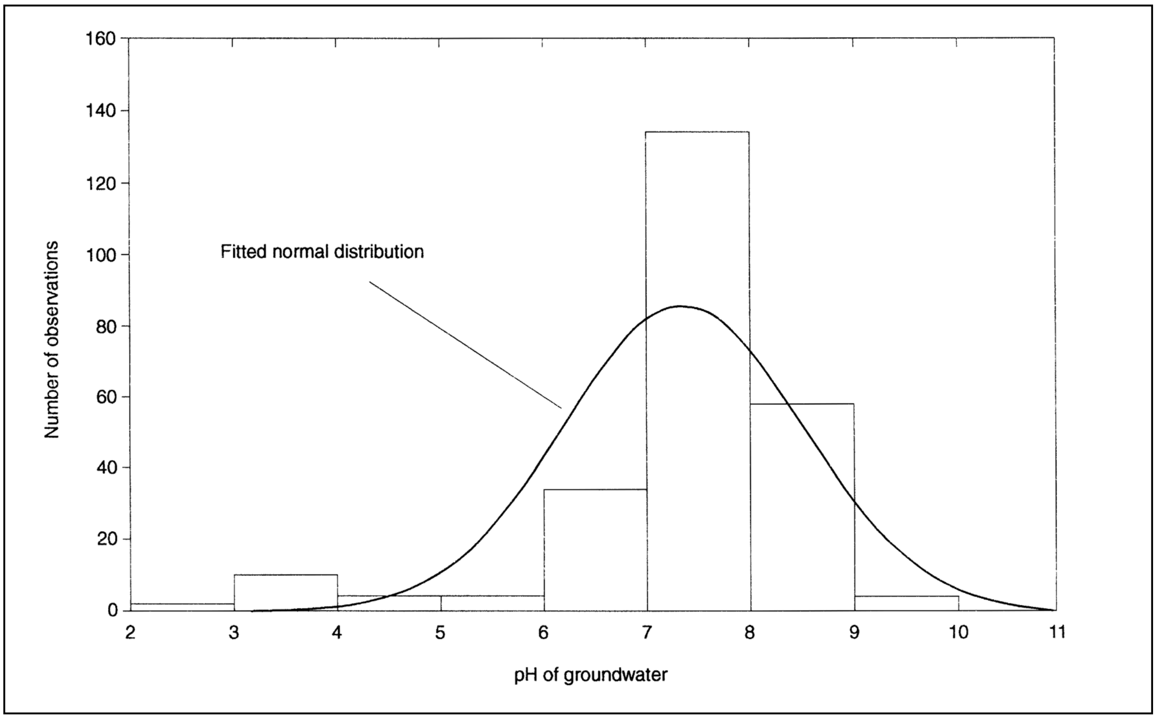
Table 4.1 Correlations between chemical constituents in water from a mining environment (sample size 250)

	Ca	*Mg*	*Na*	*Cl*	SO_4	*Total alkaline*
Ca	1.00	84*	0.15	0.07	0.97*	0.40
Mg	0.84*	1.00	0.21	0.12	0.85*	0.19
Na	0.15	0.21	1.00	0.90*	0.17	0.23
Cl	0.07	0.12	0.90*	1.00	0.07	0.13
SO_4	0.97*	0.85*	0.17	0.07	1.00	0.50
Total alkaline	0.40	0.19	0.23	0.13	0.50	1.00

∗ Denotes a correlation better than 0.70.

Histograms may be used to display distributions of variables (Figure 4.15). Of particular value is the bivariate histogram (Figure 4.16), in which two variables may be considered simultaneously. The interrelationship between sulphate concentrations and the pH of the water becomes apparent in the latter presentation.

Figure 4.15 Histogram of groundwater pH levels from a mining environment



4.4.2 Time series

Many hydrogeological variables are time-dependent. This data may therefore be analysed by time series methods, such as trend analysis and correlation methods.

Most commonly, polynomial regression methods are used to depict trends (for example,in water-level fluctuations, for instance: Figure 4.17). A comparison between water-level responses in different boreholes may be done by cross correlation, while auto correlation may be used to investigate possible repetitive responses (Box and Jenkins, 1976).

Figure 4.16 Bivariate histogram of groundwater pH and sulphate from a mining environment

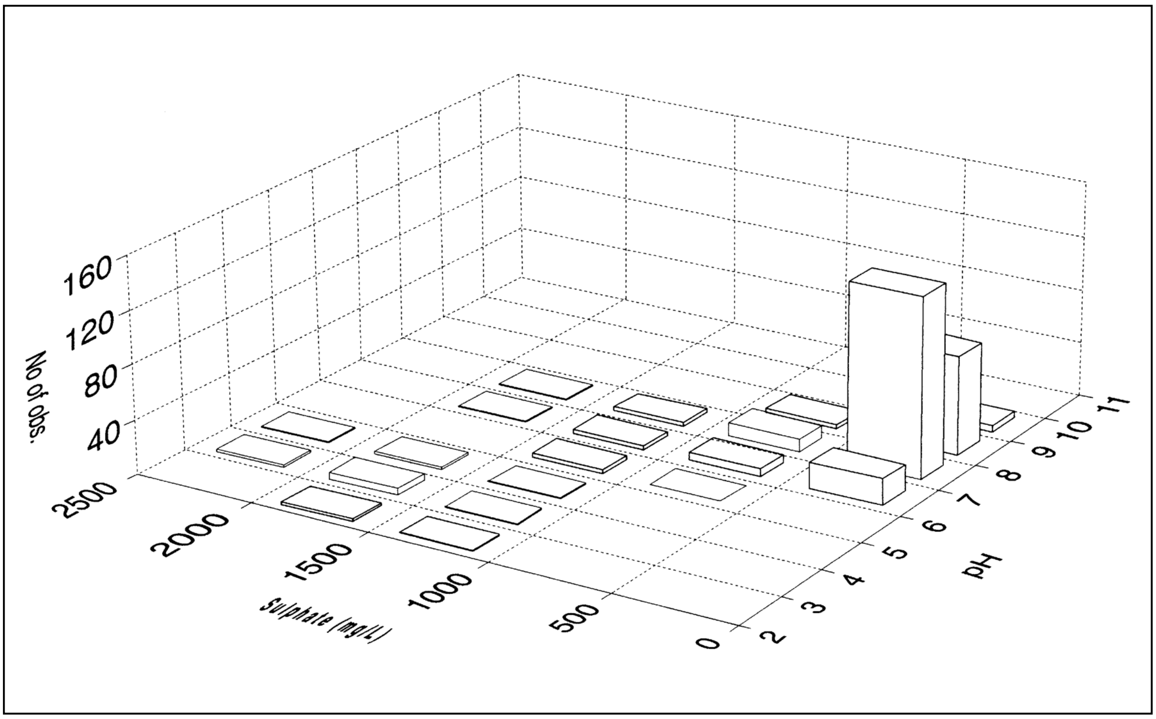
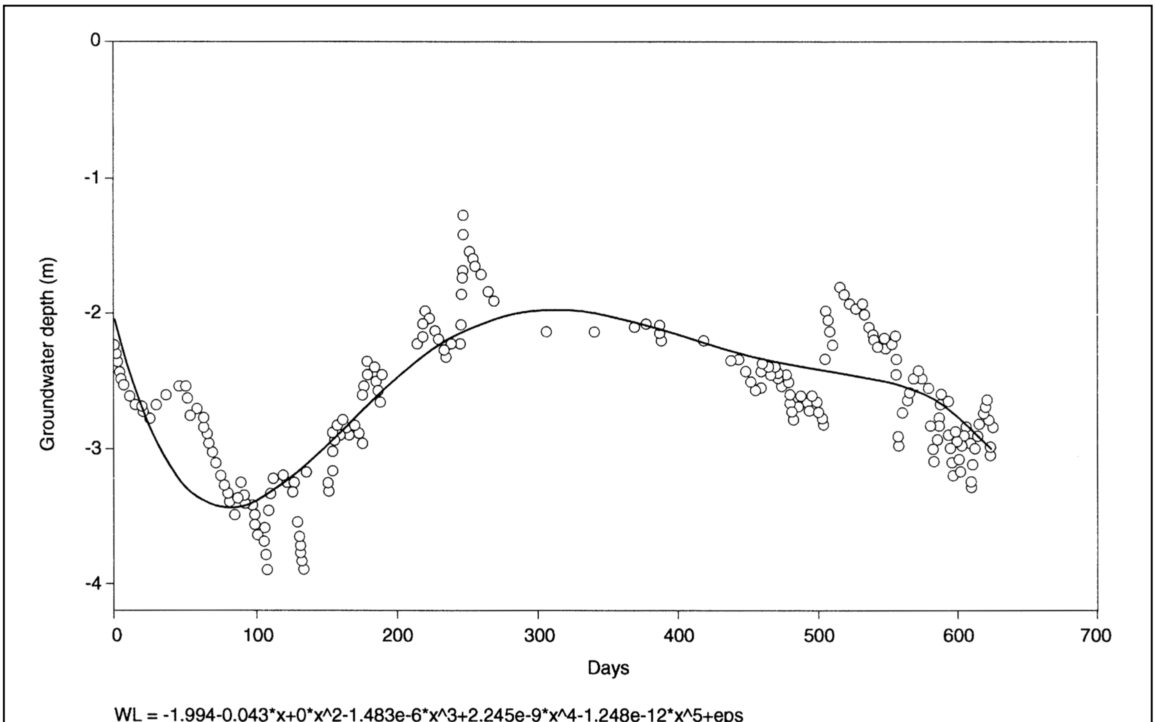
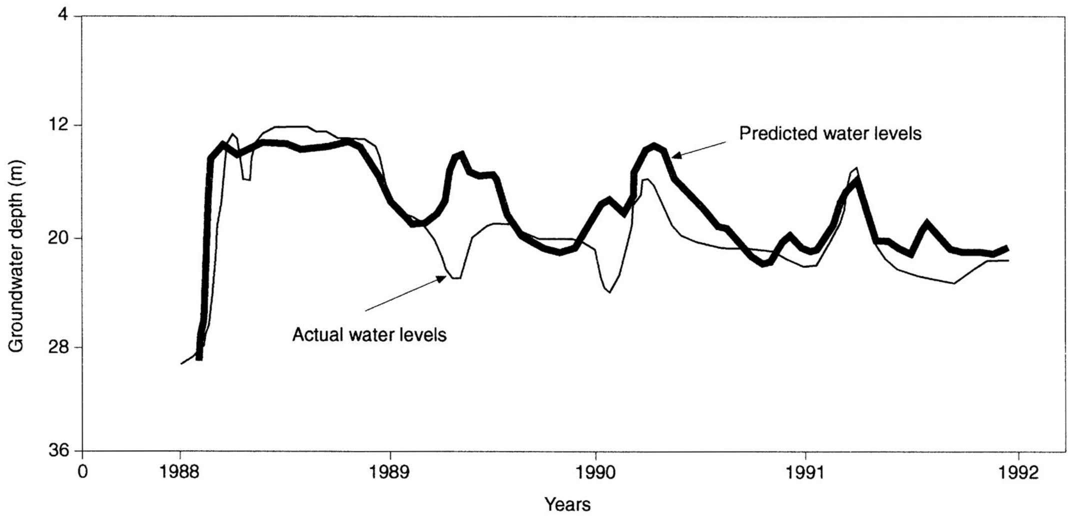


Figure 4.17 Fifth order polynomial regression interpolation of groundwater levels



Multivariate analysis can take on many forms. A commonly used predictive tool is multiple linear regression.It enables a number of variables to be related to a single variable that needs to be predicted (Hodgson, 1978). A typical example of how water levels may be predicted and related to rainfall and abstraction is shown in Figure 4.18. The equation used for the prediction of the water levels indicates a time lag of two months between the rainfall event and maximum recharge to the aquifer. It should be stressed that prediction of this kind is only valid within observed limits. If, for instance, a rainfall event occurs that exceeds all previous rainfall intensities, predictions may no longer be valid.

Figure 4.18 Multiple linear regression showing actual and simulated water levels



$\hat{y} = 12.963735 + 0.000056\, x_{1t} - 0.041470\, x_{2(t-2)} + 0{,}394976 \hat{y}_{(t-1)}$

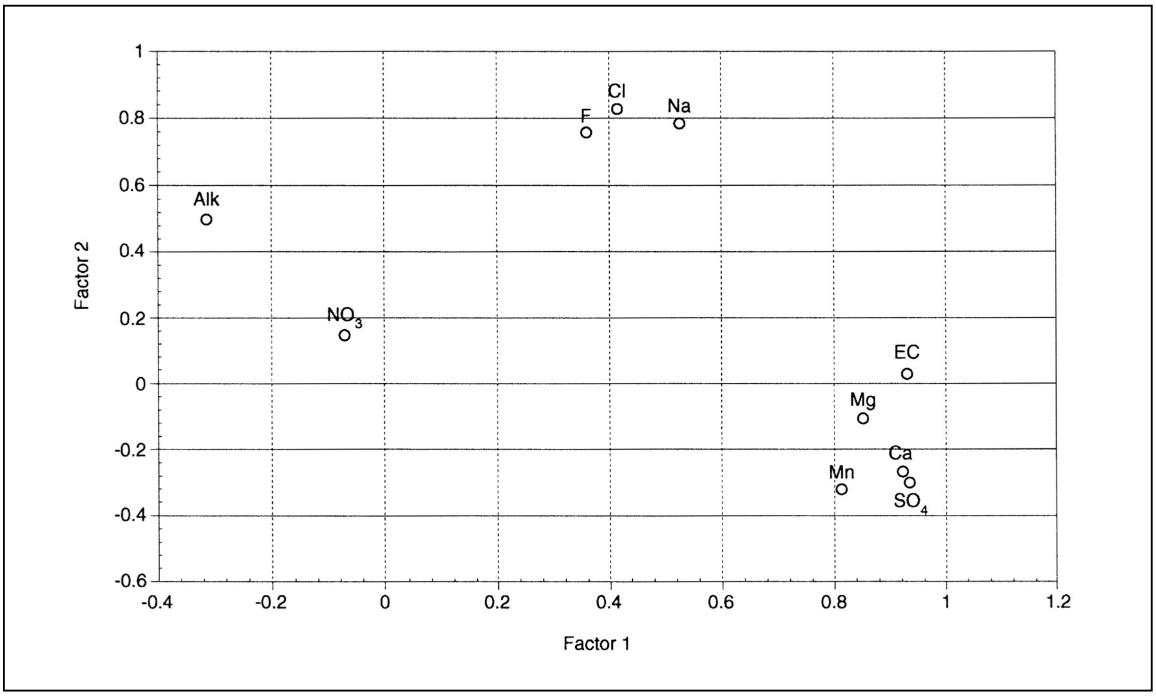
where x_1 = pumpage; x_2 = rainfall; t = months

Another form of multivariate analysis is that of grouping of variables that show relational trends. Test statistics which may be used by the hydrogeologist include principal component analysis, factor analysis and cluster analysis. All three of these evaluative procedures are based on the principle of recognising patterns in data sets that contain many observations (Stevens, 1986). These techniques are excellent in, for instance, distinguishing between different populations. The results are similar to those obtained diagrammatically from the Piper and other specialised hydrochemical diagrams.

Figure 4.19 is a factor analysis of the water quality data for the mining environment. The close interrelationship between mine pollution variables (electrical conductivity, Ca, Mg, SO_4 and Mn), versus nitrate pollution from farming and the natural pollution (Na, Cl and F) is demonstrated by the groupings of the variables in the diagram.

Using advanced multivariate statistics has its pros and cons. An advantage of these procedures is the wide range of tests that may be performed by informed statisticians The disadvantage is that very few managers and planners understand the significance of these tests. Many prefer graphs or maps above complex statistical arguments. This situation will, hopefully, change, as more and more individuals become familiar with statistical procedures. Many current statistical computer software packages are accompanied by well-documented manuals, making the use of statistics easy and enjoyable.

Figure 4.19 Factor analysis of groundwater chemistries from a mining environment



4.5 Spatial information systems

4.5.1 Introduction

Spatial information systems have undergone significant development during the past five years (Fulton, 1992; Haefner, 1992; Juracek, 1992). In its computerised form, these systems are referred to as 'Geographic Information Systems' (GIS). GIS in its simplest form constitutes two essential components, namely a map drafting facility and a related data base. All elements, such as points, lines, arcs and text that are drafted, are recorded as spatial entities within the data base.

All information in a GIS is structured. Structuring is done according to certain definable conditions. A basic condition is usually that variables of the same type are stored together, to allow ease of data retrieval. Each storage unit is usually referred to as a coverage. Hydrogeological data and information such as water levels, positions of boreholes and distribution networks may therefore constitute separate coverages as part of a GIS data base.

One of the main advantages of standardising on GIS is that hydrogeological information may be combined with other information sets. Typical information sets that are of use to the hydrogeologist could be topography, soils, geology, meteorology, urban development, waste disposal and mining and industrial development. The latter information is generally available within other institutions where GIS has been in place for some time. It therefore does not need to be entered by the hydrogeologist.

The advantage of having complementary information sets available at the start of a hydrogeological investigation, is enormous. It allows, amongst other things, the identification of target areas where hydrogeological investigations would be meaningful. As an example of the versatility of such a system, the following constraints may for instance be imposed onto existing information within the GIS, for the siting of a domestic waste facility:

- it should not be closer than 1,000 m from a water-supply borehole;
- it should not be within 500 m of urban or nature conservation areas;

- it should not be on primary aquifers, chalk or dolomite;
- the natural groundwater level should be at least 2 m below the surface;
- the area should not have a surface gradient greater than 1:100;
- it should be accessible from an existing highway;
- sufficient clay should be available within a 2 km radius, for leachate control.

GIS is particularly well-suited for the above evaluation. It will, within seconds, identify all the areas that conform to these requirements and show them on the computer screen. Detailed hydrogeological investigations may then be concentrated within these areas, thus effecting a considerable cost savings in terms of initial planning and site visits.

Many other scenarios may be investigated using GIS. The only constraints are:

- the availability of information sets;
- the quality of the information.

4.5.2 Entry of spatial information

Entry of information into GIS is complex and time-consuming. Point data may be imported electronically from existing databases. Existing maps either have to be digitised or scanned. Digitising involves the physical tracing of all lines, points and text into GIS coverages. Electronic scanning of existing maps is initially much faster. This provides a raster image that may be used as a coverage in the GIS. Raster images are not true GIS images and no intelligence can be attached to them. They consist of millions of dots, with each dot representing a single scanned pixel. Such images occupy vast storage space on computer disks. In view of these limitations, it is almost always necessary to convert raster images to vector images.

Vector images may be obtained from raster images by activating suitable computer software. This software is capable of recognising patterns and converting them into lines, text, symbols and shadings. Vectorisation of raster images is a major task, mainly due to the relatively poor printed quality of most maps. It may, for instance, take more than a month to vectorise a single topographical map. Nevertheless, scanning and vectorising is the preferred technique for inputing information into a GIS.

4.5.3 Conversion of point data

The conversion of point hydrogeological data into spatial information is the next logical step in GIS application. This is most commonly done by contouring point values. Examples of variables that may be contoured are transmissivities, storage coefficients, hydrochemical data and water levels.

Various methodologies and numerous software packages are available for contouring of data. Most of these methodologies are based on the transformation of irregularly spaced data to regularly spaced values. This may be typically done through inverse distance or kriging methodologies. Kriging provides the added advantage that confidence limits may be attached to contours – a facet that is often overlooked during contouring. A set of contours should therefore ideally be accompanied by a second set of contours that show upper and lower limits for contoured values.

Another aspect of contouring which is easily overlooked is that contouring should not be done in areas where insufficient data are available. A check on the sufficiency of data may be built into computerised programs, by limiting search distances. Areas where insufficient data are available will thus be identified and no contours will be drawn there (Figures 4.20–4.21).

Contours should preferably be accompanied by a posting of the positions where data are available, thus explaining the absence of contours in certain areas.

Three-dimensional contours are sometimes used to demonstrate specific aspects. It is

Figure 4.20 Smoothed water-level contours, using a 5,000 m search radius

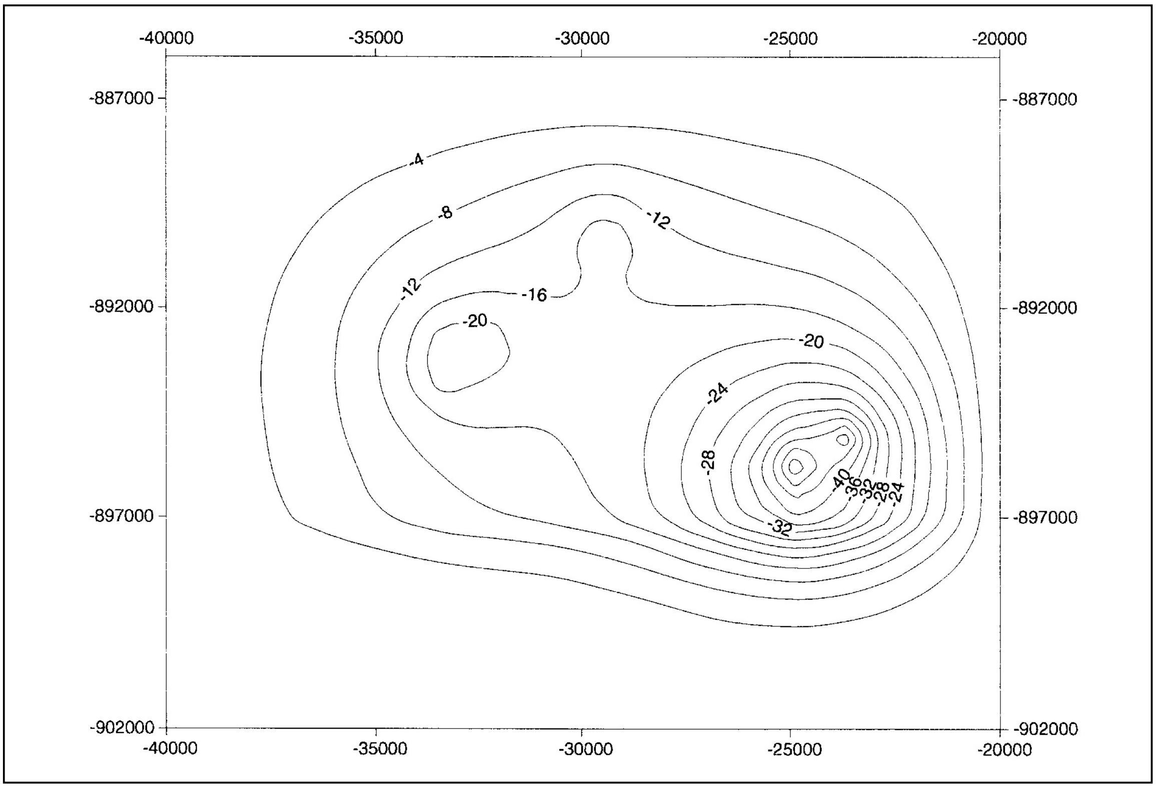
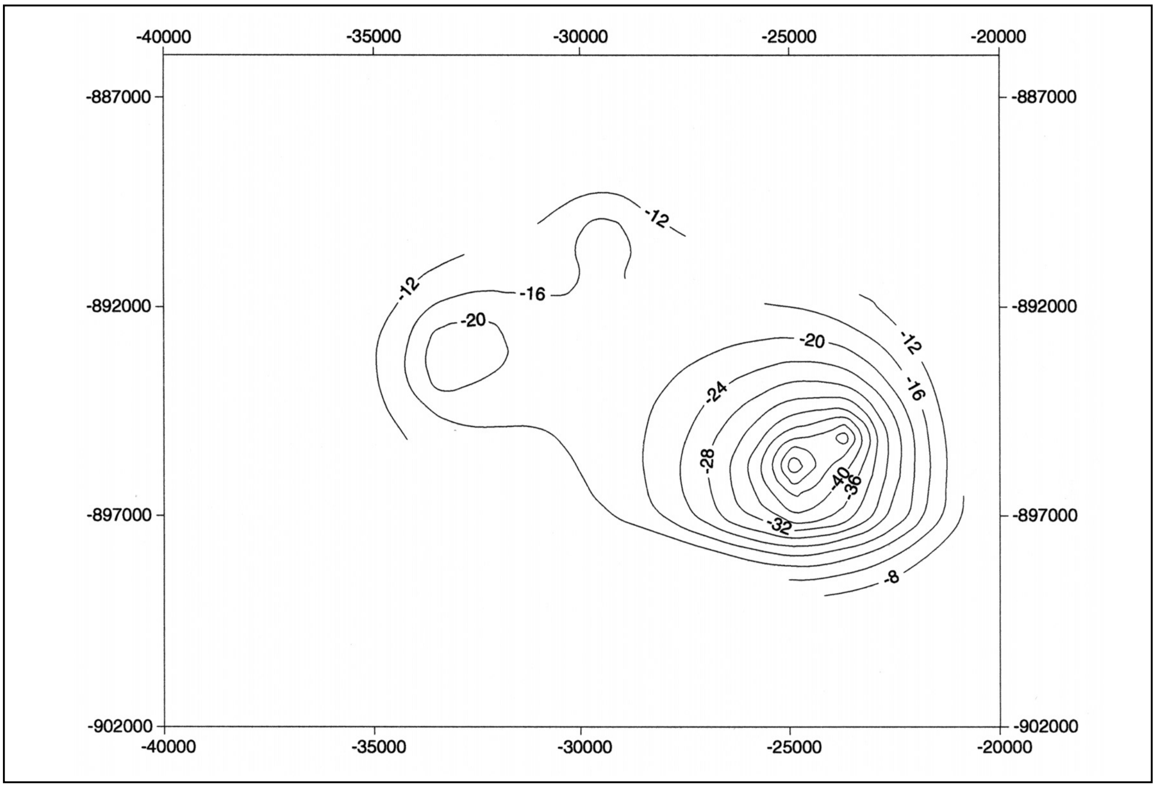


Figure 4.21 Smoothed water-level contours, using a 500 m search radius



difficult for the viewer to orient himself with respect to these presentations. For this reason, these contours should be accompanied by (i) two-dimensional contours, (ii) a vertical scale and (iii) an indication of the drawing orientation. In addition, equi-potential contours may be embedded onto the three-dimensional surface, thus enabling the quantification of variations in the diagram (Figures 4.22–4.23).

A recent development in GIS is the transformation of spatial information into grid images. In these applications, the grid images are very coarse, typically consisting only of a 40 x 30 matrix. The purpose of such transformations is to convert spatial information into averaged grids (Figure 4.24) which in turn, may again be analysed by using standard GIS tools. A typical application would be the transformation of water-level and topographic information into two raster images of identical dimensions. Thereafter, the interrelationship between these two variables may be studied. The application of grid procedure particularly lies in the comparison of complex information sets.

The combination of GIS technology with hydrogeological tools such as specialised interpretations and modelling, has been under development for a number of years. Examples are the NWIS-II (USGS, 1991), REGIS (TNO, 1993; NITG-TNO, 2000) and GGIS (IGS, 1994) software. All three software packages will have the capability of converting point data to spatial data, integrating data sets with GIS and executing flow and mass transport models from within the GIS environment.

4.5.4 Hydrogeological maps and the UNESCO code

Hydrogeological maps to depict groundwater related information (see also Chapter 10) have been around for many years. Examples of such maps are plentiful (UNESCO, 1983). UNESCO suggested standardising hydrogeological codes and maps in 1963; they were revised in 1983. These publications have been well received and many maps have been produced using this code or variations of it.

One of the many examples where the UNESCO code has been used is the Hydrogeological Map of England and Wales (United Kingdom, 1977). Many other hydrogeological maps have been published of Europe and other parts of the world. The UNESCO code of 1983 lists numerous hydrogeological maps world-wide.

The tendency for hydrogeological maps is to have a large central base map showing some hydrogeological variables. Surrounding this map are other map inserts. In the case of the Lesotho Map, for instance, (Ardino 1994), information on the following topics has been included on the map:

- hydrogeological legend;
- summary of hydrogeology for Lesotho;
- climatic information, including rainfall time series, contours and wind directions;
- base flow and run-off information;
- water quality information tabled and plotted onto Piper diagrams;
- geological and hydrogeological sections;
- groundwater contours and flow nets;
- tables showing individual and interesting characteristics that relate to the geohydrology at specific localities on the map.

Other features that may typically be considered are groundwater vulnerability in terms of over-exploitation and pollution.

In view of the great number of hydrogeological maps that are being produced world-wide, standardisation of colours and codes is recommended. It is recommended that the UNESCO codes of 1963 and 1983 should be used as closely as possible.

Figure 4.22 Three-dimensional presentation of groundwater levels

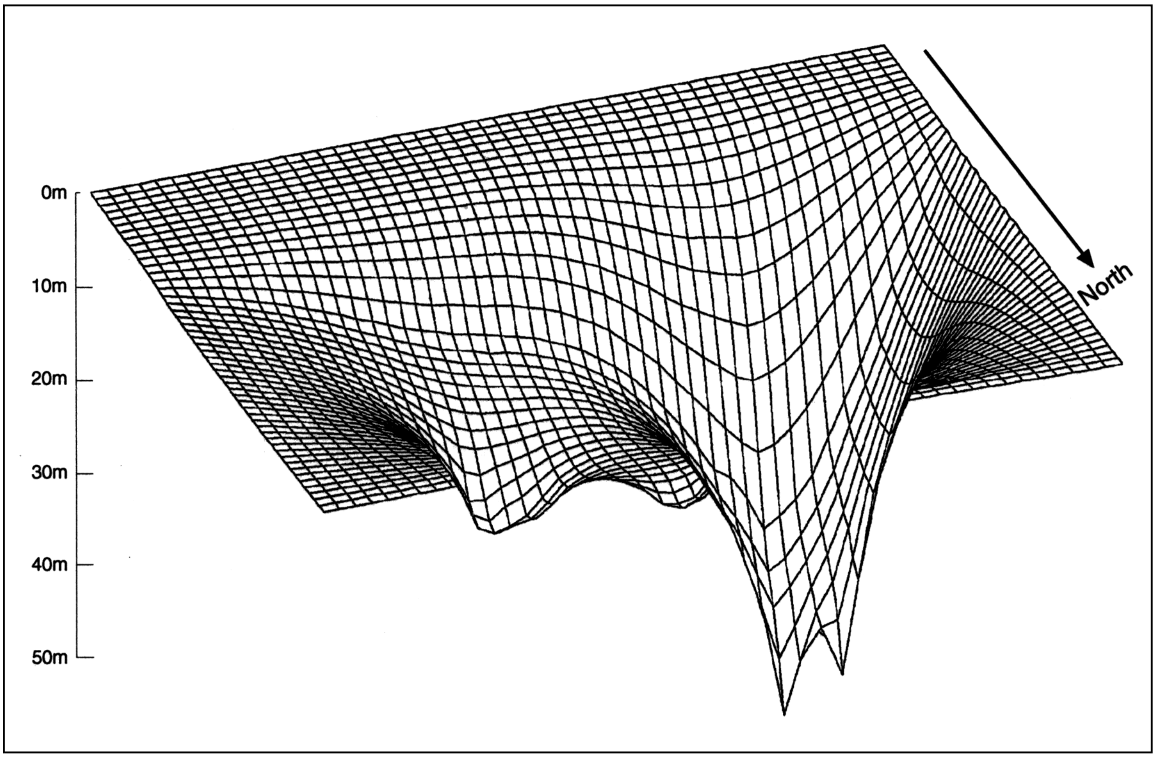


Figure 4.23 Three-dimensional presentation of groundwater contours at metre intervals

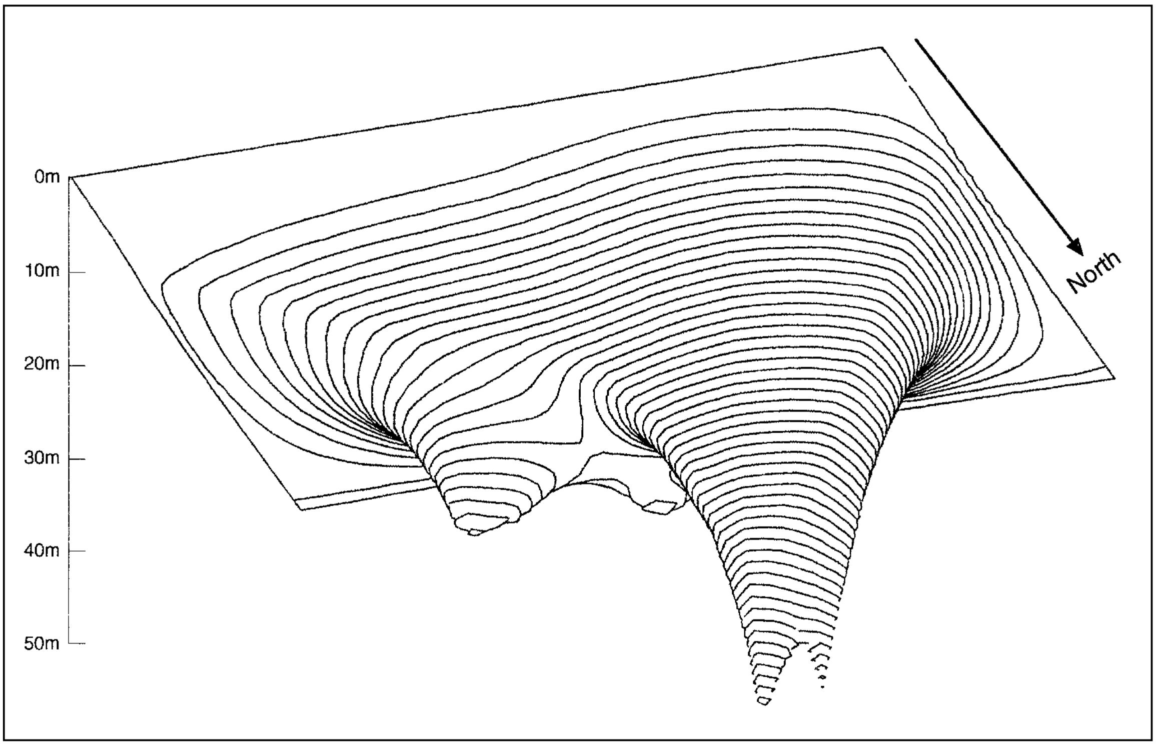
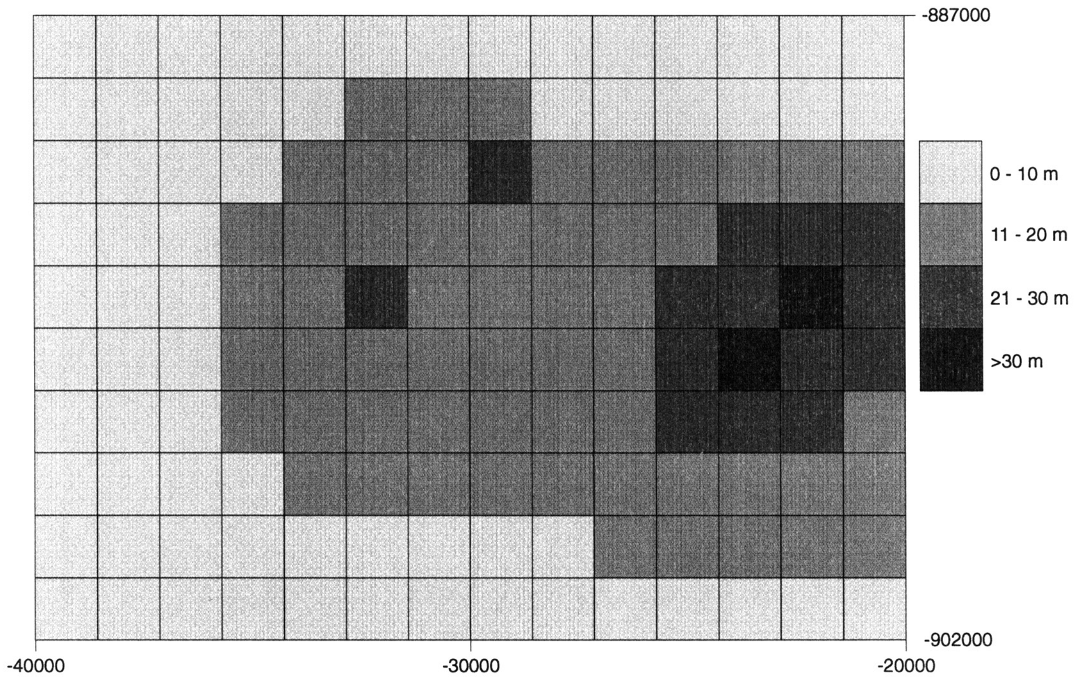


Figure 4.24 Grid presentation for evaluation of averaged groundwater levels



4.6 Conclusions

In terms of hydrological data processing and presentation, the only limit is the imagination of the individual dealing with these matters. In this Chapter, an attempt has been made to present the techniques that are available to everybody. Many more sophisticated techniques are available – often requiring significant computing power. When searching for a specific processing and presentation technique, the norm should always be:

- the level of expertise of the individual;
- the quality of the data;
- the availability of equipment such as computing equipment;
- the level of expertise of the individuals to whom the information is to be presented.

4.7 References and additional reading

ARDUINO, G.; BONO, P.; DEL SETTE, P., 1994. *Hydrogeological Map of Lesotho, Scale 1:300,000.* Department of Water Affairs, Maseru.

BOWER, C. A.; OGATA, G.; TUCKER, J. M., 1968. Sodium Hazard of Waters as Influenced by Leaching Fraction and by Precipitation or Solution of Calcium Carbonate. *Soil Science,* Vol. 106, pp. 29–34.

BOX, G. E. P.; JENKINS, G. M., 1976. *Time Series Analysis: Forecasting and Control.* Holden-Day, San Francisco.

DUROV, S. A., 1948. Natural Waters and Graphic Representation of their Composition. *Dokl. Akad. Nauk SSSR*, Vol. 59, pp. 87–90.

FULTON, J. L., 1992. Development of Spatial Data Guidelines and Standards: Spatial Data Set Documentation to Support Hydrological Analysis in the U.S. Geological Survey. *Geographic Information Systems (GIS) and Mapping – Practices and Standards,* STP 1126, ASTM, Philadelphia, PA, pp. 30–7.

HAEFNER, R. J., 1992. *Use of a Geographic Information System to Evaluate Potential Sites for Public-Water-Supply Wells on Long Island, New York*. Open-file Report 91-182, USGS, Denver, 37 pp.

HAYS, W. L., 1988. *Statistics (fourth edition).* CDS College Publishing, New York.

HODGSON, F. D. I., 1978. The Use of Multiple Linear Regression in Simulating Groundwater Responses. *Ground Water*, Vol. 16, pp. 249–53.

IGS, 1994. *GGIS – A Geographic Groundwater Interpretation System under Arc/Info.* Open-file Report to the WRC, Institute for Groundwater Studies, Bloemfontein, South Africa, 120 pp.

JURACEK, K. E., 1992. *Use of a Geographic Information System to Assist with Studies of the Availability and Use of Water in Kansas.* Open-file Report 92-142, USGS, Denver, 14 pp.

KACHIGAN. S. K., 1986. *Statistical Analysis: An Interdisciplinary Introduction to Univariate and Multivariate Methods.* Redius Press, New York.

KENDALL, M.; STUART, A., 1979. *The Advanced Theory of Statistics (Vol. 2).* Hafner, New York.

KRUSEMAN, G. P.; DE RIDDER, N. A., 1991. *Analysis and Evaluation of Pumping Test Data (second edition).* International Institute for Land Reclamation and Improvement, Publ. 29, Wageningen, 377 pp.

LLOYD, J. W.; HEATHCOTE, J. A., 1985. *Natural Inorganic Hydrochemistry in Relation to Groundwater*. Clarendon Press, Oxford, 296 pp.

LLOYD, J. W., 1965. The Hydrochemistry of Aquifers of North-Eastern Jordan. *Journ. Hydrol.,* Vol. 3, pp. 319–30.

NITG-TNO, 2000. Regional Geohydrological Information System (REGIS). *REGIS Demo CD*. NITG-TNO CD, Delft.

PIPER, A., 1944. A Graphic Procedure in the Geochemical Interpretation of Water Analysis. *Trans. Am. Geophys. Union,* Vol. 25, pp. 914–23.

RUNYON, R. P.; HARPER, A., 1976. *Fundamentals of Behavioural Statistics.* Addison-Wesley, Reading, MA.

SCHOELLER, H., 1955. *Géochimie des eaux souterraines, application aux eaux des gisements de pétrole*. Revue de L'Institut Français du Pétrole et An. de Combustibles Liquides, 213 pp.

STEVENS, J., 1986. *Applied Multivariate Statistics for the Social Sciences.* Erlbaum, Hillsdale, NJ.

STIFF, H. A., 1951. The Interpretation of Chemical Water Analysis by Means of Patterns. *J. Petr. Technology,* Vol. 3, pp. 15–7.

TNO, 1993. *REGIS – Regional Hydrogeological Information System.* Information Document, TNO Institute for Applied Geoscience, Delft, 19 pp.

UNESCO, 1963. *International Legend for Hydrogeological Maps.* Unesco, Paris.

UNESCO, 1983. *International Legend for Hydrogeological Maps. Revised Edition*, Unesco, Paris.

United Kingdom, 1977. *Hydrogeological Map of England and Wales, 1:625,000*. Institute of Geological Sciences, London.

USGS, 1991. *System Requirements Specification for the U.S. Geological Survey's National Water Information System II*. Open-file Report 91-525, United States Geological Survey (USGS), Denver, 622 pp.

WILCOX, L.V., 1955. *Classification and Use of Irrigation Xaters*. U.S. Department of Agriculture. Circular No. 969, 19 pp.

5 Remote sensing techniques for groundwater prospection

5.1 Introduction

5.1.1 General

The remote sensing (RS) techniques discussed in this chapter are based on the interaction of electromagnetic radiation and matter. This interaction takes place at the surface of the earth (the depth penetration of the traditional RS techniques is negligible).

The image of the surface can be interpreted using a variety of techniques. The visual interpretation of a photographic hardcopy has long been a standard technique in earth and vegetation sciences, but now that computers and appropriate software have become easily affordable, many techniques that used to be restricted to specialised data processing centres have an interpretation of an area should also involve extrapolation to the unseen deeper part become available even to earth scientists with limited computer facilities. Personal computers can now be used to process satellite images, scan aerial photographs, apply Geographical Information Systems (GIS), and run sophisticated statistical packages and mathematical models.

5.1.2 Techniques and approaches

The many aspects of hydrology and hydrogeology that may be mapped include geological, soil or geomorphological units, drainage pattern, slope, vegetation, water bodies, lineaments, geological structures, man-made structures, water points and villages. The speed and accuracy of mapping can be increased if aerial photographs and/or photographic prints of satellite images are used (Allum, 1966).

The availability, distribution and quality of groundwater as well as its natural flow pattern is largely determined by the characteristics of the subsoil and the climatic conditions. Since remote sensing techniques basically reflect properties of the surface, in order to better understand the hydrogeology of the area, an interpretation of an area should also involve extrapolation to the unseen underground. This will lead to a conceptual three-dimensional model of the groundwater system. No standard recipes can be given for the development of such a model. The interpreter has to integrate the features he can see on his image with all the other available information on the area, such as thematic maps (geological, vegetation), meteorological data, hydrological data. Fieldwork is imperative to check the validity of the interpretation ('ground truthing').

For simple groundwater exploration, aerial photographs and RS are used to map features likely to be high-yielding zones, and to identify favourable structures or deposits for groundwater accumulation.

The applications of remote sensing data in water resources studies (Hansmann et al., 1992; Ling Hengzhang, 1985; Nieuwenhuis and Thunnissen, 1990; Schultz and Barrett, 1989; Waters et al., 1990) can be divided into several approaches.

(i) Visual interpretation using computers with image processing

Modern image processing software facilitates interpretation. Using the traditional prints of an image a trade-off had to be made between the overall quality of the image and the detectibility of specific features. Today's high quality colour monitors enable the interpretation to be done from the screen. Several digital enhancement techniques are usually available in the image processing software, the analyst can use information from several bands, and a selected window can be enlarged; this improves image interpretation. All these techniques can be fine-tuned in order to enhance the detectibility of a specific feature, e.g. the drainage pattern, the lineaments, the vegetation cover and the lithology. Once one feature has been mapped, another specific process can be applied to another feature.

(ii) Modelling the interaction of EM radiation with matter

In water studies modern image processing software mainly focus on the energy balance at the ground surface, the correlation between spectral response and vegetation characteristics and meteorological applications (Baily, 1990; Belward, 1991; Lagouarde and Brunet, 1993).

The actual evapotranspiration and soil water content plays a very important role in the energy balance at the ground surface. There are theoretical models that attempt to relate the thermal infrared radiation with the soil moisture content and the rate of evaporation (Mauser, 1990). For example, the so called 'cold cloud temperature' registered by meteorological satellites can be correlated to rainfall. The spectral response of vegetation can be modelled and used to estimate vegetation density, vegetation type and moisture conditions.

(iii) Classification, automatic interpretation and feature extraction

Much research is being done on the computerisation the interpretation of satellite images. Geologists want geological maps, soil scientists pedological maps and hydrologists hydrological maps – all readily made by a computer. The processing is based on the fact that different thematic units may have a different spectral response. There are simple or sophisticated rules, usually statistically based, for deciding how to interpret the area. Often the algorithms must be fine-tuned for the theme of interest. Pattern recognition techniques may also be used to decide how to classify an area.

(iv) Integration of RS, GIS and modelling techniques

Modern RS/GIS software enables digital RS data to be merged with other data.

Digitised maps, e.g. road networks, or geological maps, can be combined with the images, making the interpretation easier. An existing thematic map can be visualised together with the information from a satellite image.

In most studies the conceptual models essential as the backbone of the investigations, need to be quantified. It is not sufficient to know where water is stored; amounts, quality and fluxes also need to be known. Here again the techniques combining RS, GIS and simple models can be used to estimate fluxes. Mapping units can be based on geomorphology, geology, soils and vegetation. Some of the components of the hydrological cycle, such as actual evapotranspiration and groundwater recharge potential, may be estimated for different mapping units. Irrigated area, and the associated consumption of surface water and groundwater can be estimated using RS techniques. Better interpolation of rainfall data is possible if the data are combined with vegetation indices and digital terrain models.

Hydrological models may be directly programmed in the GIS, or be linked to the GIS, facilitating the preparation of input files for the model and serving as a post processor to display the results of the models calculations done by the model.

Flow systems can be schematised as cross-sections. Simple cross-section flow models can be used to explain the observations, or force the investigator to review his model when the results do not tally with the observations.

Airborne geophysical information may be displayed together with RS data, or a satellite colour image may be wrapped around a Digital Terrain Model (DTM) in an perspective view.

It must be stressed that digital techniques offer many advantages but cannot replace a sound knowledge of geomorphology, geology, hydrology, and soil and vegetation sciences.

On the following pages an introduction in remote sensing techniques and a selection of case studies will be presented to demonstrate the different aspects of the use of RS, aerial photography and GIS in hydrogeology.

5.2 Principles of remote sensing and GIS

5.2.1 Remote sensing

Remote sensing, in general terms is defined as collecting and interpreting information about a target without physical contact with the object (Sabins, 1986). The observations are usually made from aircraft or satellites. The definition would include the measurement of gravity, electrical and magnetic fields, radioactivity and electromagnetic radiation, but these techniques are considered part of airborne geophysics (see chapter 6).

The term 'remote sensing' is commonly restricted to methods that employ electromagnetic energy (light, heat, radar waves) as a means of measuring the characteristics of the earth's surface. The most commonly used RS products are aerial photographs and satellite images.

5.2.2 Physical fundamentals

(i) *Electromagnetic radiation*

Electromagnetic radiation is a wave phenomenon propagating with the speed of light, which is $3 \cdot 10^8$ m/sec.

For waves the velocity (c), wavelength (l) and frequency(f) are related by the formula $c = l \cdot f$.

The different classes of EM radiation, such as X rays, UV, visible light, infrared, radar and radiowaves are based on the wavelength or frequency. Figure 5.1 gives an indication of the nomenclature of different wavelength ranges.

(ii) *Interactions with matter*

Figure 5.2 schematically depicts the interaction mechanisms between electromagnetic radiation and matter. When the radiation encounters matter it is called the incident radiation. The following interactions can take place: reflection, scattering, transmission, absorbtion and emission.

The interaction between EM radiation and matter is usually a combination of the basic

Figure 5.1 Electromagnetic spectrum

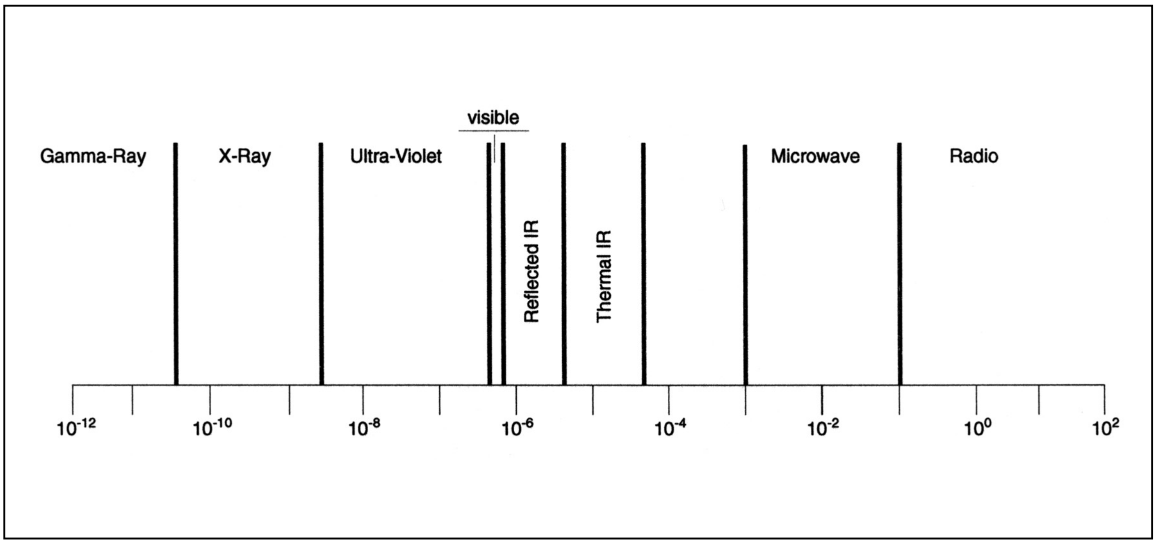
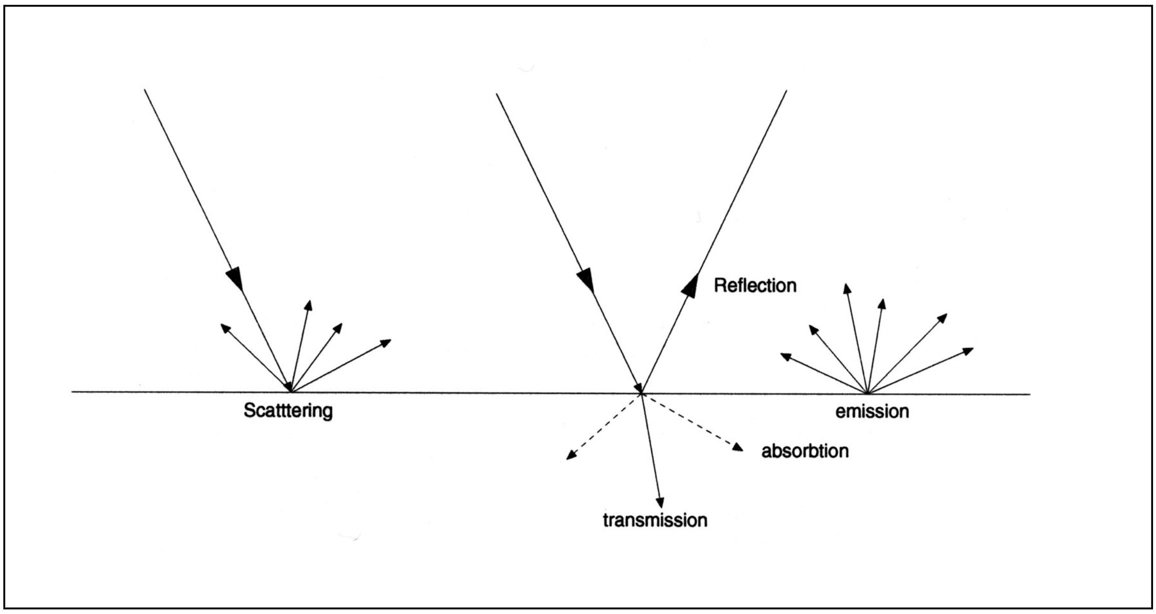


Figure 5.2 Interaction between EM radiation and matter

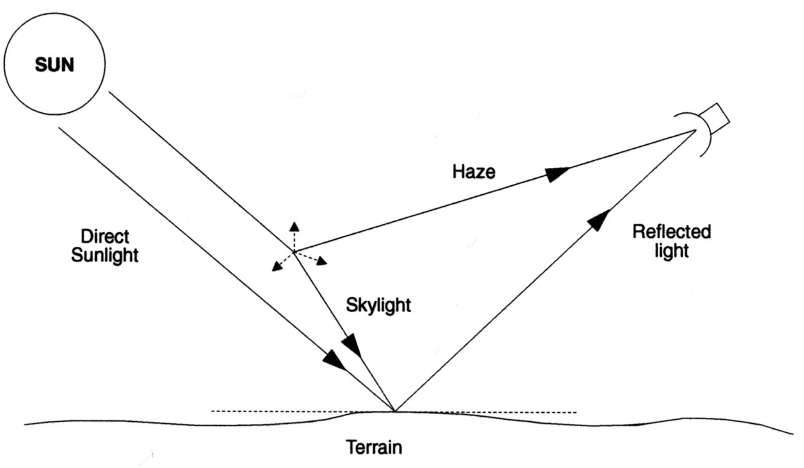


interaction phenomena described above. What is important is that all interactions vary as a function of the wavelength. Some of these interactions are recorded by a remote sensing system and the characteristics of the matter may be interpreted. By far the most important properties recorded are the reflection characteristics.

5.2.3 Aerospace imaging systems

The radiation returning from the surface of the earth is recorded by a photographic or scanning system. The two types of imaging systems are described below. A schematic representation of the path of the radiation through the atmosphere is presented in Figure 5.3.

Figure 5.3 Path and interactions between sun and sensor



(i) Photographic systems

Black and white films record the average intensity of the reflected light, colour films record the intensity of the three primary colours similar to the human eye, whereas infrared films are sensitive to near-IR radiation. Aerial photographs are usually taken from an aircraft though many photographs are have been taken during manned space missions. The aerial photographs taken in such a way that there is approximately 60% overlap between adjacent photos. Examination of the shared part of two photographs (a so-called stereo pair) through a stereoscope yields a three-dimensional impression of the landscape. These stereo pairs are used for a wide range of applications, such as the production of topographical maps and the mapping of soil, vegetation, land use, forestry, geology, geomorphology and hydrology features. (Verstappen, 1977; Meijerink and Van Wijngaarden, 1997; Sabins, 1986; Von Bandat, 1962).

(ii) Scanning systems

Scanning systems are more complicated than the lens-based photographic system. They divide an image into an array of millions of small square areas, called pixels, (picture element). These pixels are scanned and the incoming radiation for one or more wavelength intervals is recorded for each individual pixel. The wavelength intervals are called bands. These systems can be mounted in an aircraft, but most such data is collected by satellites.

Three types of resolution are distinguished. The spatial resolution is the size of the pixels, e.g. 30 x 30 metres for the Landsat Thematic Mapper, (Landsat TM). The spectral resolution is determined by the number of bands. The temporal resolution is determined by the frequency of the satellite overpasses. The recorded radiation for every pixel and every band is recorded as an 8 bit digital number (1 byte), which is equivalent to a number between 1 and 256.

5.2.4 Frequency domains

The imaging system may record one or more frequency or wavelength intervals.

A black and white film will record all visible light. Colour film will record the three primary colours. Sensors are designed to detect a specific wavelength or frequency interval. A major division in domains is made between reflected and emitted radiation.

(i) Reflected radiation

Visible and near infra red radiation is reflected, absorbed and scattered by the earth's surface. The reflected light is recorded by the sensor or photographic film. The differences on the image are mainly due to the difference in reflectivity of the Earth's surface.

(ii) Emitted radiation

Matter emits electromagnetic radiation. The intensity and wavelength are functions of temperature and properties of the radiating material. The wavelength of the radiation increases with temperature. At high temperature the wavelength enters the domain of visible light (e.g. on heating, a piece of iron changes in colour from deep red to yellow).

Thermal IR images may yield valuable hydrogeological information. In many cases the discriminating power of thermal IR with respect to lithology is higher than that of visible or near-infrared bands. Seepage and springs discharging into rivers, lakes or the sea may be detected, due to the temperature differences. Fracture patterns may show up very distinctively, as will be shown in one of the case studies. Thermal images can be used to estimate actual evapotranspiration and its spatial distribution.

Nevertheless, thermal IR imagery is rarely used, because thermal IR images are most useful if they are taken during night time, when solar heating is absent. Therefore, IR surveys have to be flown at night and cannot be combined with, for example, a survey of normal aerial photographs. Several satellites have thermal IR sensors. At present the Landsat TM with a pixel size of 120 x 120 metres has the highest resolution. This resolution is not enough for detailed investigations. Moreover, most of the TM images are acquired in the daytime, as during the night only one of the seven bands is recorded.

5.2.5 Imaging satellites

Figure 5.4 shows the characteristics of common satellites.

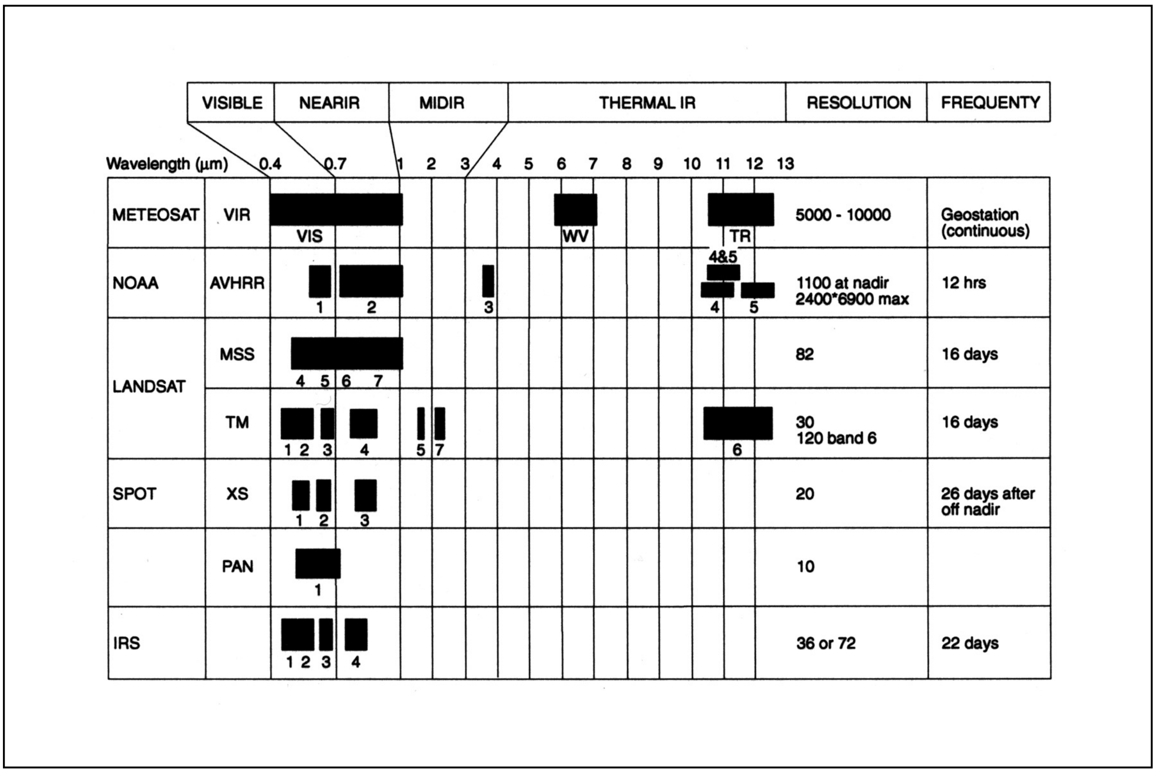
A special feature of the French SPOT Satellite is its capability to look sideways (off-nadir). This allows of an image of the same area to be produced from different orbital passes. This property makes it possible to produce satellite stereo images that can be studied with a normal stereoscope, as well as the high frequency imaging of an area. As the product has to be ordered specifically, it is rather expensive.

5.2.6 Digital image processing

The raw data sent by the satellites needs processing to transform it into visual information, either as a print or as an image displayed on the screen of a computer. The processing can be done by the seller of the data, a specialised firm, or directly by the user on a PC or a work-station.

The following data processing categories can be grouped:

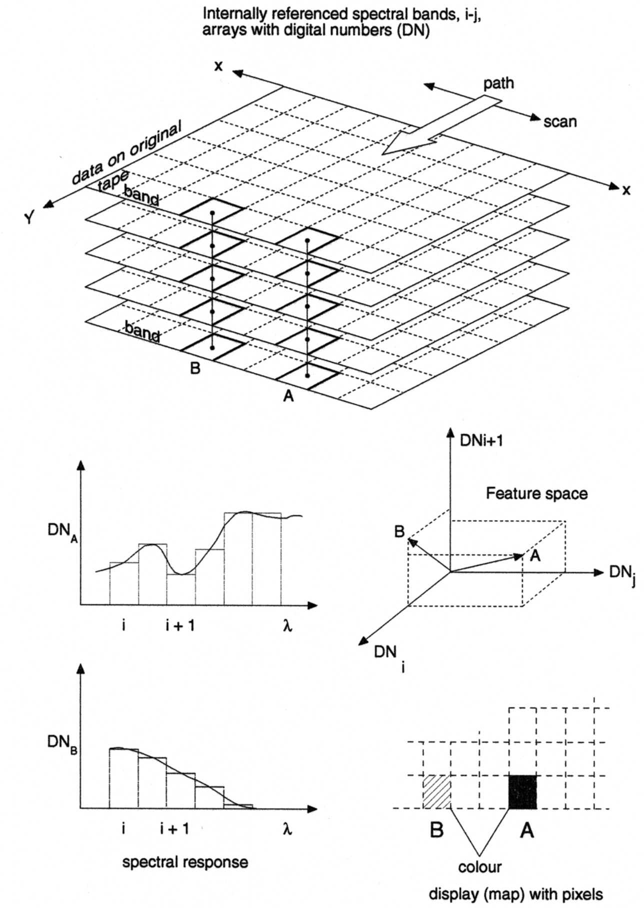
Figure 5.4 Satellite characteristics



- Corrections:
  - correction for technical distortions;
  - correction for atmospheric distortions;
  - correction for geometrical distortions;
  - image enhancements;
  - contrast enhancements;
  - filtering (edge and lineament enhancements);
  - colour coding;
  - transformations:
    - principal component images;
    - ratio images.
- Classification:
  decision rules are used to automatically classify (interpret) an image in different thematic classes.These classes may be type of soil, type of geology, landuse, vegetation, etc.

The image processing is schematised in Figure 5.5. The figure depicts four data arrays representing four spectral bands. Every element of the arrays represents a pixel and is stored as a digital number (DN). The higher the number, the higher the energy received (reflection) for that wavelength. The two XYgraphs represent the spectral reflection curves or spectral signatures for two pixels, A and B. They represent the reflectivity of a pixel as a function of the wavelength (λ). This curve is representative of different land and vegetation types. A combination of three bands can be represented by a three-dimensional vector. Clusters of points in n-dimensional space representing n bands of a set of pixels are called the feature space. A band can be represented as a grey-tone image by assigning black to the lowest digital number and white to the highest; the intermediate values are represented by grey shades. Three bands can be

Figure 5.5 From scanning to display



represented by a colour image, by assigning three bands to the intensity of the three primary colours: red, green and blue. For further reading on digital image processing and colour theory, see one of the specialised books (e.g. Lillesand and Kiefer, 1987; Mather, 1987; Mulder, 1982 and 1988; Niblack, 1986; Sabins, 1986; Wilkinson, 1991).

5.3 Geographical information systems (GIS)

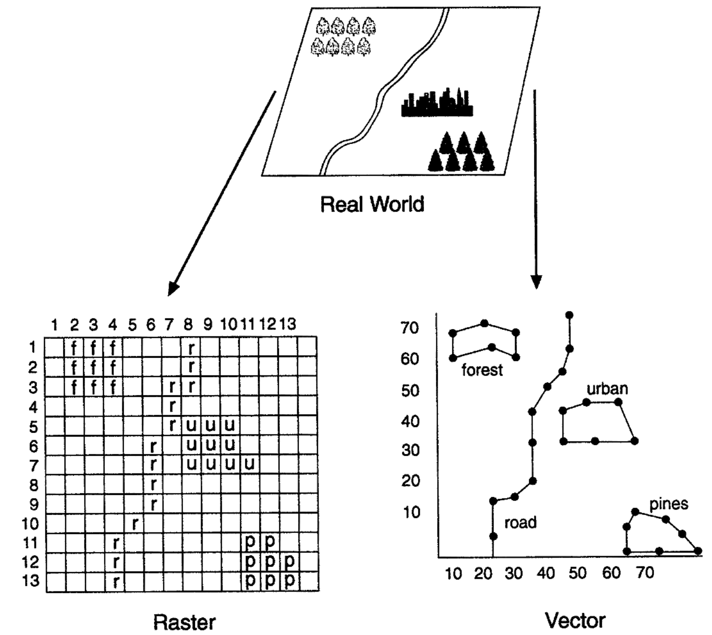
GIS has been a fast developing technology in spatial information processing. An overlap exists with pure image processing and many software packages offer both image processing and GIS capabilities. GIS can enter, store, retrieve, process and display spatial information in the form of maps or images (satellite, aerial photographs) including a database which is linked to the mapping units (Meijerink et al., 1987). This database is the equivalent of the legend in traditional maps.

On a map three elementary entities can be distinguished: points, lines or segments and surfaces or polygons. Points may represent villages, wells, met stations, lines may represent roads, rivers, faults, whereas polygons may represent countries, districts, soil classes, geological formations, or any other mappable units. Information on the points, lines, polygons are called attributes and are stored in a database. One record in the database contains the information on a geographical entity (point, line, polygon). The attributes stored in the database can be the number of lanes, quality, type, etc. of a road, the discharge, water quality, etc. of a river, the names, population, number of schools, etc. in towns and villages, the population of a district, etc., soil data like permeability, thickness, salinity, etc.

The two main data structures to store spatial information are vector and raster data structures.

- In a *vector-based* system, points are represented by a coordinate pair (X,Y), a line is represented as a series of coordinate pairs approximately defining the line and a polygon is represented by one or more lines surrounding an area (Figure 5.6).
- In a *raster-based* system, the area is overlain by a raster. A point is represented by one cell, a line by a sequence of adjacent cells and a polygon by a cluster of cells filling the polygon.

Figure 5.6 Raster and vector data structures



Data retrieval from non-spatial databases is only possible if selection criteria are applied (conditions) to the database itself, e.g. 'retrieve all wells with a yield of more than 10 m^3/s'. A GIS can also deal with geographical queries such as: 'select all wells which are at less than 3 km from a village', or a combination with a logical condition such as: 'retrieve all wells closer than 3 km from a village and yielding of more than 10 m^3/s'.

Maps may be combined or modified using GIS. For example a detailed geological map has an attribute database with the following three fields:

- Era: Palaeozoic, Mesozoic, Tertiary;
- System: Devonian, Carboniferous, Permian, Jurassic, Cretaceous, Palaeocene, Miocene;
- Lithology: granite_1, granite_2, granite_3, sandstone_1, sandstone_2, limestone.

One can combine all polygons from the same Era and the result will be a simple geological map with classes Palaeozoic, Mesozoic and Tertiary. Similarly, one can regroup the three granites and two sandstones producing a simplified lithological map.

Another process a GIS can handle is the overlaying or combining of two maps. A simple erosion index consists of a combination of slope, rainfall and soiltype. These parameters can be mapped on three separate maps. A combination gives a new map of the relative erosion hazard.

Raster and vector data structures each have their advantages and disadvantages; which approach is more suitable depends on the task to be done. In general, the vector approach is more suitable for large database-oriented GIS projects, whereas the raster-based systems are very suitable for geographical modelling and analysis. Some GIS packages have both approaches integrated, and can convert from vector to raster and vice versa. The raster data structure used in a GIS is also used to store satellite images. Many models (groundwater, surface water, erosion, crop yield) use a raster as their main data structure, and Digital Elevation Models (DTM) can also be represented in a raster data structure. Each cell of the raster has as attribute the average elevation of that particular cell. Therefore the linking and combination of technologies is straightforward. A cell of a raster based GIS representing a certain soil, may also represent a pixel with a certain DN from a satellite image, or represent a cell in a groundwater model for which the phreatic level can be calculated. The only condition is that the raster should represent the same area on the ground and the raster dimensions and cell sizes must be equal in all cases. All spatial data, can be integrated in a GIS be they thematic map units, remotely sensed images, administrative boundaries, topography, demographical data, towns, roads, wells, rivers, interpolated data from rainfall or groundwater levels, the simulated water levels, or river discharges.

The GIS has the capability of processing the information, which may reveal certain relationships, and ofvisualising different types of information simultaneously. So we can have satellite images with a road map superimposed on a three-dimensional representation of the landscape based on a DTM with a satellite colour composite image draped over it, or continuous geophysical data like gravity or magnetic field combined with satellite images. For further reading on GIS in general see Burrough and McDonnell (1998) or Aronoff (1989). For application in hydrology see Meijerink et al. (1994).

5.4 Required resources

Preprocessing yields two main products:

a) The digital data on computer compatible storage medium like tape or CD-ROM.
b) A negative or a photographic colour print.

(i) Aerial photographs

Existing aerial photographs can often be obtained for the cost of printing (typically a few US$). However, even for a relatively small area the complete coverage may consist of hundreds of photographs. This is also due to the large overlap.

(ii) Satellite images

Satellite images are rather costly. In 1995 commercial prices ranged between US$4,000 and US$4,500 for a full scene. It is also possible to order subscenes, ranging in price from a few hundred to several thousands of dollars, depending on the size. However, many images dating from before 1985 can be acquired for prices ranging from US$250 to US$500. It is likely that prices will fall when the raw unprocessed data is available at low cost and commercial companies can add value by processing. Information on products can also be obtained via Internet.

The requirements for hardware are modest. Most software is available for Windows, requiring a computer with a 486 or higher processor. Until recently, special equipment was needed to read the computer compatible tapes containing the data. Since data can now be ordered on a CD-ROM, only a CD-ROM player is required to transfer the image to the hard disk of the computer.

Colour inkjet printers can produce images of acceptable quality; however, specialised and expensive equipment is still needed for the high-quality photographic prints. With Microsoft Windows evolving as a dominant operating system, the integration of maps and images in a document will become routine.

5.5 References and additional reading

ALLUM, J. A. E., 1966. *Photogeology and Regional Mapping.* Pergamon Press, Elmsford, N.Y.

ARONOFF, S., 1990. *Geographic Information Systems: A Management Perspective.* WLD Publications, Ottawa.

BAILEY, J. O., 1990. The Potential Value of Remotely Sensed Data in the Assessment of Evapotranspiration and Evaporation. *Remote Sensing Reviews,* Vol. 4, pp. 249–377.

BELWARD A. S., 1991. Spectral Characteristics of Vegetation, Soil and Water in the Visible, Near-Infrared and Middle-Infrared Wavelengths. In: Belward and Valenzuela (eds.), *Remote Sensing Applications and GIS for Resources Management in Developing Countries (Eurocourses).* Kluwer Acad. Publ., Dordrecht, pp. 31–54.

BURROUGH, P. A.; MCDONNELL, R. A., 1998. *Principles of Geographical Information Systems*. Oxford University Press, Oxford.

COLWELL R. N., (ed.), 1983. *Manual of Remote Sensing*. Amer. Soc. of Photogrammetry, Virginia.

DURRET, H. J. (ed.), 1987. *Color and the Computer*. Boston Academic Press Inc., Boston, 299 pp.

ELACHI C., 1987. *Introduction to the Physics and Techniques of Remote Sensing*. John Wiley and Sons, New York.

HANSMANN, B. C.; MEIJERINK, A. M. J.; KODITUWAKKU, K. A. W., 1992. An Inductive Approach for Groundwater Exploration. *ITC Journal* 1992–1993, pp. 269–76.

HENGZHANG, L., 1985. Applications of Remote Sensing Technology to Water Resources Exploitation in Mountainous Areas in Southwest China. In: *Remote Sensing Applications to Highland Development. Proc. Regional Seminar, UNDP/ESCAP, 7–11 Nov. 1985, Kathmandu, Nepal,* pp. 80–5.

JACKSON R. D., 1983. Spectral Indices in N-Space. *Remote Sensing of the Environment*, No. 25, pp. 295–309.

LAGOUARDE J. P., 1991. Use of the NOAA AVHRR Data Combined with an Agrometeorological Model of Evaporation Mapping. *International Journal of Remote Sensing,* Vol. 12, pp. 1853–64.

LAGOUARDE, J. P.; BRUNET, Y., 1993. A Simple Model for Estimating the Daily Upward Longwave Surface Radiation Flux from NOAA-AVHRR Data. *International Journal of Remote Sensing,* No. 14, pp. 907–25.

LILLESAND, T. M.; KIEFER, R. W., 1987. *Remote Sensing and Image Interpretation.* John Wiley and Sons, New York.

MAUSER, W., 1990. Modelling the Spatial Variability of Soil Moisture and Evapotranspiration with Remote Sensing. In: *Proceedings, International Symposium Remote Sensing and Water resources (IAH-BCRS), Enschede, The Netherlands,* pp. 249–260.

MEIJERINK A. M. J. ET AL., 1994. *Introduction to the Use of Geographical Information Systems for Practical Hydrology*. ITC-UNESCO, Enschede.

MEIJERINK, A. M. J.; VALENZUELA, C. R.; STEWARD, A., 1987. *The Integrated Land and Watershed Management Information System (ILWIS)*. ITC Publication No. 7, 115 pp.

MULDER, N. J., 1982. Methodology of Colour Coding MSS and Other Data. *Archives Int. Soc. Photogrammetry and Remote Sensing,* Helsinki Symposium, Commission III.

MULDER, N. J., 1988. Digital Image Processing, Computer Aided Classification and Mapping. In: A. W. Küchler and I. S. Zonneveld (eds.), *Vegetation Mapping.* Kluwer Ac. Publ., Dordrecht.

NIBLACK, W., 1986. *An Introduction to Digital Image Processing.* Prentice Hall International, 215 pp.

NIEUWENHUIS, G. J. A.; THUNNISSEN, H. A. M., 1990. Application of Remote Sensing in Water Management. In: *Proceedings, International Symposium Remote Sensing and Water resources, (IAH-BCRS), Enschede, The Netherlands*, pp. 229–48.

SABINS, F. F., 1986. *Remote Sensing: Principles and Interpretation*. W. H. Freeman & Co, New York.

SCHULTZ, G. A.; BARRETT, E.C., 1989. *Advances on Remote Sensing for Hydrology and Water Resources Management.* UNESCO, Techn. Doc. IHP, Project 5.1, Paris.

VERSTAPPEN, H. TH., 1977. *Remote Sensing in Geomorphology*. Elsevier Scient. Publ. Co.

VON BANDAT, H. F., 1962. *Aerogeology*. Gulf Publishing Compagny, Houston, Texas.

WATERS, P.; GREENBAUM, D.; SMART, P. L.; OSMASTON, H., 1990. Applications of Remote Sensing to Groundwater Hydrology. *Remote Sensing Reviews,* Vol. 4, pp. 223–64.

WILKINSON, G. G., 1991. The Processing and Interpretation of Remotely-Sensed Satellite Imagery, A Current View. In: Belward and Valenzuela (eds.), R*emote Sensing Applications and GIS for Resources Management in Developing Countries (Eurocourses).* Kluwer Acad. Publ., Dordrecht, pp. 97–110.

6 Geophysical techniques in groundwater investigations

6.1 Introduction

For many years geophysical exploration techniques have proved to be efficient tools in groundwater exploration, not only in the direct detection of the presence of water, but also in the estimation of aquifer size and properties and groundwater quality. As objectives in groundwater exploration may be named: detection of the position and extent and volume of potential aquifers, location of faults and fracture zones below undifferentiated surface covers, detection of fresh/saline water interfaces, estimation of the depth of the water table, detection and tracing of groundwater pollution, detection of aquitards and connections between different aquifers. Through correlation with boreholes and pump tests an estimation can be made of the variation of porosity, permeability and transmissivity through an aquifer.

This chapter presents an overview of methods which are at present applied to groundwater exploration and it does not pretend to be complete. It should be noted that there has been enormous progress in the development of applications since the early 1980s, so geophysical methods other than the 'traditional' ones have been included.

This chapter discusses the various methods in a sequence of large-scale surveys (regional scale, airborne methods) to small-scale ones (local scale, ground-based methods). The rationale behind this is that the surveying process aims at investigating large areas with fast, efficient and relatively cheap methods, to select those parts which seem attractive or even favourable for further, more detailed surveys. These will give more information about the target, but are at the same time more costly. By surveying in a sequence of large scale to small scale an efficient and cost effective result may be obtained. It should be realised here that the airborne and waterborne techniques are actually variations on the original ground-based techniques.

6.2 Magnetic methods

From the point of view of mapping and exploration using the magnetic method, the whole earth may first be considered in three parts: core, mantle and crust.

6.2.1 The source of the earth's magnetic field

(i) Data collection

The magnetic field of the earth originates in the molten part of the earth's core. This 'core field' varies smoothly and predictably over the surface of the earth and is defined conventionally by the *International Geomagnetic Reference Field* (IGRF) which changes at any one locality slowly and progressively with time. The field at any given point is usually defined in terms of its *total field strength* (T), its *inclination* (I) and the deviation between true and magnetic north or *declination* (D). In geophysical mapping, the scalar magnitude of the geomagnetic field is (with few exceptions) the *only* element that has been measured; this frees the geophysicist from the need to relate his observation to any reference direction and so makes rapid surveying, even from a moving aircraft, relatively straightforward using suitable electronic magnetometers.

The earth's mantle is too hot to be magnetic since the magnetic properties of rocks generally disappear when they are heated above the Curie point temperature. This falls in the range 550–600°C so, with typical geothermal gradients, this temperature is reached at depths towards the base of the crust.

(ii) The crust, the source of magnetic anomalies

Most of the rocks within the earth's shallow crust are cooler than the Curie point temperature and may display magnetic properties, either *remanent* or *induced*. *Remanent magnetisation* is a permanent setting of the magnetic fabric of a rock that may persist for hundreds of millions of years and may have been acquired when the rock was first formed or when it was re-heated during metamorphism. *Induced magnetisation*, on the other hand, is a temporary effect, acquired whenever a susceptible material is located in a magnetic field such as that of the earth's core. It is very difficult to say whether the magnetism present in rocks *in situ* is induced or remanent, except in the case of outcropping rocks that can be sampled and subjected to laboratory measurements.

Both the induced and the remanent magnetisation of rock are carried only by the magnetic minerals present. Magnetic properties are addressed as magnetic susceptibility: the degree to which the mineral particles can be magnetised. Table 6.1 presents some susceptibility values of minerals, ores and rocks.

Table 6.1 Magnetic susceptibilities of minerals, ores, and rocks

Mineral/ore	*Susceptibility (S.I.)*	*Rock type*	*Susceptibility (S.I.)*
Pure magnetite	15	Sediments	0.0–0.001
Magnetite ore	0.07–14	Metamorphic rocks	0.0001–0.001
Ilmenite ore	0.3–4	Granites	0.0001–0.01
Pyrrhotite	0.001–0.1	Basic plutonics	0.0001–0.1
Hematite ore	0.0004–0.01	Volcanic rocks	0.00–0.1
Pyrite	0.0001–0.005	Banded iron formations	Very high remanent magnetisation

The combined effects of the induced and remanent magnetisation of rocks imposes local variations – *anomalies* – on the earth's main magnetic field. Departures of actual magnetic field values from those predicted by the IGRF may be mapped out in a survey using ground or airborne techniques to produce maps or images that reflect the distribution of magnetic properties of crustal rocks. Since most sedimentary rocks are virtually non-magnetic the anomalies mapped out over sedimentary basins – or over areas of surficial cover, dense vegetation or water cover – provide an important means of interpreting the depth and structure of underlying igneous and metamorphic rocks. This has led to the widespread application of aeromagnetic surveying to extend the mapping of the so-called magnetic basement into the large areas of the world where the igneous and metamorphic rocks do not outcrop.

(iii) Measuring the magnetic field and its anomalies; data collection

The magnetic field can be measured with sensors based on different principles; the oldest is the fluxgate sensor. Modern instruments for ground operation use a sensor based on the proton precession meter or the so-called (optically-pumped) caesium magnetometer. The latter is also used in airborne operations, allowing fast sampling of the magnetic field.

The strength of a magnetic field is usually expressed in Tesla. The scalar magnitude of the IGRF falls within the range 25,000 to 60,000 nT (*nanoTesla*) over almost all of the earth's surface. Anomalies due to igneous and metamorphic rocks measured at typical aircraft survey altitudes (50 to 100 m above terrain) seldom exceed one or two thousand nT and mostly are of the order of tens or hundreds of nT. Ground-based magnetometers record variations that are perhaps ten times greater. Magnetometers are capable of measuring magnetic variations with a sensitivity of at least 1 nT and often to small fractions of this quantity. Modern magnetometers measure the absolute value of the so-called *total magnetic field* (i.e. core-field *plus* anomalous component, added vectorially) to this accuracy in a measuring time of one second or less. So, magnetic anomalies with amplitudes from less than 1 nT to several thousand nT can be detected and mapped out conveniently and accurately in magnetic surveying, even though they often represent less than 1% of the total magnetic field at any given location (Reeves, 1992).

The survey procedure is complicated by short-term variations in the geomagnetic field (*temporal variations*) that have amplitudes of 50–100 nT on an approximately daily cycle. They can have much larger and unpredictable variations due to so called magnetic storms that have their origin in events on the sun. Temporal variations can be eliminated from field observations by recording them during the survey at a base station, or by measuring the magnetic gradient, using two short-spaced sensors simultaneously.

The general procedure for data collection is to make observations at regular intervals *(stations)* along a profile line. For *mapping* purposes, profile lines are surveyed at more or less regular intervals *(line spacing).* For mapping purposes the line spacing must be such that no significant anomalies will escape between adjacent lines. This is particularly important for airborne operations, where closer lines lead to more flight kilometres and consequently to higher survey costs. For most surveys, lines should not be separated by more than 4–5 times the vertical distance from source to magnetometer. For the detection of shallow sources this leads to a high line density in ground-based surveys. For airborne surveys a flight elevation of 100 m (above terrain) would allow a line spacing of 400–500 m, while at 50 m flight elevation a spacing of 200–250 m would be required. Increasing the flight height will reduce the survey cost, but the consequence of that is a reduced resolution of smaller and deeper sources. Ultimately, the cost of an airborne survey is determined by the desired resolution. The limit of the resolution is again set by how low an aircraft can be flown with safety. Good sampling of a magnetic anomaly requires at least 10, but preferably 20, observations over the length of the anomaly. An aircraft flying at a height of 50 m at a velocity of 70 m/sec and fitted with sensors which can sample 10 times per second, will sample at an interval of 7 m on the ground. For good progress (i.e. production) a ground survey would require an interval of 5 or 10 m, risking poor definition of anomalies over shallow sources, particularly in areas with high levels of 'geological noise' (laterites, magnetic boulders).

When large areas need to be surveyed, an airborne operation will generally prove to be most cost effective. For detailing of selective areas a 'follow-up' by a ground based survey can then be carried out, which reduces also the operational problems associated with less accessible terrain.

6.2.2 Data processing and interpretation

When the local IGRF value is subtracted from the local magnetometer measurement to leave the local anomaly, the value of the anomaly is a very close approximation of *the component of the anomalous field in the direction of the inducing field*. Since the distribution of magnetic properties in rocks can be quite complex, magnetic anomalies can be both positive and negative. Even a single very simple geometrical source (e.g. a buried magnetic cube) will produce a magnetic anomaly that has both positive and negative parts. Most of the negative part flanks the side of the body nearest to the magnetic pole of the hemisphere, i.e. on the north side for the northern hemisphere and on the south side for the southern hemisphere.

The magnetic anomalies over simple buried geometrical shapes can be predicted from the physical and geometrical parameters of the sources by *forward modelling* calculations. Taken one stage further, the model may be adjusted in shape and depth to produce an anomaly which closely fits an observed anomaly. This process is called *indirect interpretation* and can be further refined into a process of *inversion* whereby the physical and geometrical parameters of the body are adjusted to generate a response which closely fits the observed magnetic anomaly. This is particularly useful where an estimate of, for instance, the depth of a sedimentary layer or the thickness of weathering needs to be estimated.

(iii) Estimating depths of magnetic sources

By forward modelling it can easily be demonstrated that, as the depth of a given body increases, the shape of its anomaly changes. The most obvious change is the reduction in amplitude of the anomaly. Depending on the extent of the source body in depth and strike-length, the amplitude is found to decay by a power usually between one and three of the distance between source and magnetometer. Doubling the distance between source and sensor will cause the anomaly to reduce in amplitude by somewhere between one-half and one-eighth. However, the amplitude is *directly* proportional to the magnetisation of the rock which can have a range that varies over *several* orders of magnitude. It follows that the amplitude of an anomaly is of little use in estimating the depth of its source, since the magnetisation of the source body is usually unknown and even reasonable estimates could be in error by two or three orders of magnitude.

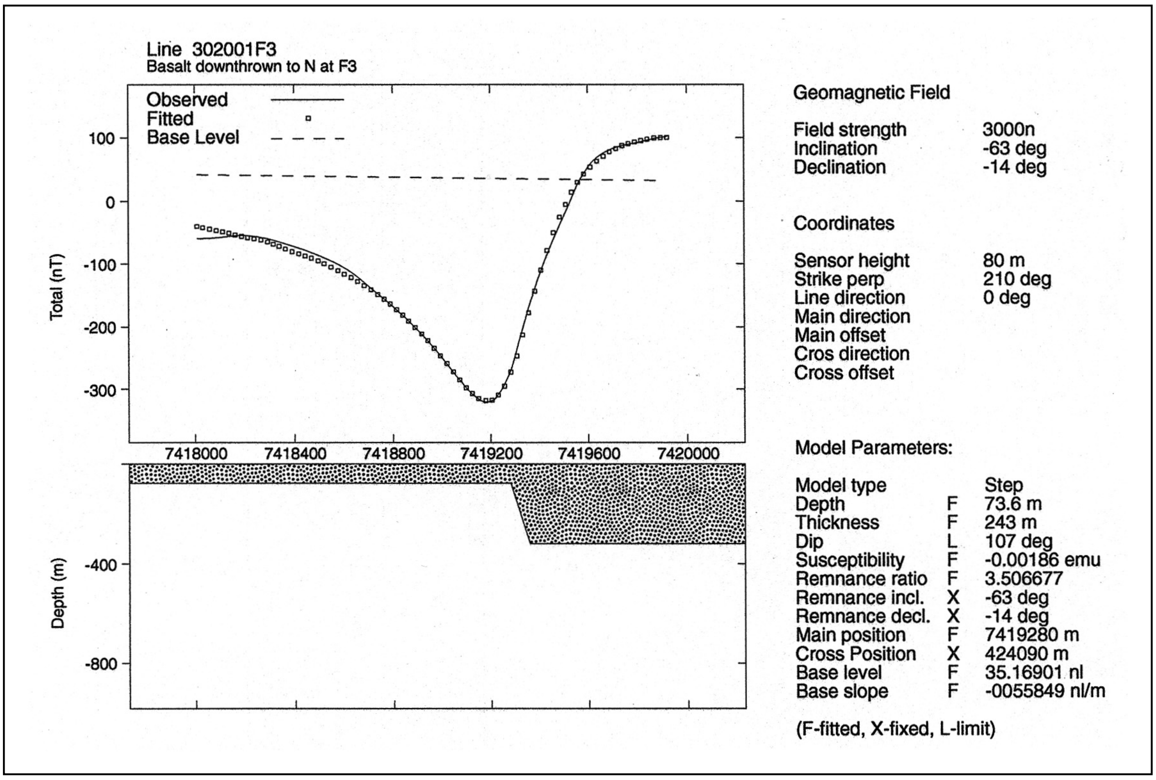
More helpful in estimating the depth of the source of an anomaly is the fact that the 'wavelength' of the anomaly increases approximately linearly with increasing depth of source (at least for compact sources that are much smaller than their depth), regardless of the magnetisation. In the days before computer-based interpretation became routine Several graphical methods to estimate the wavelength were used. Well defined anomalies may be interpreted by inversion to produce quite reliable estimates of the depth and geometry of the source body, given some reasonable geological constraints such as the likely shape of the source (dyke, fault, horizontal plate, down-thrown block, etc.), as shown in Figure 6.1.

(iv) Maps and images

After removal of temporal variations the survey results are presented as contour maps (Figure 6.2) or images (Figures 6.3).

Most land areas of the world have by now been covered by aeromagnetic survey (though if carried out many years ago these surveys are sometimes of inferior quality) and the groundwater explorer should be aware of the potential value of surveys that already exist. Where aeromagnetic data are available in digital format, display and enhancement using methods of image processing has led to many methods of data presentation that can be tailored to the needs of the user by seeking the clearest possible expression of the geological features of interest.

Figure 6.1 Quantitative interpretation of the magnetic anomaly over a normal fault in a basalt sheet at position D in Figure 6.4. The basalt is 300 m thick north of the fault and only 100 m thick to the south (Courtesy Dept. of Water Affairs, Government of Botswana)



Perhaps the most important development resulting from this has been as simple as making aeromagnetic images attractive and more intuitively interpretable for those who are not necessarily specialists in geophysics.

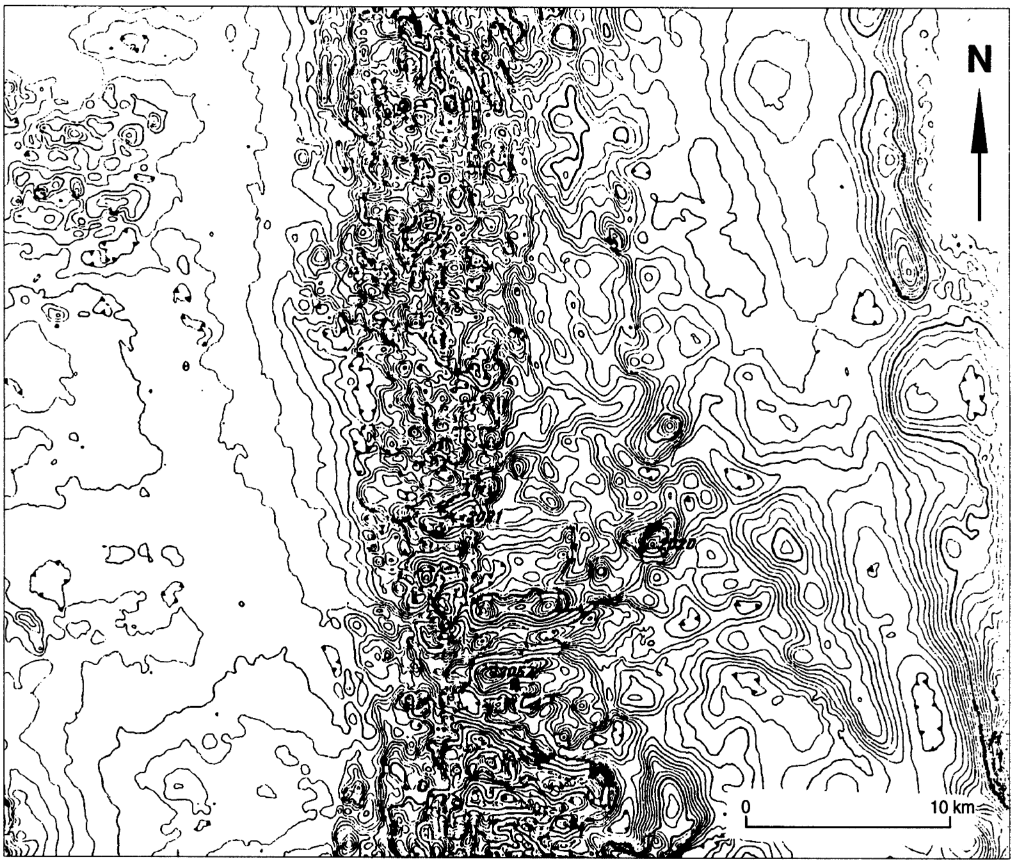
(v) Interpretation

Interpretation of aeromagnetic maps is usually both qualitative and quantitative. Probably the most widespread purpose is to correlate the mapped magnetic expression with known geological mapping and to interpolate and extrapolate the magnetic response to areas with obscured, unknown, geology. A full discussion of this process is beyond the scope of this volume. However, some attention must be drawn to some of the deductive processes that are particularly valuable in groundwater exploration.

Perhaps the most obvious application would be in typically hard rock areas. In such areas the search is concentrated on localised areas of enhanced or secondary porosity that may be of value. These could be areas with thick weathered bedrock or with buried channels. In view of the uncertainty (approx. 20% of the source to sensor distance) in estimating the depth of source rocks from magnetic data, the success of an airborne survey in this application would probably be limited. This is particularly so in areas where intrusive granites appear which have such a featureless magnetic signature that they can easily be confused with areas of sediment. Still, magnetic ground surveys are useful for locating weathered bedrock and buried channels.

A second, more promising, example is in the tracing of faults and fractures through such hard rock areas. Distribution of magnetic properties in rocks gives rise to a complex magnetic anomaly pattern or *signature* which may be associated with a particular assemblage of rocks.

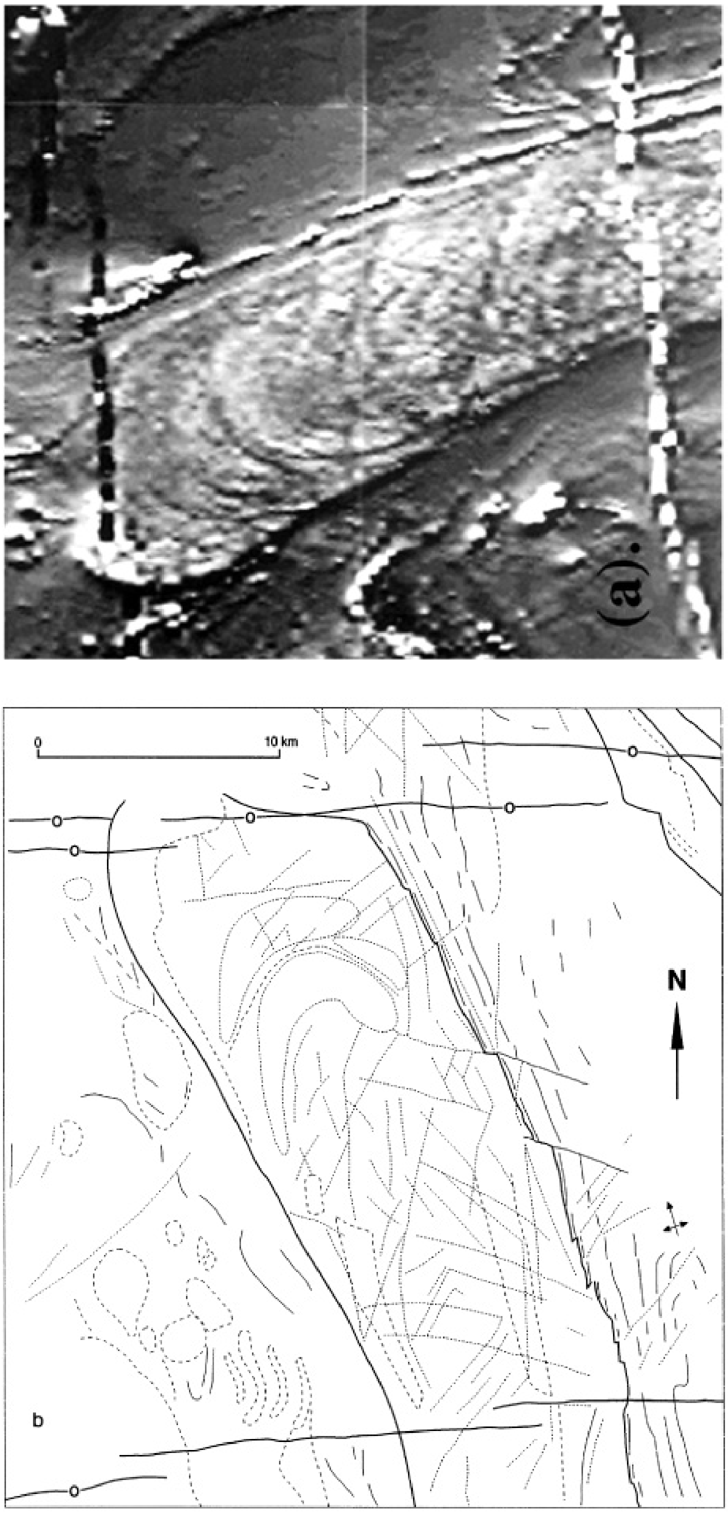
Figure 6.2 Contour map of magnetic anomalies for a survey flown at 100 m ground clearance, east-west lines at 400 m spacing and a contour interval of 2 nT. Relatively magnetic metasediments occupy the centre of the map, with a granite body to the east and sediments to the west (Courtesy of the Australian Geological Survey Organisation)



This pattern may differ from the pattern over an adjacent rock type of different lithology, in amplitude, in number of anomalies, in shape of the anomalies. etc. Criteria of this sort allow many hidden boundaries to be mapped out between often only limited areas of exposure suitable for conventional field mapping. Similarly, faults or joints can be revealed by the offset of the magnetic pattern, as in Figure 6.3. More subtle expressions of such faults or joints can often be recognised. In some cases, dykes are clearly seen to produce distinctive magnetic anomalies when magnetite-rich mafic magma has been injected along fault planes.

The aeromagnetic interpretation maps resulting from interpretations of features of all these sorts are often criss-crossed by interpreted dykes, faults and joints. Though not very interesting for the geological mapping of the area, they may be crucial in the siting of successful groundwater boreholes. A ground-breaking study by Astier and Paterson (1987) showed that, for an area in Burkina Faso, West Africa, the yields of water boreholes in hard rock terrain decreased markedly with distance from interpreted aeromagnetic fractures. Boreholes sited within 300 m of a lineament were far more likely to give substantial yields than those sited further away. The same study showed that, in the test area, far more lineaments were evident from aeromagnetic data than from satellite and air photo interpretation. It can be concluded that in many areas the

Figure 6.3 (a) Shaded relief aeromagnetic image of a granite in the Yilgarn Block of Western Australia with illumination from the east. (b) Interpretation of the fracture pattern revealed from the magnetic data and the outline of the granite. Proterozoic dykes striking east-west cut across the area near the north and the south margins of the figure (Courtesy of the Australian Geological Survey Organisation)



preferred locations for detailed ground follow-up studies for siting water boreholes would be near the linear features interpreted from aeromagnetic maps.

(iv) Digital elevation models from airborne surveys

A further potentially useful development is the use of GPS heights for the aircraft, coupled with the radar-altimeter distance between aircraft and ground surface, to give accurate ground heights along each flightline of a survey. This information can be converted into a digital elevation model of the survey area, with terrain blocks (pixels) of dimension about 50 m and an accuracy of better than 2 metres. Such information, particularly in areas that are poorly surveyed, can be most useful in groundwater studies, not only in providing a digital terrain model that may be important for modelling aquifer recharge, but also in determining reasonably accurate heights for the collar of boreholes.

6.2.3 A case-study of aeromagnetic surveying for aquifer exploration

As an example, a case study from Botswana is presented in outline. Most of eastern Botswana is underlain by Precambrian rocks which are relatively unpromising for large-scale water supply. However, Palaeozoic/Mesozoic rocks of the Karoo Supergroup overlie the Precambrian rocks in some areas and contain a massive porous sandstone which has proved to be a very productive aquifer, particularly where it is capped by the relatively impermeable basalt and which forms the topmost ('Stormberg') stage of the Karoo Supergroup. The extensive Kalahari sand-cover makes mapping the geology difficult, so outlining the areas of sub-outcropping Karoo rocks requires innovative approaches.

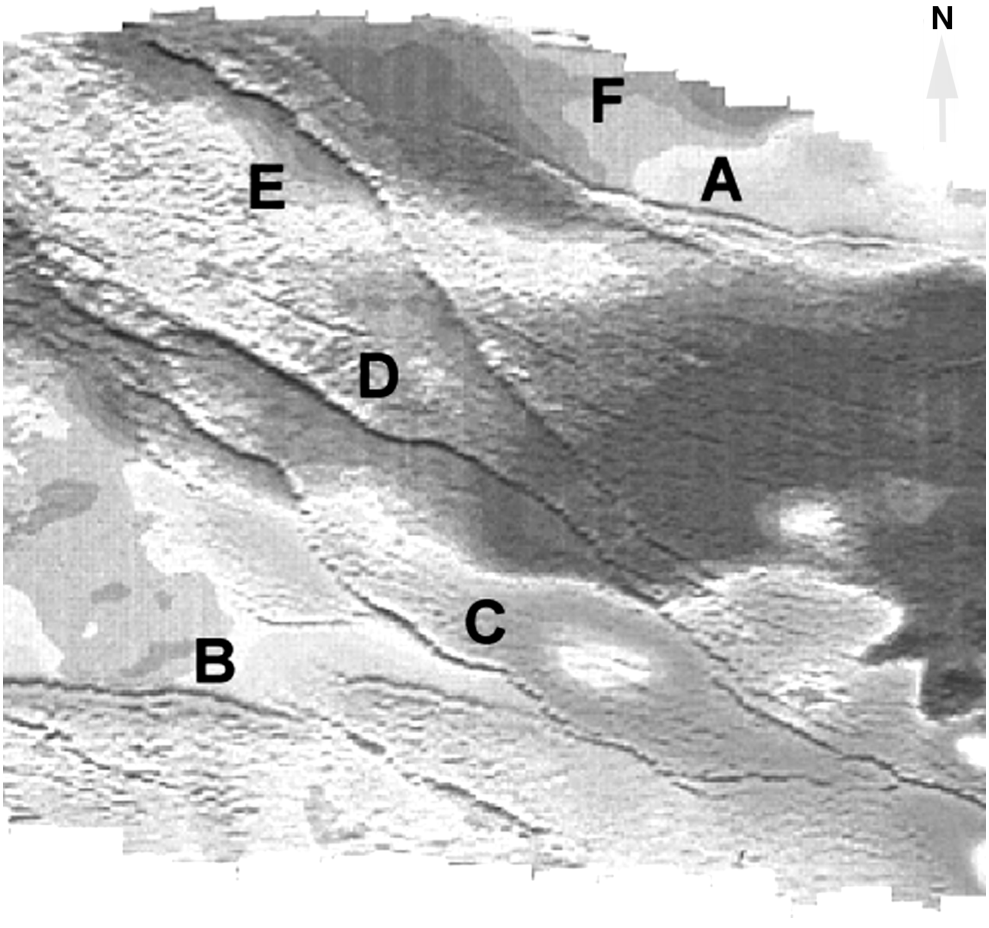
When data from a low-level magnetic survey are presented as an enhanced image (Figure 6.4) the ripple signature of areas underlain by basalt are distinctive and the faulted margins of areas of different sheet thickness are clearly displayed. Areas devoid of basalt show a comparatively smooth 'glossy' texture.

Extremely strong anomalies from highly magnetic rocks in the Precambrian sequence below the Karoo tend to swamp the rather more subtle basalt anomaly patterns in the east of the study area. Modelling of specific step-fault anomalies, as illustrated in Figure 6.1, enables the thickness of the basalt to be estimated and the areas where the sandstone aquifer is confined by an appropriate thickness of basalt can be mapped out in an interpretation map such as Figure 6.5. With support from suitable ground geophysical studies and indications from simultaneous interpretation of satellite imagery, plans can be made for drilling and testing the potential for the sandstone aquifer where it is confined (unshaded, Figure 6.5). Less promising areas, where the sandstone is unconfined – or where Karoo rocks are totally absent – can be avoided.

6.3 Electromagnetic (EM) methods

EM techniques are used to survey the electrical properties of the subsurface materials. One of the key feature of the EM techniques is that most applications do not require a galvanic contact, i.e. there is no need for electrodes making a physical contact with the ground. This allows work to be done over highly resistive top surface material (such as lateritic cuirasse, duricrust, caliche or permafrost), to survey from airborne platforms (helicopter or fixed wing aircraft) and normally it permits a rapid rate of survey progress because the time required to install/remove electrodes is eliminated. Usually EM and Resistivity techniques (see Section 6.4) are complementary. In specific cases EM techniques may be more appropriate than resistivity and vice versa. There is

Figure 6.4 Magnetic anomalies of the Palla Road area, eastern Botswana, processed to second vertical derivative and presented as a shaded-relief grey-scale image (white high and black low) with illumination from the south. Note the positive anomalies over the north edges of the basalt sheet (e.g. positions A and B) and negative anomalies over the south edges (position C) and where faulting thins the sheet to the south (as at D). The 'ripple' signature of the basalt (as at E) is in contrast to the areas such as at F, where basalt is absent (Courtesy of the Dept. of Water Affairs, Government of Botswana)

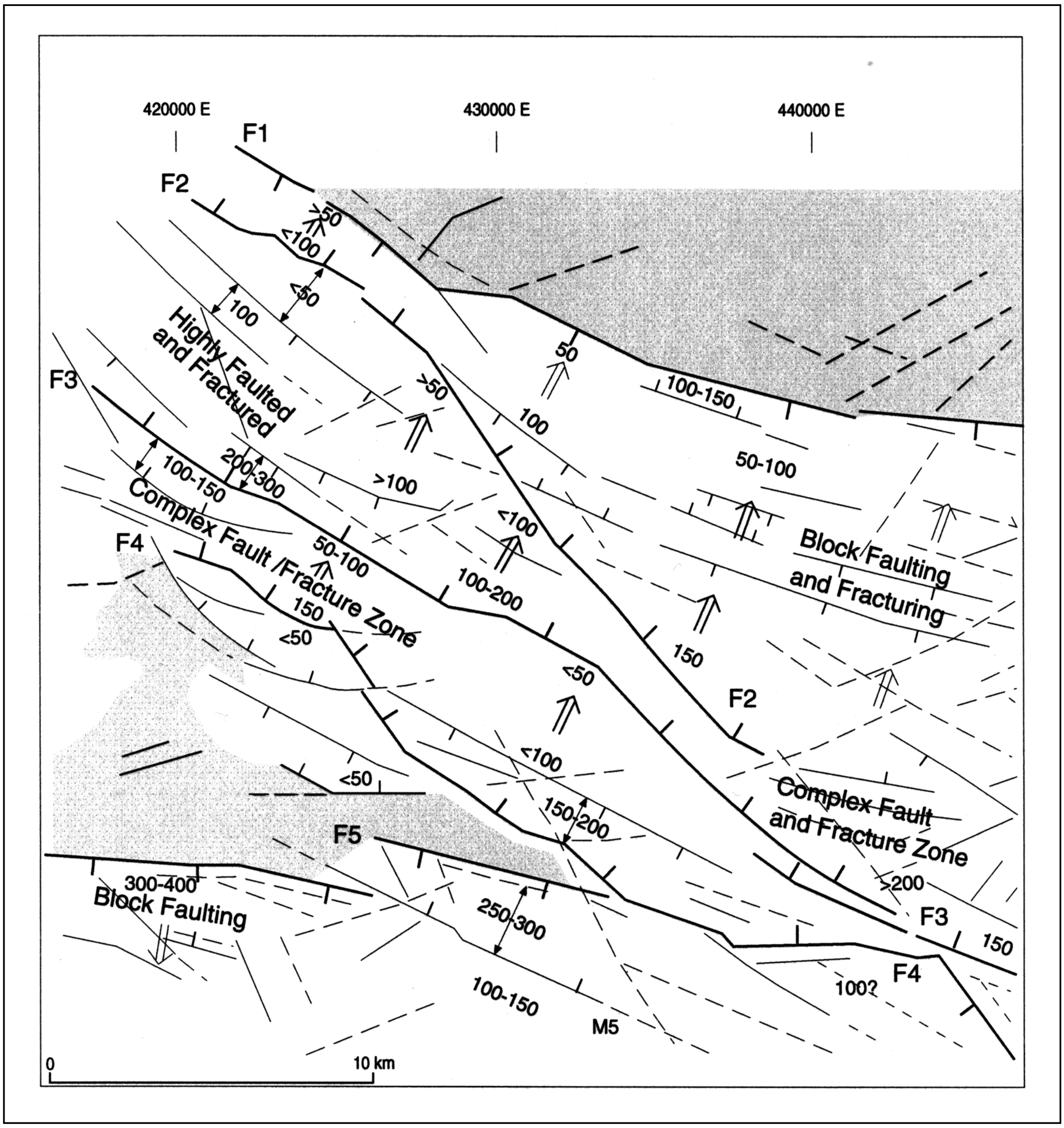


a wide variety of practical implementations for the EM techniques, but only a few commonly applied techniques will be discussed in this Section.

6.3.1 Principles and survey techniques

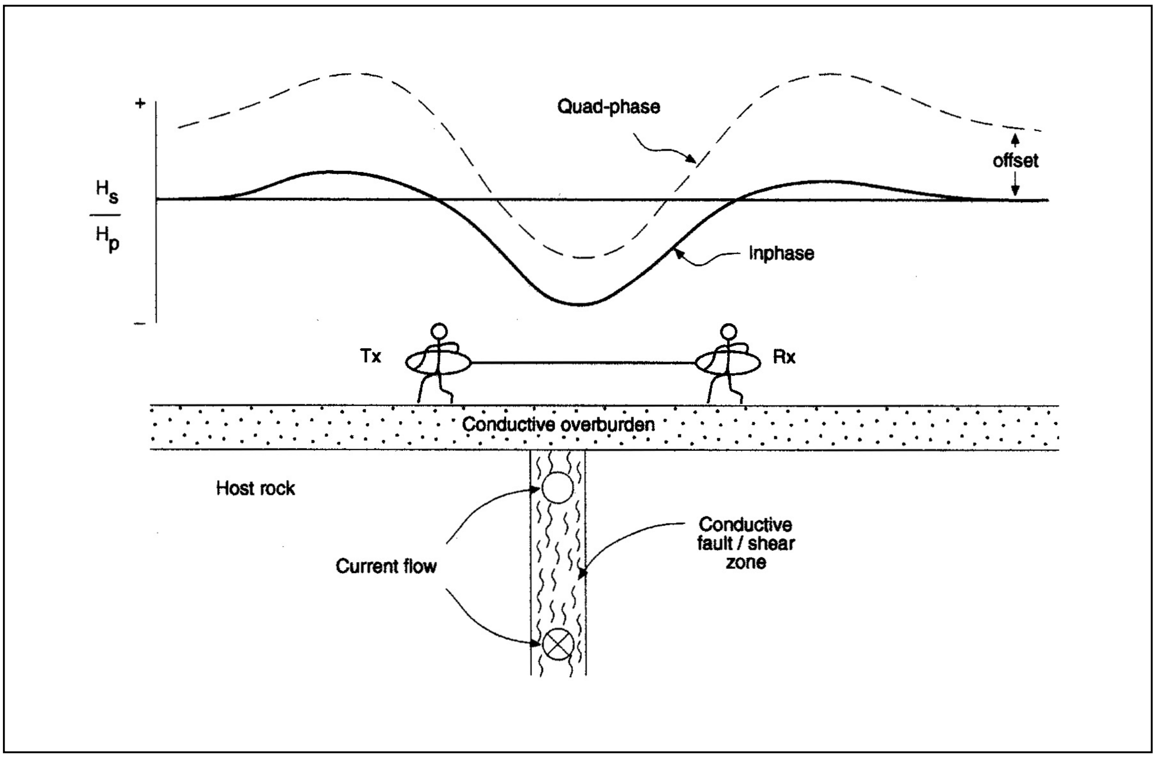
EM surveying is based on the principle that an alternating current running through a loop wire generates an electromagnetic field with the same frequency, and vice versa. The fundamentals are properly described by Maxwell's Equation. The Slingram or Horizontal Loops (HLEM) EM system illustrates the concept (Figure 6.6). On one side a current-carrying loop with a vertical axis produces, when energised, a magnetic field time varying over time. This loop is called the EM transmitter (Tx), the magnetic field is called the *primary field.* If within the space around the

Figure 6.5 Interpretation of Figure 6.4, showing areas devoid of basalt (shaded), fault lines and basalt thickness (Courtesy of the Dept. of Water Affairs, Government of Botswana)



transmitter a body with good electrical conductivity is present, the primary field will induce an *electro-magnetic force* (emf) in that body and an alternating current will flow. This induced current will in turn create an electromagnetic field in and around this body. This field is called the *secondary field.* Primary and secondary fields and the induced current are all of the same frequency. At the surface and at a given distance from the EM transmitter, another loop of wire with a vertical axis is located. This loop detects the vector sum of the primary and secondary em fields and is called the receiver coil (Rx). If we move the Tx-Rx assembly as a unit over the earth, i.e. maintaining a constant Tx-Rx separation and orientation, the measured resultant field will follow the variation in conductivity of the subsurface. In effect, the strength of the secondary field depends on the electrical (i.e. conductivity) properties of the subsurface. Therefore we can

Figure 6.6 Principle of the Slingram EM technique. An electromagnetic field is generated by a transmitter coil. Through induction this primary field generates a secondary electromagnetic field, which is detected by a receiver coil (From McNeill, 1990)



detect and localise conductive features without any electrical contact with the ground. Because the receiver observes the total strength of primary and secondary fields, the transmitter is fitted with a reference coil which detects the strength of the primary field. To establish the secondary field, the primary field is subtracted from the total field. In practice the receiver is designed in such a way that it measures not only the magnitude of the resultant field but also its delay relative to the field produced by the transmitter. This delay is called the *phase shift*. To analyse this delay, the receiver measures two phase components of the resultant field: one 'in phase' with the primary field and the second the 'out of phase' or 'quadrature' which is delayed by ¼ of a full (sines) cycle relative to the primary field. This information will be useful at the interpretation stage in order to determine a conductor's position and depth and its rating as a conductive body. The principles of EM are well described by McNeill (1990).

(i) Electromagnetic properties of rocks

The significant rock properties for EM surveying are the electric conductivity σ and the magnetic permeability μ. The magnetic permeability is often very nearly equal to that of vacuum (μ_0). Even when a formation contains a large concentration of a magnetic mineral such as magnetite, the magnetic permeability is usually less than 10 times μ_0. The rock's electric conductivity σ is mostly a function of the rock mineralogy, its porosity and its pore water's conductivity. The most common conducting minerals are the clay minerals, the metallic oxides and sulphides and graphite. The relationship between the rock conductivity, or rather resistivity, and its pore contents is described in Section 6.4.

The dependence of the rock properties σ and μ are subject to the condition that $\sigma >> 2\pi f \varepsilon$

(in which ε is the electrical permittivity or dielectric constant of the medium and f the frequency). This is usually the case when f is below 100kHz. At higher frequencies, so-called displacement currents are generated. Most EM systems operate at frequencies below 100 kHz and the condition is met.

The depth capability (or 'investigation depth') of EM systems is a function of the geometry of the EM transmitter-receiver pair (coil separation) and of the resistivity of the medium in which the magnetic field propagates. Amplitude (and energy) of the EM waves are attenuated in a conductive medium: the lower the resistivity and/or the higher the frequency, the higher the attenuation and the shallower the penetration of this medium. This effect is called the *skin depth limitation*. The *skin depth* is defined as the depth at which the amplitude of the EM field is reduced (attenuated) to ~1/3 (1/e exactly, with e = 2.7....) of the value it would have if the material were a perfect insulator. At the relative low frequencies normally used with EM systems, its numerical value is equal to:

$$\delta = \frac{503}{\sqrt{\sigma \cdot f}}$$

where δ is the skin depth (in metres), σ is the electrical conductivity of the medium (in Siemens/m [= inverse of resistivity]) and f is the frequency (in Hertz).

(ii) EM induction: frequency and time domain

The previous paragraph described the generic operation of an EM instrument. Variations relate to the size of and distance between the transmitter and receiver coils used.

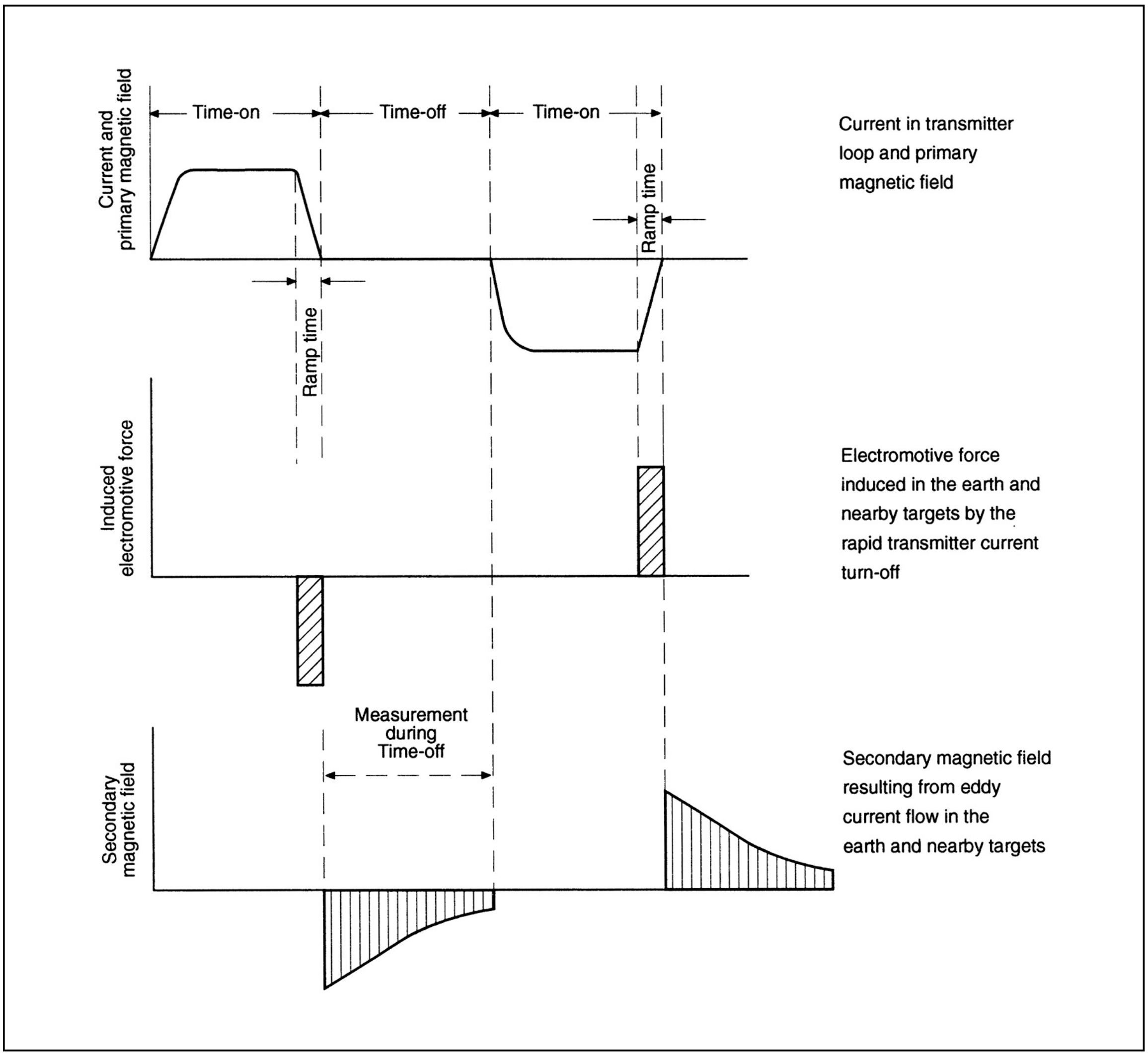
The now almost obsolete TURAM technique uses a large (several hundreds of metres in a square) transmitter coil with a small portable roving receiver measuring strength and dip angle of the secondary field. The selected frequency will control the depth investigation.

The so-called VLF system uses existing EM transmitters located at a very large distance. VLF refers to the *very low frequency* (15 kHz to ~ 30 kHz) of submarine radio communication stations, which are distributed all over the world. The field strength of these stations is large enough to generate a detectable secondary field thousands of kilometres away. Modern VLF instruments can also use the higher frequencies of long wave (LW) frequency civil broadcasting stations. The VLF system is a simple and fast technique but with a relatively shallow depth of penetration. Many VLF receivers measure selected geometrical characteristics of the resultant field (dip angle, ellipticity etc.) rather than its amplitude, which is a function of various external, uncontrollable, factors including the wave propagation conditions.

The SLINGRAM systems use relative small, equally sized and portable transmitter and receiver coils. Frequencies are selectable in the range of 200 Hz to ~ 56 kHz, allowing for a variety of depth investigations. They will display either 'in phase' and 'quadrature' response or, for particular configurations, directly the electric apparent conductivity. Some instruments display both. A setting where the target (aquifer) has distinctive electric properties and is located inside a complex stratification of alternating resistive and conductive layers, can be resolved with Slingram systems, provided a range of frequencies and coil separations can be used. This may prove to be a rather time consuming task.

The so-called Time Domain system or Transient EM (TD or TEM) has been specifically designed to cope with the more complex situations and to replace the Slingram system. The principle is based on the transmission of an impulse or signal with a step function (Figure 6.7). When the transmission is switched off, the induced emf will decline in time. With this, the secondary field will also decay in time. The time span in which these effects take place is very short, usually a period of microseconds. The rate of decay and the duration depend on the electrical properties of the earth. The decay is recorded by measuring the response in time slices

Figure 6.7 Principle of time domain EM. During the off-time of the transmitter, the decay of the induced secondary field is recorded in 'time slices' or 'time windows' (From McNeill, 1990)



or *time windows*. The effects of shallow layers are represented in the 'early' slices, while the deeper layers are presented by the 'late' slices. The geometry of the Tx-Rx systems is often different from FD systems (such as VLF and SLINGRAM systems), although there is considerable overlap between FD and TD systems geometry. Time domain systems are often implemented with medium to large transmitter coils (a square of tens to hundreds of metres per side) and little or no separation between Tx and Rx. The controlled variables are the loop size and some characteristic of the basic timing such as the loop current pulse width. Usually the loop size is optimised according to the expected local geoelectric section. Time domain systems often use a moving square loop laid out on the ground. The secondary magnetic field is detected by a loop or a magnetic sensor or with an assemblage of three orthogonal sensors.

(iii) EM survey techniques

Both TURAM and SLINGRAM techniques are used in a profiling or mapping fashion, depending on the objective of the survey. The TURAM technique is presently not very favourable, as it needs the layout of a very large transmitter loop, which limits speed of operation and flexibility. Profiling is carried out by making observations at more or less regular intervals (stations). Mapping consists of series of more or less parallel profile lines, preferably equally spaced. This allows a similar processing and presentation as in magnetic surveying. Profiling and mapping will result in an overview of the horizontal variation of the EM response (and consequently the earth's conductivity) in a surveyed area. These techniques are therefore comparable with resistivity profiling, as discussed in Section 6.4.

The SLINGRAM technique can also be applied in a depth sounding technique (frequency sounding), by making in one position a series of observations with different frequencies and coil separations. Processing of this data will result in a geoelectric cross section of the surveyed location in terms of layer thickness and layer conductivity.

The TD systems can both be used in a profiling and a depth sounding mode. For a while, clear distinction was made between profiling and sounding work. Currently many surveys are done in a hybrid mode: the survey is run in a profile mode but the quantity of information recorded at each station is somewhat equivalent to a sounding. The depth sounding mode requires observations at a range of selectable frequencies. Modern instruments will collect the data automatically according to a programmed sequence. EM survey techniques – both FD and TD – can also be performed in an airborne mode. To this end transmitters and receivers are mounted either on a fixed wing aircraft or on a helicopter. As with other airborne geophysical techniques, large areas can be covered in a relatively short time. Airborne EM surveys must be conducted at a low ground clearance in order to maintain response levels. Also the line spacing must be minimised. As a consequence, this type of surveying is more expensive than airborne magnetometry. The use of a helicopter is particularly expensive.

6.3.2 Data processing and interpretation

Simple interpretation tools are available in the form of interpretation diagrams. They are the result of the compilation of analogue scale modelling data. Responses of scale models have been collected for popular EM systems, such as the Slingram. Discrete sheet-like conductors have been modelled in particular. In such cases, the effects of variation of azimuth, position, depth, dip and conductance (sheet thickness times sheet conductivity) of a hypothetical target have been studied. The implementation of analogue scale models has been reviewed by Frischknecht (1971).

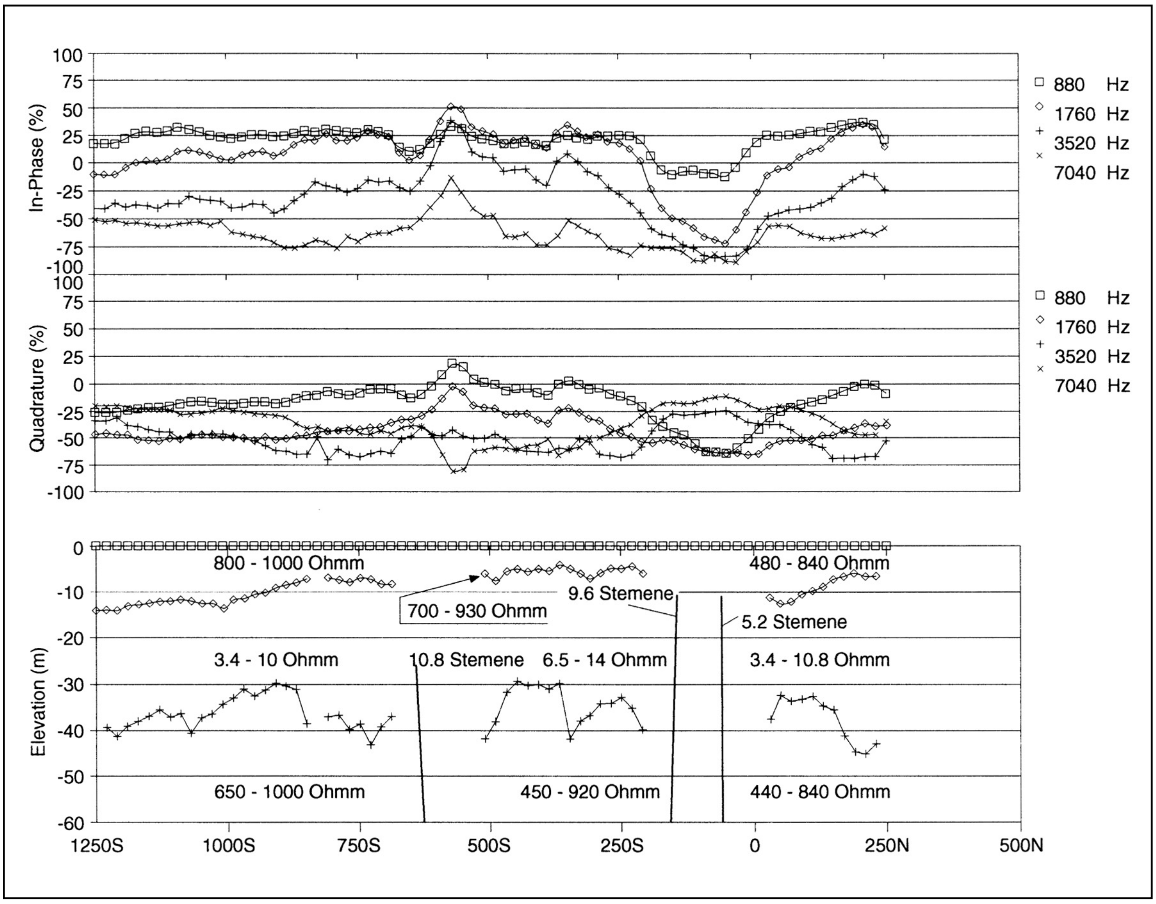
Simple interpretation schemes for a TD EM system include the determination of the apparent resistivity from the decay curves characteristics, and of the conductance of a discrete conductor using a simplified conductor's shape and dimension together with the response time constant. For a data set from a tri-axial receiver, the dip of the secondary field may be calculated for several stations along a profile and the migration of an equivalent induced current axis is mapped as a function of delay time after the termination of the primary pulse; this gives an indication of a conductor's location, depth and attitude or dip.

Numerical modelling of common FD and TD systems is widely available for the 1D case and to a lesser extent for 2D and 3D cases. A significant contribution in this domain was made by Anderson (1979). His routine is used in a number of application programs. Typical application programs include a forward modelling mode, which calculates the response of a given model for a specific EM system, and a data inversion mode, where the forward modelling component is included in a least squares inversion routine to find the best fitting model to a set of field data. This combination is particularly useful for the data interpretation stage, as well as for the survey planning and execution stage. Various programs are available for data entry, editing, reduction and display. These programs includes text editors, general purpose numerical spreadsheet software packages which include a graphic display capability, programs specially designed for geoscientific mapping (e.g. GEOSOFT) and EM numerical modelling packages . Low cost numerical EM modelling program are available from the USGS, Denver, CO, United States of America (Anderson, 1979).

(i) Case studies

Figure 6.8 shows an example of ground surveyed frequency domain EM as applied to a groundwater survey in Botswana ('Palla Road' project). The geology can be roughly described by a sequence of a few tens of metres of Kalahari Sands covering 100 to 200 m of basalt, under which silty and sandy sediments occur. The survey was carried out along a profile, taking multiple frequency readings at every station. Data processing and interpretation show that a thin conductive layer occurs above the basalt, probably representing saline or brackish groundwater. In several places the data cannot be interpreted as a layered earth model due to the presence of faults or fault zones. Several zones are indicated, with their conductivity thickness, depth and dip. These zones are potential locations for borehole drilling.

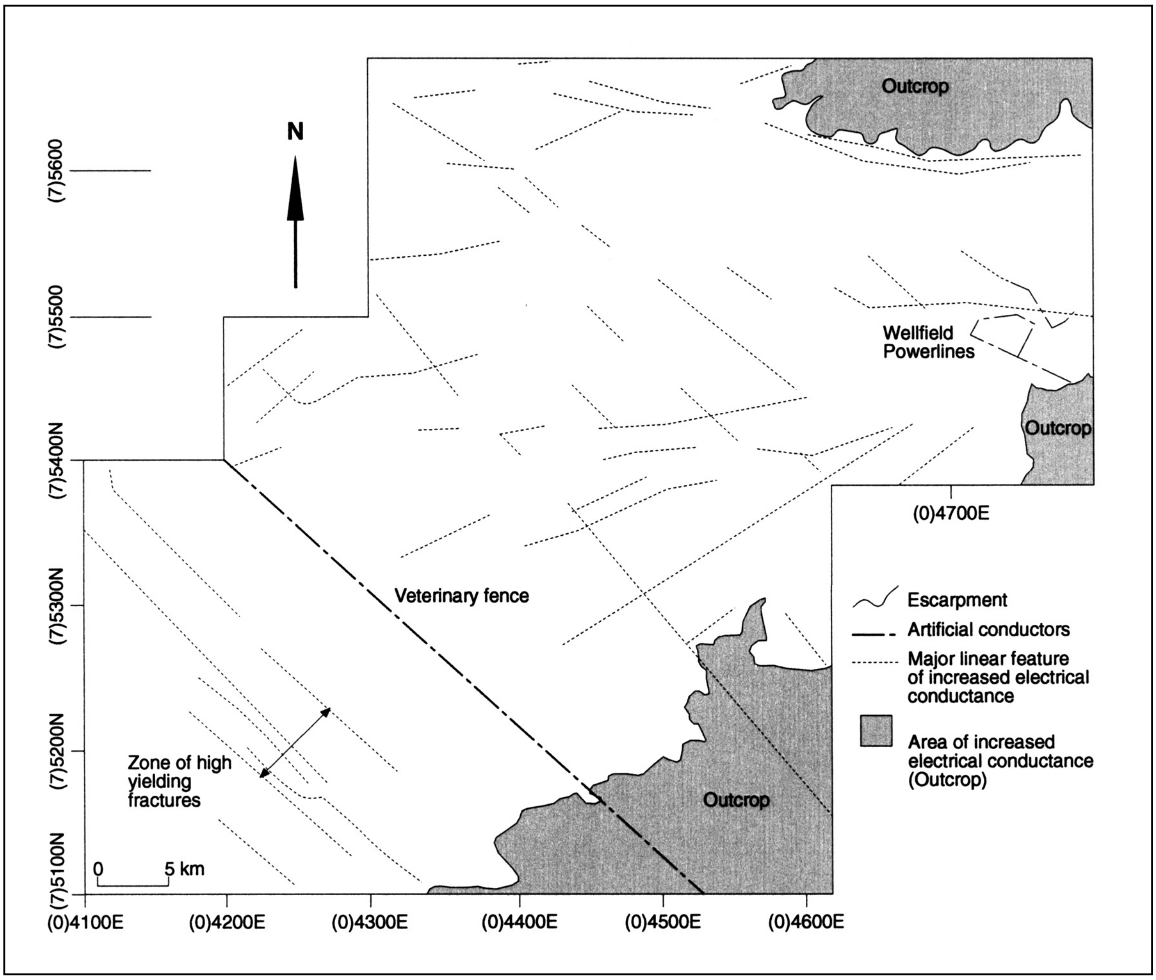
Figure 6.8 Example of a frequency domain EM profile over dyke intrusions and faults in basalts (Palla Road area, Botswana), made with a MaxMin horizontal loop instrument. The dyke intrusions generate less obvious anomalies than the faults, which can be recognised by a typical sequence of relative high response ('shoulder') flanking a low response ('trough'). The geometry and magnitude of the response allows interpretation of the more conductive faults in terms of conductivity thickness and establishment of dip and dip direction. Note that the resistive dykes are much better defined in the resistivity profile of Figure 6.4.3 (Courtesy of the Dept. of Water Affairs, Government of Botswana)



Helicopter-borne EM systems are usually multi-configuration (both coaxial and coplanar coil geometry) and multi-frequency implementations on a bird (rigid frame) towed 15 to 20 metres below the helicopter. Sengpiel (1986) published an example of such a survey with an operating frequency of around 900 Hz, using a coaxial and coplanar coil configuration over the sandy island Spiekeroog, off the German North Sea coast. The data was used to calculate the thickness of a fresh water lens. The survey was flown at a 100 metres line spacing and totalled 40 line-km. The interpretation, supported by vertical electric sounding, showed the largest thickness of the lens to be some 56 m. Similar surveys were carried out in Pakistan.

Several airborne VLF systems have been used on a commercial basis. From a groundwater exploration point of view the systems from which the dip of the resultant field could be determined are particularly interesting. Such a system was used in the Serowe project in Northern Botswana (Bromley et al., 1994). A regional sandstone aquifer is covered with a basaltic layer also providing an excellent magnetic marker horizon in a region of horst and graben covered by the Kalahari beds (sand, silcretes and calcretes). Weak VLF elongated conductors were mapped as a result of 7,500 line km multiparameter airborne survey. These VLF conductors (Figure 6.9) correspond to fractured zones which give the highest yields in the investigated area.

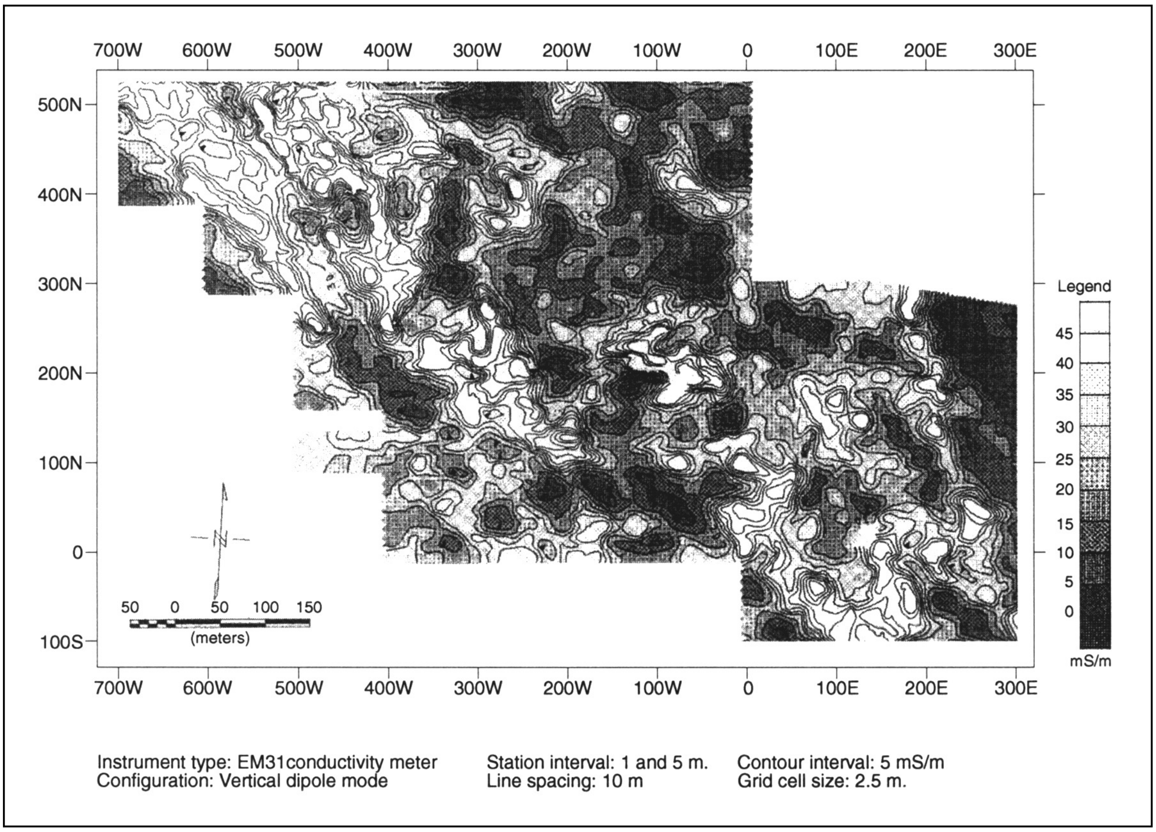
Figure 6.9 Example of the interpretation of airborne VLF data from a survey over Serowe, Botswana. Interpretation shows position of major fractures, which have potentially the highest yield of groundwater (From Bromley et al., 1994)



The interpretation strategy also involved aeromagnetic interpretation in a very effective structural and fracture mapping technique.

Figure 6.10 shows the results of a high density sampling of conductivity, using a Geonics EM 31 conductivity meter. This instrument presents direct reading of the apparent conductivity, using a single frequency of 9.6 kHz and a coil separation of 3.6 metres. The depth investigation is about 6.5 m. The survey was conducted over a dolomitic terrain south-east of Moura, Portugal, with shallow karst dissolution features and is an adaptation of airborne techniques brought down to the ground. Sampling was done at a 1m interval and with a line spacing of 10 m. Conductivity readings would vary from 8 mS/m to 60 mS/m over a distance of 20 m or less. The high conductivities probably correlate with karst zones which represent areas with enhanced rainwater infiltration.

Figure 6.10 Results of a single frequency horizontal loop EM survey (Geonics EM31) in the Moura area, Portugal. Areas with high conductivity are associated with shallow karst features and faults, which represent areas with enhanced infiltration of precipitation



(ii) Recent developments

The availability of convenient digital recorders and key electronic components and the improvements in instrumentation performance have led to EM systems with progressively higher resolution. From a spatial point of view readings may now be made relatively efficiently at a high sampling density. This is especially valuable for detailed shallow investigations. Another development is the extension of the equivalent frequency range and recording of a large number of time windows (20 or more) in TEM. Work is presently being done to bridge the gap between traditional EM frequencies and radar frequencies: New instrumentation and interpretation tools are being developed along these lines (Pellerin et al., 1994). The interpretation tools will no longer be limited to the quasi-static approximation when the condition $\sigma >> 2\pi f \varepsilon$ is met.

6.4 Resistivity method

6.4.1 Resistivity of rocks

In resistivity surveying, the electrical properties of rock materials are established by measuring the resistance to a galvanic current flow (direct current, DC). Current transport in rocks can take place by means of free electrons, as in pure metals, by non-compensated electrons within imperfect mineral crystals or by free ions within mineral crystals, which could be regarded as 'solid electrolytes'.

The resistivity of earth materials is a material property. Usually the Greek letter ρ (rho) is used for this property. The dimension of ρ is Ohmmeter or Ohmm (Ωm). The inverse dimension for ρ is *conductivity* (Siemens/m: $\sigma = 1/\rho$).

Rocks usually have a high resistivity because they consist of an agglomeration of minerals of which the bulk is resistant. High resistant minerals are quartz, feldspars, calcites and dolomites. Many minerals are conductive, such as clay minerals, magnetite, hematite, graphite, pyrite, pyrrhotite and many sulphidic minerals. However, these minerals occur only in a relatively low concentration in rocks and do not make these rocks conductive. Table 6.2 lists a few rock types with their resistivities.

Table 6.2 Resistivities of some rock types and fluids

Fluids	*Resistivity (ohmm)*	*Rocks*	*Resistivity (ohmm)*
Oil, gas, air	∞	Sandstones	100–10,000
Rain water	30–1,000	Shales	10–1,000
Soil moisture	1–100	Limestones	$50–10^5$
Sea water	~ 0.2	Dolomites	$100–10^5$
Clays	5–100	Metamorphic rocks	$100–10^6$
(River) sands	50–10,000	Gneisses	1,000–10,000
		Lavas	300–10,000

The electrical resistivity of rock is controlled by two components: the solid particles (with the binding cement) and the pore fillings. If the pores are filled with water the rock generally shows a relative low resistivity. The relation between the resistivity of the bulk of the rock ρ_f and other rock parameters, also known as Archie's Law (1942) is:

$$\rho_f = a \cdot \rho_w \cdot \varphi^{-m} \cdot s^{-2}$$

in which:

- a = texture factor, ranging from 0.5 to 1.5;
- ρ_w = resistivity of the pore water;
- φ = porosity as a volume fraction;
- m = cementation factor, ranging from 1.3 to 2.6; and
- s = saturation of the pores (as a volume fraction).

This relation is also presented by:

$$\rho_f = F \cdot \rho_w$$

in which F is called the electric formation factor or formation resistivity factor. These two expressions play an important role in groundwater exploration, as they can be used to correlate formation resistivities with the hydraulic properties of aquifers of sedimentary formations. Archie's Law is an empirical one and based on studies of sandstones. The presence of clay minerals in the rock matrix leads to serious deviations from this relation. The resistivity of pore water ρ_w is an important parameter, as it is directly related to the TDS (Total Dissolved Solids) in water and therefore gives an impression of the 'quality' of the pore water.

6.4.2. Resistivity measurements

(i) Principles and instrumentation

The resistivity of rocks is usually determined in the field using an array of four electrodes, which are placed in a straight line (Figure 6.11a). The instrumentation consists of a power supply (which can be an arrangement of batteries or a generator), current and potential electrodes, cables and a sensitive and accurate voltmeter and current meter.

Electrodes A and B are used to run the electrical current I into the ground (*current electrodes*). By running the direct current (DC) into the ground, an electric potential field is generated. The M and N electrodes measure the difference ΔV between the potentials at the

Figure 6.11a Principle of the resistivity method, using two current electrodes and two potential electrodes. Strata with different resistivities affect the current distribution in the earth

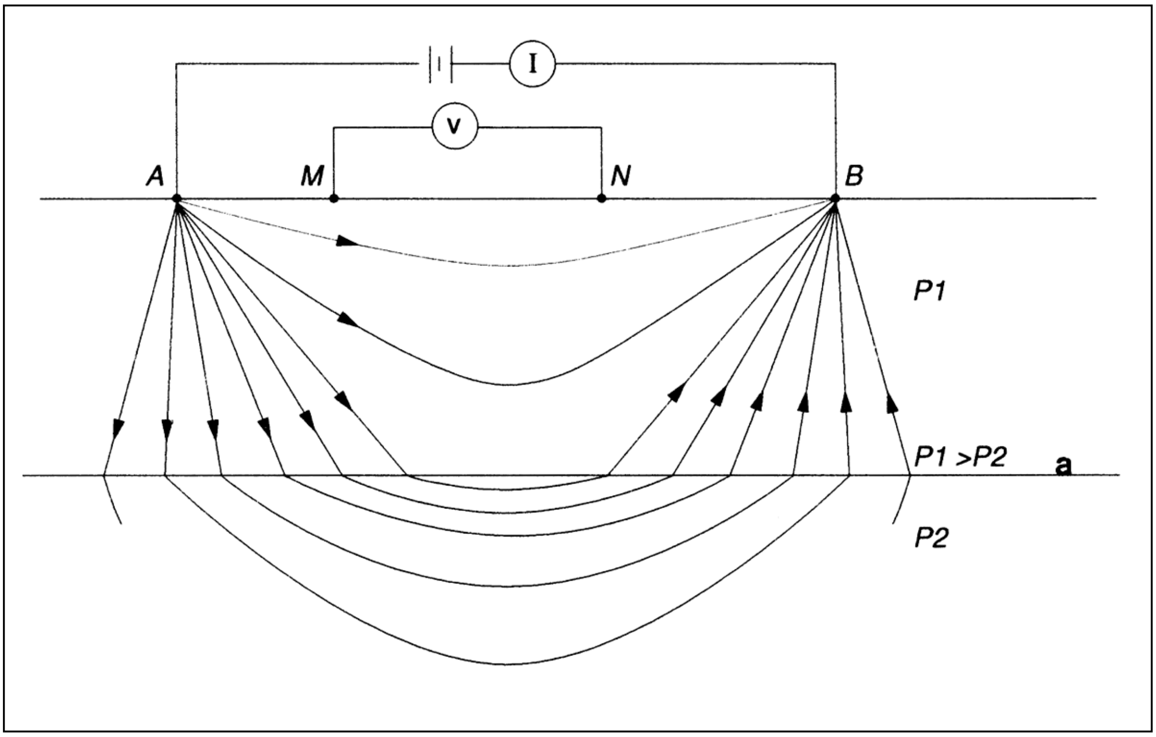
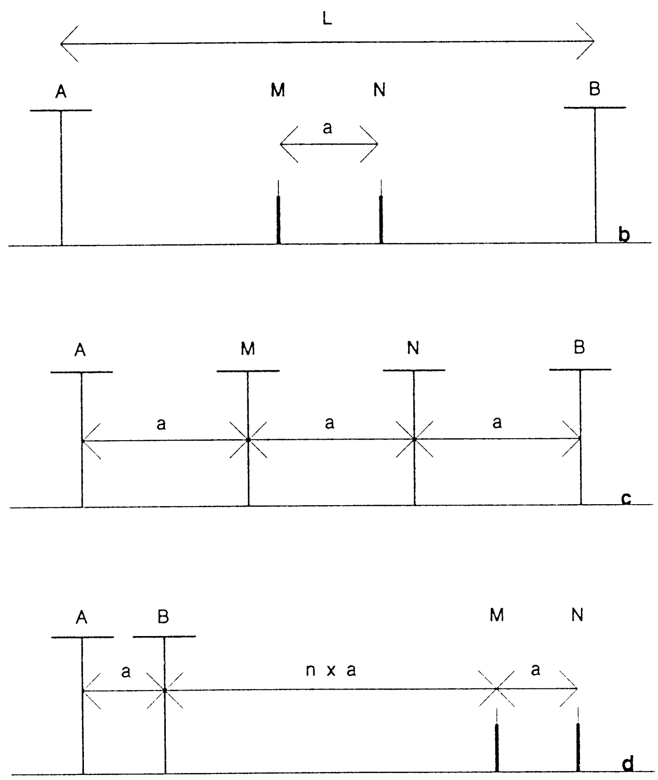


Figure 6.11b, c, d Layout of different electrode arrays: Wenner, Schlumberger and Dipole-Dipole



positions of these electrodes *(potential electrodes).* The potential difference varies with the position and the geometry of the four electrodes. The resistivity is calculated from the relation: $\rho = K.\Delta V/I$, in which K is called the *Geometrical Factor*.

Because the earth is not homogeneous and isotropic, the measured resistivity is generally addressed as *apparent resistivity* ρ_a: the resistivity appears to belong to a homogeneous earth, which is not the case.

The variation of resistivity for an inhomogeneous earth is caused by the distortion of current flow lines and with that, of the electric potential field around the potential electrodes (Figure 6.11). For a stratified earth, the distortion is systematic, with the consequence that through processing and interpretation the stratification of the earth can be derived from a systematic collection of resistivity data at one location.

(ii) Electrode configurations

The most common arrays are the Wenner, the Schlumberger and the Dipole-Dipole (Figure 6.11b, c and d).

In the Wenner array, the electrodes are placed at equal distances a. In the Schlumberger array the distance AB must be at least 5 x MN. AB can be 100 x MN or more, depending on the available power supply. By this layout the potential electrodes measure the potential gradient. In the Dipole-Dipole array, the distance a between A and B is equal to the distance between M and N, while the distance between the two dipoles is a multiple of a. For special type of surveys, the Dipole-Dipole array can be used in a variety of non-linear configurations.

Each of the different arrays has its advantages and disadvantages and should be used according to the technical and economic requirements of the planned survey. Sometimes one current electrode and/or one potential electrode are placed at an 'infinite distance', resulting in so-called half-Wenner or half-Schlumberger arrays (3 electrodes) or pole-pole arrays (2 electrodes).

In resistivity profiling, or trenching, apparent resistivity observations are made with a fixed electrode array at different locations (stations) along a survey line. The interval between stations is preferably kept constant. The profiling technique allows the variation of earth resistivity along the profile line to be studied, while the depth of investigation is more or less constant. In practice, often two, three or more observations per station are made with increasing current electrode separation, so that, to a limited extent, also the variation of the resistivity with depth is recorded. An example of profiling is shown in Figure 6.12. The two and three electrode arrays can be particularly useful in mapping structural features like faults and contacts by profiling. A detailed discussion of responses can be found in Telford et al. (1990) and in Keller and Frischknecht (1966).

Resistivity mapping is a variation on the profiling technique. It consists of a number of not necessarily parallel and regularly spaced profile lines, so that apparent resistivity data is obtained over the whole of the survey area. These mapped resistivities can be contoured, thus yielding an overview of anomalous levels of apparent resistivity.

Generally the three survey techniques can be used with any of the electrode configurations and very often a survey consists of a combination of depth sounding and profiling or mapping.

(iv) Data processing and interpretation

The processing of profiling data is usually minimal and consists of presenting the data graphically. Figure 6.12 shows part of the results of a full Schlumberger array profile over an East-West striking resistive dyke in Botswana. The anomaly shows two resistive peaks, separated by a minimum. The response is caused by a single dyke with a width approximately equal to the current electrode separation (AB/2 = 100 m). It is interesting to notice that this dyke would be less clearly present in an EM survey with moving coils run along the same profile line.

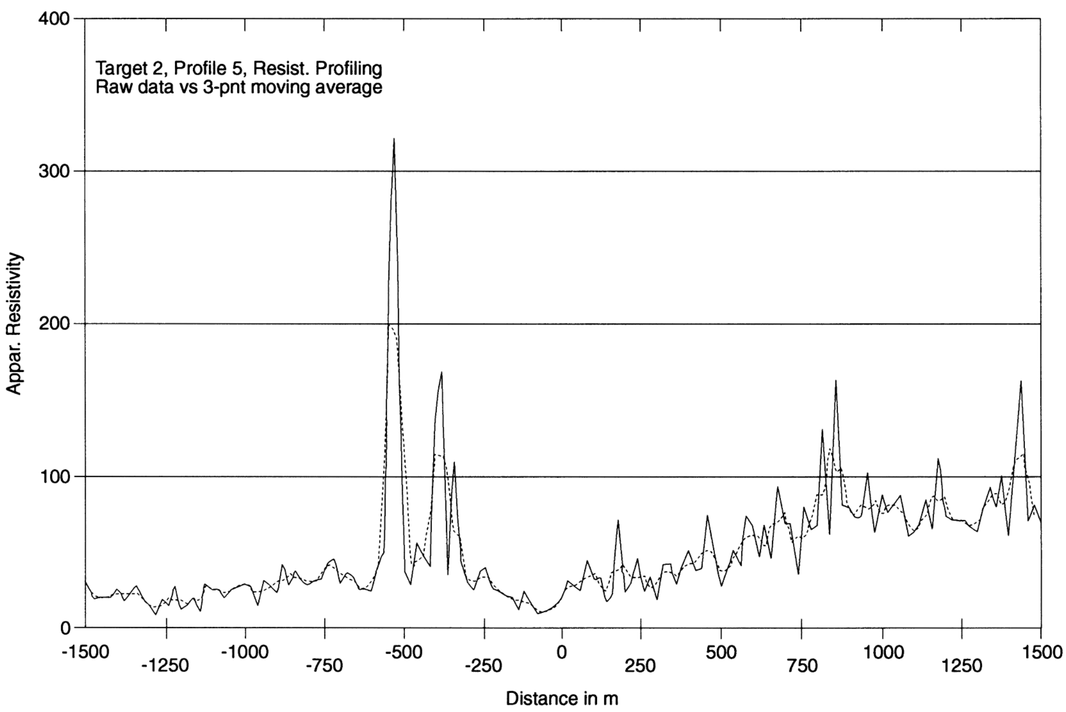
Depth sounding data requires more preparation before being interpreted in terms of the thickness and the resistivity of layers. A 'traditional' way of interpretation is the curve-matching technique, but such a process is tedious, time consuming and not very accurate for multi-layered field curves. Fortunately, it is now common to use a microcomputer. Simple programs allow 'curve matching' on the screen. More sophisticated ones allow fully automatic direct interpretation of field data, or controlled optimisation of layer parameters through the use of inverse modelling techniques. Use of such computer programs also make it possible to handle large numbers of VES data in so-called batch processing. Figure 6.13 shows an example of the output of the RESIST program that can handle Wenner, Schlumberger and Dipole-Dipole array data (Van der Velpen and Sporry, 1992).

The interpreted earth models can be used to construct resistivity cross sections that can be correlated with geology and lithology to model the survey area. Cross sections can be combined into fence diagrams, presenting a subsurface model in three dimensions. Such a model can be used to support the aquifer modelling in terms of extent, thickness, depth and variation of lithology (read: permeability) in the aquifer.

(v) Resistivity interpretation and aquifer evaluation

An important step in aquifer evaluation is the correlation between geophysical and hydrogeo-

Figure 6.12 Example of resistivity profiling over a sequence of Kalahari sands and basalts in Botswana. The high resistivity peaks around station –500 are caused by a sub-outcropping resistive dyke intrusion (Courtesy Dept. of Water Affairs, Government of Botswana)



logical parameters. During the 1970s and 1980s quite a few researchers worked on this problem (Kosinsky and Kelly, 1981; Mazac et al., 1985; Sporry et al., 1991). The importance of such a correlation (if successful) is that hydraulic transmissivity values can be predicted for every depth sounding location, thereby dramatically improving the control of the groundwater model. Figure 6.14a presents an example of such a correlation for the La Paz-El Carrizal Graben south of La Paz in Baja California, Mexico. The resulting transmissivity Contour Map is presented in Figure 6.14b.

(vi) Recent developments

Availability of low cost electronics has led to the development of equipment and procedures based on multicore cables and multi-electrode arrays, which allow the operator to switch through a complete sequence of electrode separations and positions without leaving his central operating position. Computer controlled instruments can store the acquired data automatically on magnetic diskettes, for processing later. Recently, instrumentation for continuous profiling and/or depth sounding has been developed. In-field processing is at present not an exception. A disadvantage is that the increased cost of these instruments vis-à-vis the 'traditional' ones cancels out the gain made in production and in the reduced number of manpower.

The data of Figure 6.13 was collected using a multicore *Barker cable* (Barker, 1981) in the Offset Wenner array. In principle, this sounding can be made in the field by one single operator, although work would proceed much more efficiently with the help of a field assistant.

Figure 6.13 Example of a VES in the Wenner array, representing apparent resistivity data (crosses), the interpretation in a block diagram (dotted line) and the model curve based on the layer parameters (table). The fit between the model curve and the field data is displayed in terms of RMS error. This depth sounding was performed using the Wenner Offset system over a sequence of fresh water (38.7 Ohmm), brackish water (10 Ohmm) and saline water (1.6 Ohmm). Location: Kapelle-Biezelinge, the Netherlands

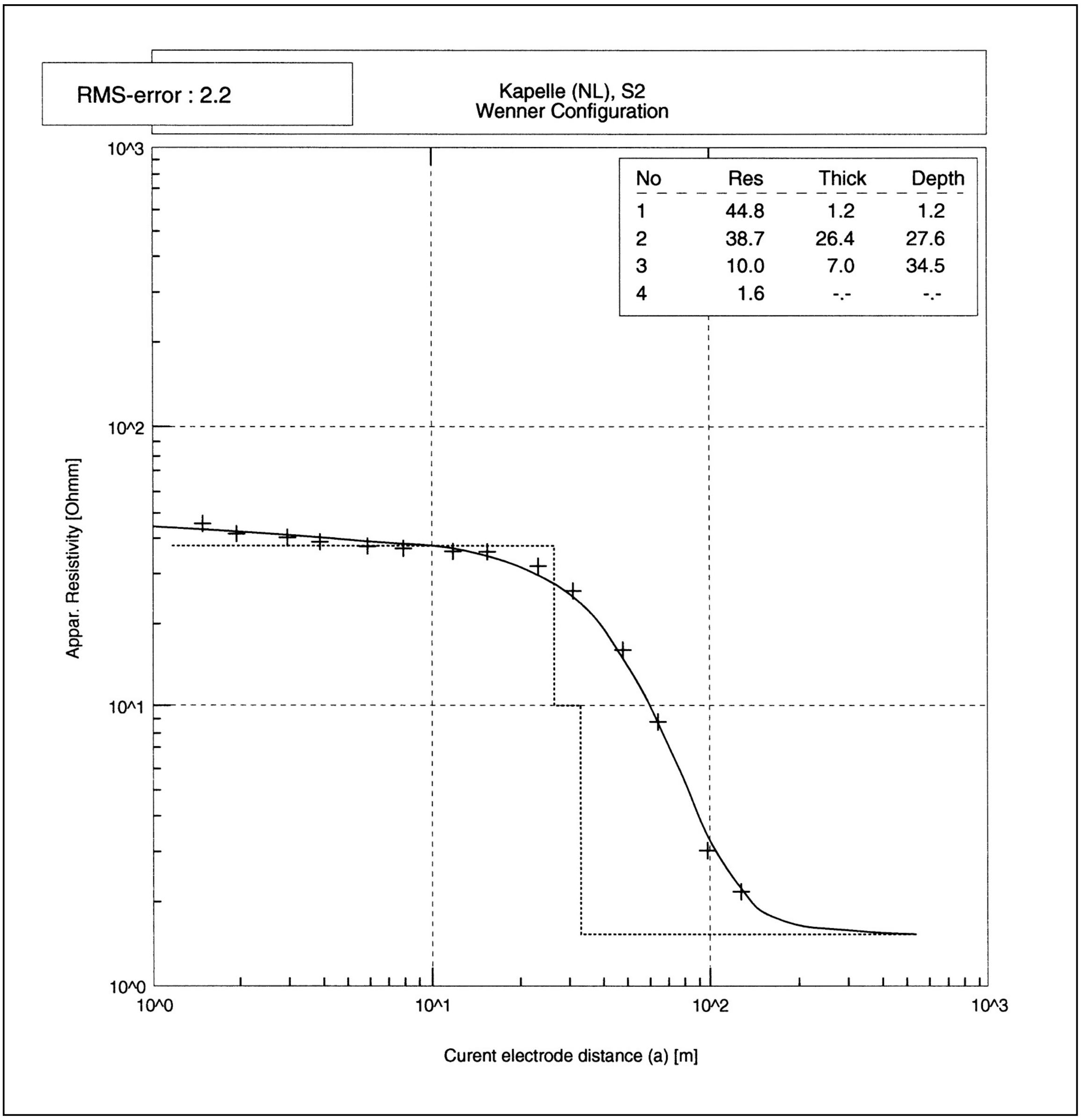
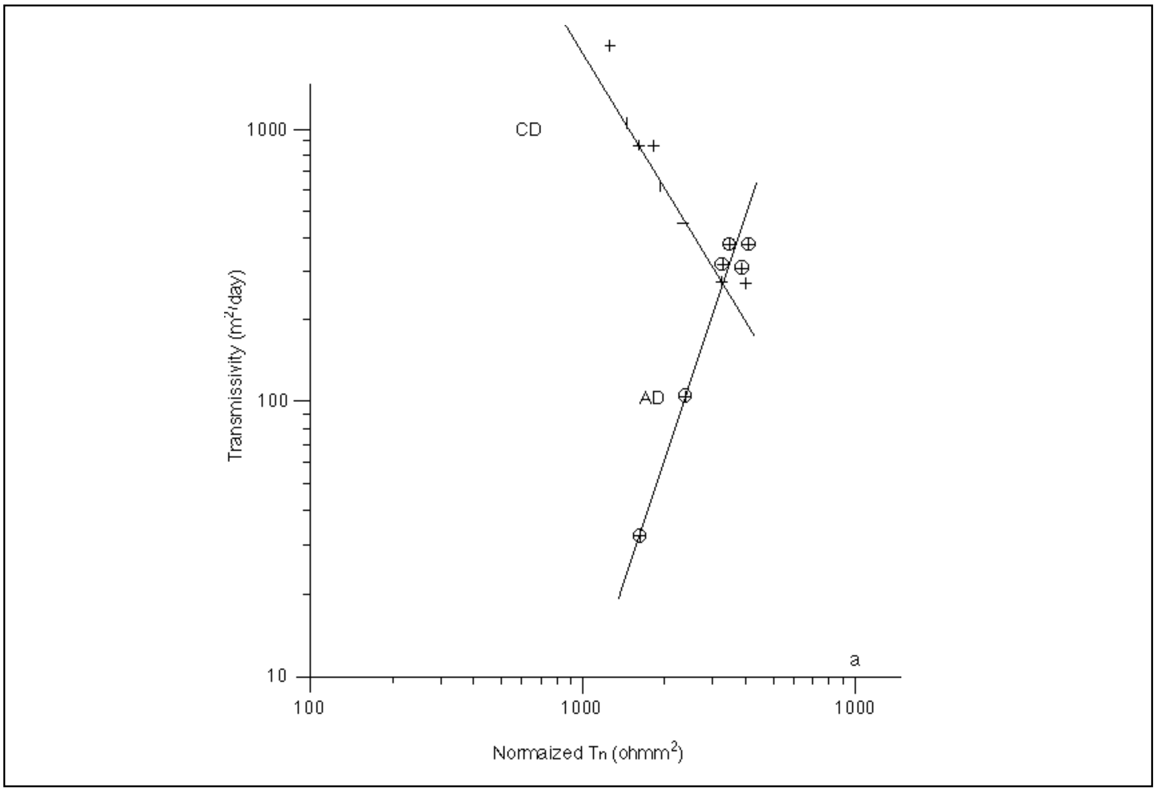


Figure 6.15 shows an example, from The Netherlands, of the great detail that is possible with a continuous sounding system (Van Overmeeren and Ritsema, 1988).

6.5 Induction polarisation

The Induction Polarisation method or IP is an electrical method combining the physical effect called induction polarisation, with the principles of the resistivity method. Consequently the technique of data collection is similar, but more complex equipment is used. As with the EM

Figure 6.14a Correlation of hydraulic transmissivity values with normalised transverse resistance T_N, calculated from aquifer formation resitivities for the aquifer of the La Paz-El Carrizal Graben, Baja California, Mexico. Two different correlations (AB and CD) can be observed, valid for the locations A, B, C and D on the transmissivity map of Figure 6.14b (From Sporry et al., 1991)



methods, this method originally was developed for mineral exploration, but eventually it was realised that groundwater exploration could benefit from it too.

6.5.1 Principles and procedures of IP

An IP survey is in principle carried out with resistivity equipment, normally in the Dipole-Dipole configuration, although the Gradient array (a variation on the Schlumberger array) is also used. It therefore requires current and potential electrodes, connecting cables, a power supply and the instrumentation to measure current and potential difference. While using the equipment to measure the IP effect, the apparent resistivity is also measured in the conventional way. When a current is run into the ground and switched off, a decaying potential difference may be observed *after the moment of current switch-off.* This effect is called induction polarisation.

The observed potential at the potential electrodes will thus show a response as illustrated in Figure 6.16. The decay of the voltage may last milliseconds to a few seconds, basically 1,000 times longer than in EM methods. The level of this effect and the period of decay depends on the IP properties of rocks.

(i) Sources of the IP effect

IP is caused by two major effects, called the *membrane effect* and the *electrode polarisation effect.* Both are electro-chemical processes and are caused by the interaction between the minerals of the rock and the electrolytic behaviour of the pore water. When current flows through or along the boundaries of the solids, the transport of electrical charges is hampered by reduction or

Figure 6.14b Hydraulic transmissivity contour map, based on the transmissivity values derived from aquifer formation resistivities using the correlation chart of Fig. 6.14a. The N-S oriented graben is broken up into blocks by two faults running E-W (not shown) (From Sporry et al., 1991)

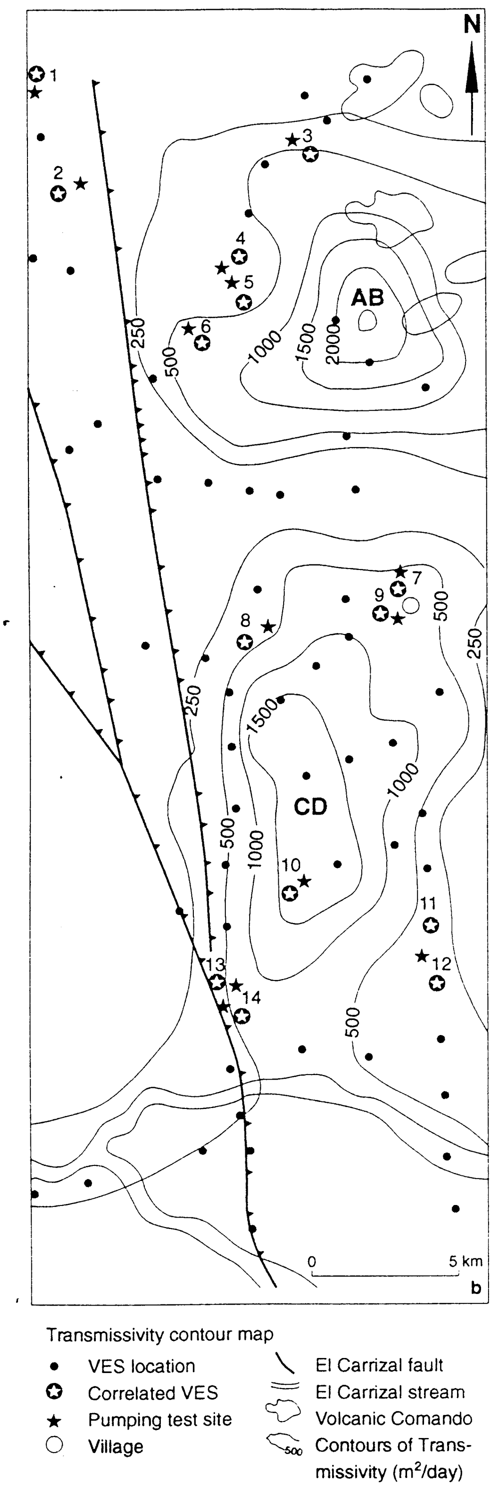
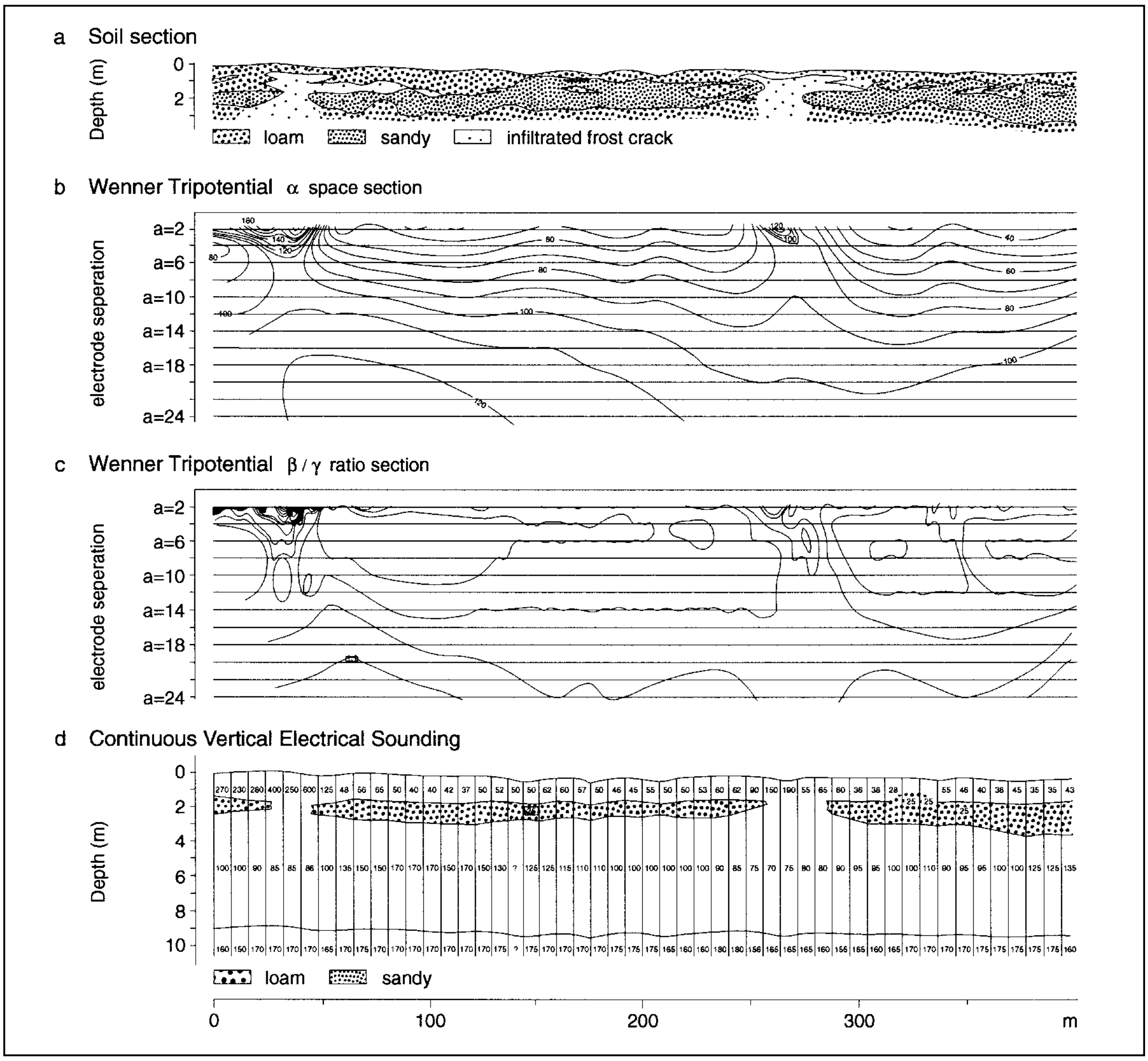


Figure 6.15 Example of a continuous depth sounding survey by electrode array scanning, imaging the shallow subsurface. Sand-infilled frost cracks representing hydraulic connections between surface sands and deeper sandy formations can clearly be identified (Van Overmeeren and Ritsema, 1988)

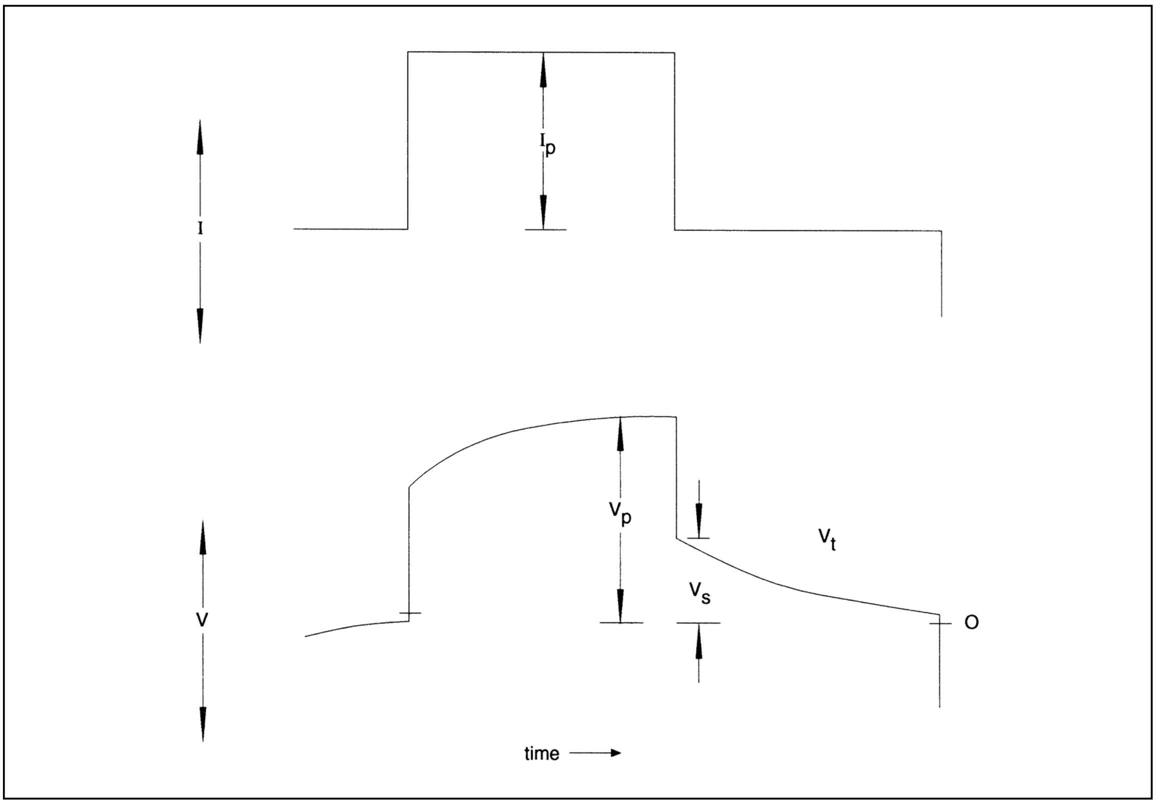


oxidation processes as well as by different mobilities of ions. As a result, electrical charges will locally build up (polarisation) and diffuse, or accumulate with a certain time lag in relation to the inducing current. A formation containing such polarisable grain surfaces will show a time-dependent current/voltage relation, which can also be presented as a frequency-dependent specific resistivity combined with a capacity-like phase shift.

The *membrane potential* is caused by an ion 'cloud' in the pore water. This cloud moves under the influence of an applied electric field. Some non-metallic minerals (clays) possess a negative surface charge, which is neutralised by 'free' positive ions in the electrolyte, thus forming an electrically charged 'membrane', consisting of alternating 'layers' of negative-positive-negative charges. When a current is applied, the neutral membrane is disrupted and when the current is switched off, a movement of ions due to redistributions takes place. The ionic movement creates a weak current flow, which is detected at the potential electrodes.

The *electrode polarisation* occurs when mineral particles block the flow of ions, carried by the electrolyte. The current flow is then carried by electrons within the mineral grains, which

Figure 6.16 Principle of measuring the chargeability in the time-domain IP technique. The induced primary current I_p is detected as a primary voltage V_p. At current turn-off, the voltage drops to a secondary level V_s and the transient voltage V_t decays with time. A theoretical measure of chargeability M is: $M = V_s/V_p$ (From Sumner, 1976)



results in a current-opposing build-up of electrical charges at the interface of particles and electrolyte (pore water). when the current is switched off, the bound ionic charge does not immediately disappear, but decreases in time due to diffusion back into the pore water. The electrode polarisation is generally the strongest effect and is particularly generated by sulphidic minerals. The membrane polarisation, however, is particularly generated by clay particles and is therefore useful in groundwater exploration. Where the resistivity and EM methods are not capable of differentiating between conductive clay-containing formations and groundwater formations, IP can indicate the presence of clay.

(ii) IP techniques

The IP effect can be measured in three modes: the *Time-Domain* method, the *Frequency-Domain* method and the *Phase* method. In the Time-Domain technique, the decay of the IP voltage is measured in 'time slices' or 'windows', as in Time Domain EM. The number of windows is in the order of 4 to 12. Because of the short period in which the decay takes place (less than 1 or 2 seconds), the voltages are measured and stored in a memory by a microprocessor. The physical property measured is called the *chargeability.* The chargeability M is defined by:

$$M = \frac{1}{V_c} \int_{t_1}^{t_2} V(t)dt$$

where M is the apparent chargeability, V_C is the potential value near the end of the on-time interval, t_1 and t_2 define start and end of the time window interval and V(t) is the decay potential (Figure 6.16).

In the Frequency Domain technique, two frequency excitation modes are used, usually in the range between 0.1 and 10 Herz. Due to the IP response the apparent resistivity will be lower for the highest frequency. These different resistivity values are used to calculate the so-called *frequency effect,* FE, and the *metal factor,* MF.

In the Phase Mode, the phase shift Φ between the excitation current and the IP potential is measured. This phase shift is directly proportional to FE. In this approach only one excitation frequency is needed.

(iii) Field procedures

In IP surveying the 'normal' apparent resistivity is measured (during on-time), together with the IP effect (during off-time). The process is repeated (in the Dipole-Dipole array) with increasing separations of the dipoles. The dipole separations run from 1 x a to 6 x a (a is the electrode separation within the dipole). Investigations to different depth can be performed using differently sized dipoles.

Although the apparent resistivity part of the survey is fairly straightforward, the observations of the IP effect are rather time-consuming, which makes the IP method a less popular survey method.

6.5.2 Data processing and interpretation

The observed apparent resistivities are plotted in a *pseudo section*. Conductive zones show up clearly in such a section. The chargeability values are also plotted in pseudo sections. A qualitative evaluation will reveal the presence of conductive fault or fracture zones, which may be potential (small) groundwater sources. More importantly, IP data can be helpful in differentiating between thick or thin layers containing clay, which is useful information in establishing the transmissivity of formations. Also, the IP method can be helpful in detecting pollution by organic compounds, such as oil and organic solvents.

Modern computer techniques have brought about the possibility to perform quantitative interpretation through the application of the finite element method in modelling. This entails varying the model parameters, in an attempt to get the model response to fit the field observations.

6.6 Seismic method

6.6.1 Principles and instrumentation

The seismic method is based on measuring the propagation velocity of elastic waves induced in rock formations by an explosion or impact. The velocity V is calculated from the travel time T of a wave over distance X between a source and a sensor: V=X/T. The waves propagate through the earth and are subject to reflections and refractions at interfaces between rock strata with different physical properties. The seismic signals which return to the ground surface are registered by arrays of sensors, called *geophones* when used on land or *hydrophones* when used in water. Cables connect the geophones to a recording device, which is the seismograph. The geophones consist of a magneto-dynamic element, which converts vibrations picked up from the earth into a small current, which is relayed to the seismograph. Modern seismographs consist of signal amplifiers and analogue-to-digital converters. The signal is digitally recorded on a magnetic storage device. The recorded signal can be displayed on a CRT or LCD screen and it can be printed on a paper printer for quality control. Multichannel recorders allow the simultaneous use of 12, 24 or more geophones. Because of the digital data storage techniques, computers can be used to process the field data efficiently.

A large variation of sources can be used. The smallest is a heavy hammer striking a steel plate on the ground. Other possible sources are dropping weights of a variety of sizes and from various heights, explosives and a range of mechanical or electromechanical devices generating impacts or vibrations. Each source type generates energy with a typical frequency bandwidth, which may be crucial to the success of the survey. Some of these sources put high demands on the skills, training and organisation of the field crews. The use of explosives is generally subject to licences for storage, transport and use. Sources that are cheap and easy to maintain or to replace (like a hammer and plate) have the disadvantage of a limited level of energy output.

The propagation of elastic waves through the earth is subject to several processes. It is not only the wave velocity that is important, but also the frequency content, attenuation (reduction) of the seismic energy, refraction, reflection and diffraction. play an important role.

Table 6.3 Elastic wave velocities in some media

Medium	*Velocity (m/sec)*	*Medium*	*Velocity (m/sec)*
Air	~ 330	Anhydrite	3,500–5,500
Water	~ 1,500	Limestones	3,400–7,000
Soil	~ 3,400	Sandstones	2,000–4,500
Sands	50–300	Shales	3,900–5,500
		Igneous rocks:	
Chalk	2,100–4,200	Granite	6,500–7,000
Dolomite	3,500–6,900	Gabbro	6,500–7,000
Gypsum	2,000–3,500	Ultra mafic rocks	7,500–8,500

(i) Wave velocities

The seismic wave velocity is expressed in m/sec. Two different groups of waves can be observed: body waves and surface waves. Most important wave types are the Compressional or P-waves and the Shear or S-waves, which are both body waves. Raleigh waves and Love waves are surface waves. In groundwater investigations, usually only P-waves are used. P-waves can travel through any medium: solids, liquids and gases. S-waves occur only in solids.

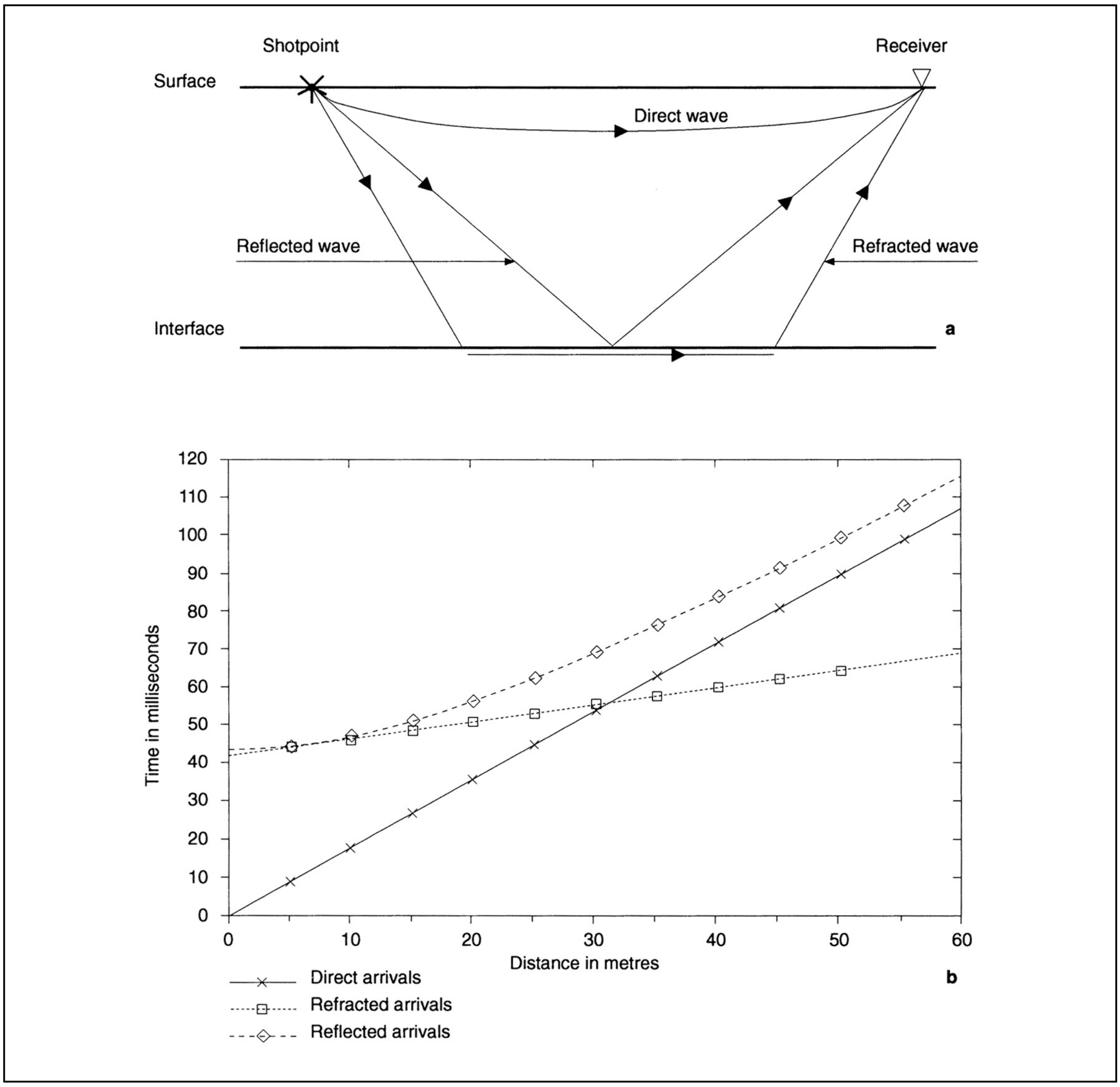
The velocity of a seismic wave is related to the elastic properties and the density ρ of the medium through which it travels. It has been found that $V_p > V_s$. Table 6.3 presents velocities of some common rock types.

Generally, it can be observed that for sedimentary rocks the velocity increases with geological age and depth of burial. All igneous rocks show high velocities in a fresh and unweathered state. The presence of groundwater affects the velocity of rocks strongly, depending on the porosity.

(ii) Wave frequencies

The seismic energy usually contains a wide range of frequencies, from as low as 1 Hertz to 1,000 Hz or more. It is found that high frequencies also lead to higher resolution, i.e. thinner individual beds can be detected (in reflection seismics) than with low frequencies. On the other

Figure 6.17a,b In a layered medium three different waves can be observed: the direct wave, the refracted wave and the reflected wave (a). Each wave has a different travel time, depending on the geometry of the layout, depth of the interface and the velocities of the individual layers (b)



hand, high frequencies suffer from a stronger attenuation than low frequencies. The result of this is that with an increased depth of investigation, the resolution also reduces.

6.6.2 Reflection and refraction of seismic waves

As the earth is composed of a sequence of sub-horizontal or dipping layered formations, the boundary effects at the interface of layers with different physical properties will result in the seismic energy being separated into a reflected part and a refracted part. These effects are controlled by the same laws as in optics. When a wave is reflected from an interface between two media, the angle of reflection equals the angle of incidence. Refraction of a seismic wave is controlled by Snell's Law: $V_1/\sin\theta_1 = V_2/\sin\theta_2$. V_1 and V_2 are the velocities of the two media and θ_1 and θ_2 are the angle of incidence and the angle of refraction respectively. Under specific circumstances ($V_2>V_1$) θ_1 can be such that $\theta_2 = 90°$. The refracted wave will then travel along the interface and is said to be critically refracted. Angle θ_1 is then called the critical angle.

(i) Travel times of seismic waves

Figure 6.17a shows that in a layered medium three waves can be observed: a direct wave, a refracted wave and a reflected wave. The travel paths are different for each of these waves, resulting in different arrival times at the geophone. This effect has led to the development of two main techniques: refraction seismics and reflection seismics. Figure 6.17b shows schematically the travel time curves of the three waves for a two-layered earth model. It is the result of measuring the travel times for different distances between source and receiver. Note that the direct and refracted waves show a linear response, while the reflected wave shows a hyperbolic response. Such a presentation is called a TX diagram or time-distance diagram.

In refraction seismics the boundary between two successive layers can only be detected if the seismic wave velocity is higher in the deeper layer than in the overlying layer. If the seismic velocity in the deeper layer is lower than the upper layer, no critical refraction will take place and the interface will not be detected. In addition, a relatively thin layer may not be detected by refraction seismics.

With reflection seismics there is no problem with detecting a low velocity layer, but here too, a thin layer may not be detected. Velocity inversion and thin layers will inevitably lead to errors in depth interpretation of refraction seismic data.

(ii) Refraction seismic surveying and data processing

In refraction seismic surveying a seismic line is made up of a sequence of one or more seismic spreads. Each spread consists of a series of geophone positions, usually 12 or a multiple of that number. A basic requirement is to fire at least two shots, one at either end of the spread: the forward shot and the reverse shot (the reciprocal method). From the records (Figure 6.18) the first arrivals are marked, listed and plotted in a TX-diagram, which forms the very core of the processing sequence. The first arrivals are always related to direct or refracted waves and are used to calculate the wave velocity for each layer, the intercept times for each geophone position and the depth of the refractors below each geophone. Relatively simple procedures, such as the 'Hawkins method' (Hawkins, 1961) or Hagedoorn's 'Plus-Minus Method' (Hagedoorn, 1959), to make these calculations can be performed in an organised table, or better still, using a spreadsheet program.

The present day approach of refraction seismic data processing is the General Reciprocal Method (GRM) (Palmer, 1980). It uses a minimum of 5 (sometimes 7) shots for each spread. One shot is always in the middle of the spread. The approach leads to a very consistent arrangement of geophonees with shots at regular intervals all along the seismic line. The GRM allows a detailed velocity analysis but is rather elaborate to process manually. Therefore the data is processed on a computer. Apart from resulting in a continuous depth section, the procedure will result in a detailed velocity analysis, presenting variations of the wave velocity along the refractor. This information can indicate the degree of weathering along a refractor, giving an indication of the presence of a potential aquifer. Changes in wave velocities may also indicate the presence of faults or fault zones covered by the seismic line.

In all processes based on forward and reverse shooting, the depth of a refractor is calculated from: $h = ½ . t_i . (V_1/\cos\Theta_c)$, in which t_i is the intercept time (travel time for X = 0). The intercept time can be calculated for every geophone position. $\cos\Theta_c$ is calculated from the velocities V_1 and V_2 (in a two layer case). For more than two layers, the depth of other refractors needs to be calculated in sequence of increasing depth.

Figure 6.18 Example of a seismograph recording with a 24-channel system. The positive amplitudes have been shaded black (variable area display VAR) to enhance the presentation of wavelets

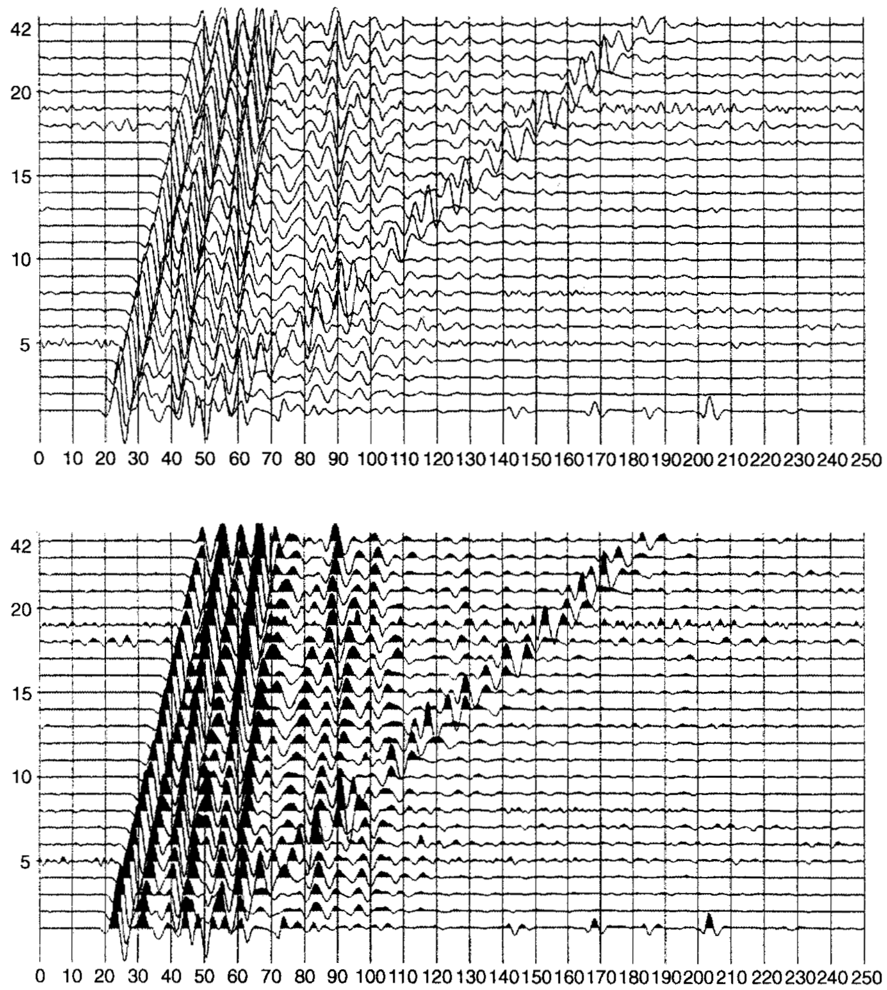
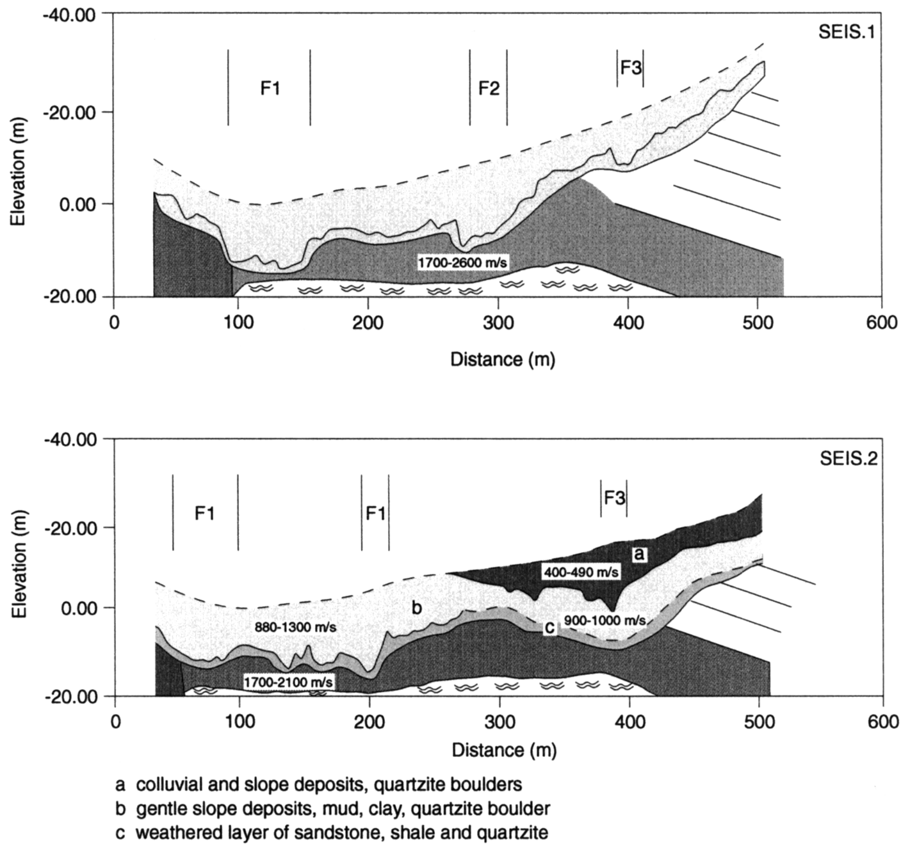


Figure 6.19 shows an example of a refraction seismic section across a narrow valley, in which fault zones are marked by increased depth of the refractor, due to weathering along the fault. The weathered zones are potentially water-bearing.

(iii) Reflection seismic surveying

Until the early 1980s, reflection seismics was not very popular in the relatively shallow applications that groundwater surveys represent. However, with the availability of cheap microprocessors and PC-compatible computers, this technique has proven to be a valuable tool for groundwater applications. Downscaling of the geometry, reduction of the amount of equipment and development of small, but efficient seismic sources has made its operation by a relative small crew of 5 or 6 men possible. The data can be processed on PC-compatible computers, with relatively cheap software, thus bringing the costs to acceptable levels for groundwater exploration. It must be realised that of all the various geophysical methods, reflection seismics is potentially capable of the highest possible resolution in subsurface mapping.

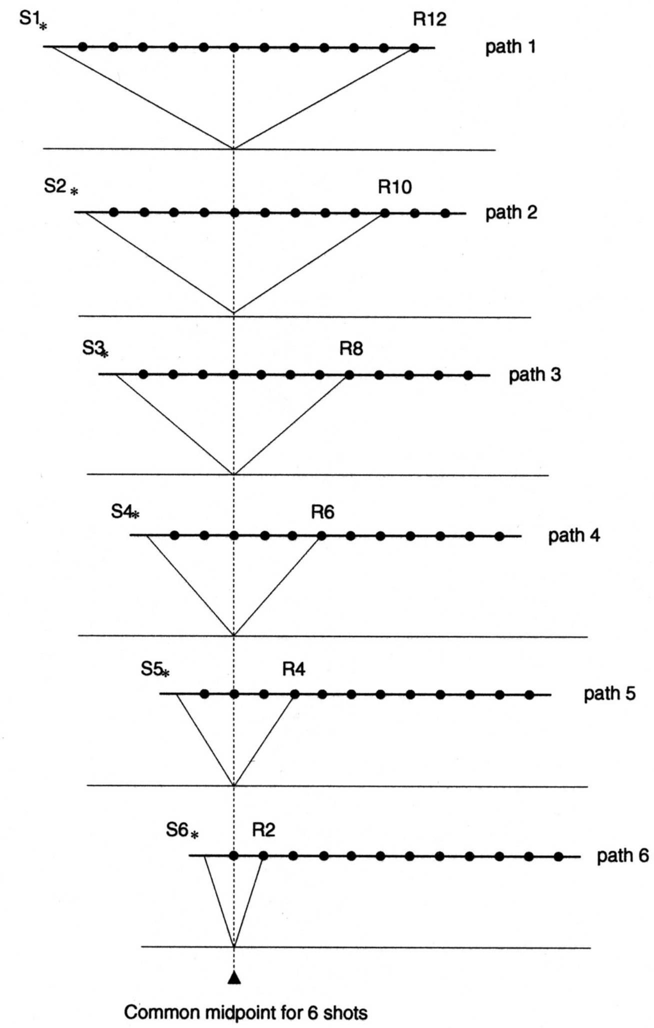
Figure 6.19 Example of a refraction seismic profile across a valley, showing the positions of three fault zones. The fault zones can be recognised by an increased depth to the refractor (bedrock) and locally lower velocities. Location: Santa Eufemia, Spain



In reflection seismics, use is made of the well established procedure of multifold coverage Common Depth Point or Common Midpoint technique. Figure 6.20 illustrates the basic concept, which starts with the data acquisition in the field. During this process, the seismic spread and the shotpoint move up one shotpoint interval (usually equal to one geophone interval) along the seismic line. Shotpoint intervals can be as small as 2 m, leading to a subsurface interval of reflection points of 1 m only. Reflections from sequential interfaces can be observed as hyperbolic events in the unprocessed field record (Fig. 6.18). Processing of this data is based on the correction of the Normal Move Out or NMO, by applying the NMO-velocity, also called the stacking velocity. NMO is the difference in travel time between a wave that travels along perpendicular to a reflector and a wave that travels at an angle to the perpendicular, as is usually the case when shotpoint and geophone are not in the same place.

The depth of a reflector can be calculated from the expression $Z = V_{nmo} \cdot t_0/2$, in which t_0 is the travel time along the perpendicular to the reflector and V_{nmo} the average (RMS) velocity of the wave to that interface. Vnmo is also called the stacking velocity. The processing sequence in reflection seismics includes trace sorting, velocity analysis, NMO correction, stacking, as well as frequency spectrum analysis, frequency filtering, corrections for surface level variations (static correction), deconvolution processes (to compensate for filtering effects of the earth) and

Figure 6.20 Principle of the CDP technique in reflection seismics. Through this survey technique t he same depth point (reflection point) is recorded 6 times (6-fold coverage), using an array of 12 geophones. Processing of this data will lead to an improvement of the signal-to-noise ratio as compared to a single coverage (Adapted from Kearey and Brooks,1991)

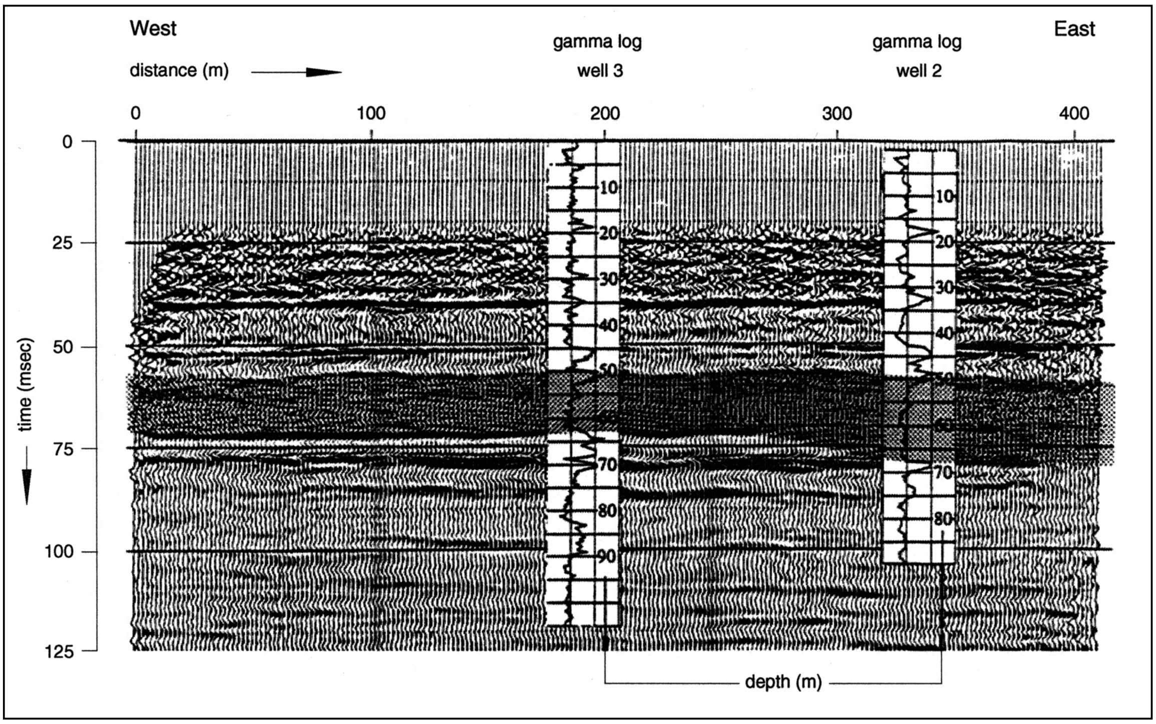


migration (to correct spatial positioning of reflectors caused by steep dips. The reader is referred to excellent publications on these subjects by Yilmaz (1987) and by Sheriff and Geldart (1982) and many others. A full discussion in this chapter would take up far too much space.

Figure 6.21 shows an example of a reflection seismic section in a geohydrological application. It demonstrates the continuity of an aquifer formation, which was suspected to wedge out, but actually continues at a reduced thickness. This example gives an impression of the excellent detailing possible with shallow reflection seismics though at a cost. One of the minor drawbacks of shallow reflection seismics is that, with a few exceptions, the first 20 metres remain ‘out of focus’.

Many examples of shallow reflection seismics have been published. Good examples are Hunter et al. (1984) and Steeples and Miller (1990).

Figure 6.21 Example of a high resolution seismic reflection section after full processing of the data, as an application in groundwater surveying. The grey zones mark an aquifer which shows a decreased thickness west of the 300 m point along the line. The section is combined with gamma logs, recorded in two boreholes in the section. Strong reflections correspond with the top of clay layers, which have a high level of natural gamma radiation (Meekes et al., 1990)



6.6.3 Recent developments

One recent development in shallow reflection seismics is the modification of 3D techniques. These, however are rather costly, requiring the use of seismographs with at least 96 channels to record on, and an even larger number of geophones. The investment in hardware required to use 3D seismics in groundwater exploration is prohibitive. However, the accelerated development of hardware and software may remove the price impediment.

Another development is seismic tomography. In this application recordings are made with the seismic source either at the surface or placed in a borehole and with the geophones in a borehole. Processing of the data allows imaging of the subsurface. Another new development is the study of surface layers, using shear waves .

6.7 Gravity surveying

Although seldom employed as a primary tool for groundwater exploration, the gravity method has proved to be extremely useful, both in the delineation of significant regional structures and in the direct detection of aquifer formations such as paleochannels, shallow karst cavities and fault zones.

6.7.1 Principles and field procedures

In gravity surveying, the quantity that is measured is the difference in acceleration due to gravity between one station and another at the earth's surface. The acceleration g is expressed in cm/sec^2 or in Gals. Practical units are milliGals and the gravity unit (g.u.), which is equal to 0.1 milliGal. The acceleration is measured with a gravimeter. The operating principle is based on the variation in length of a spring, which compensates the weight of a small mass. Modern instruments make use of electronic compensation of the weight. The differences in acceleration are caused by:

- *Earth tides:* the gravitational attraction of the sun and the moon impose a time-and position-dependent variation on the Earth's gravitational field.
- *Instrumental drift:* instruments of older design suffer from drift of the readings. Modern instruments are self-correcting.
- *Latitude:* the distance between the surface and the gravitational centre of the Earth is smaller at the poles than at the equator, causing a latitude-dependent increase of gravity from the equator to the poles. This effect is enhanced by the Earth's rotation. Corrections must be applied using the International Gravity Formula 1967 (IGF).
- *Elevation:* topographic variations also cause differences in measured gravity.
- *The differences are due to:* a) a decrease in gravity for stations at higher altitudes due to the increased distance to the gravitational centre and b) an increase in gravity for higher stations due to the extra downward attraction of the 'slab' of rock between the station level and the survey datum level. Both effects may be removed from the observations by applying 'Free Air' and 'Bouguer' corrections respectively.
- *Terrain surrounding the survey station:* hills and valleys within the vicinity of a station exert an upward attraction on the spring-balance mass of the gravimeter, thus reducing the observed gravity. This effect can be significant in hilly terrain and extremely mountainous areas. Corrections can be applied using topographical information (maps or digital terrain models).
- *Lateral variations of rock density:* Once corrections for the above factors have been made, the resulting 'Bouguer anomalies' may be interpreted in terms of variations in rock density between the observation platform (surface) and the centre of the earth. It is this very property from which geological units can be modelled. Table 6.4 presents density values of some common rocks.

Table 6.4 Density values of some common rocks

Rock type	*Density (t/m^3)*	*Rock type*	*Density (t/m^3)*
Sands and gravels	1.4–2.2	Granites	2.6–2.8
Clays	1.7–2.5	Lavas	2.8–3
Marls	1.8–2.6	Basalts/Gabbro	~ 3
Sandstones	2–2.6	Seyenite	2.75
Limestones	2.2–2.8	Diorite	2.85
Dolomites	2.2–2.8	Peridotite	3.3
Anhydrite	2.8–3	Schists	2.4–2.8
		Gneiss	2.6–3

Field procedures in gravity surveying essentially consist of making regularly spaced observations along a survey line or grid, during which station elevations also are levelled. Careful notes must be made of the elevation of the surrounding terrain and where possible, rock samples (preferably unweathered) must be collected so that the rock densities can be measured.

Since the acceleration due to gravity is proportional to the inverse square of the distance between two attracting masses, density variations closer to the observation platform have a proportionally greater effect on Bouguer anomaly values. However, deeper sources should not be neglected in the interpretation, particularly as they often contribute regional gradients to gravity data. For a full discussion of the principles of gravity measurements and data reduction, reference is made to standard geophysical textbooks (Telford et al., 1990; Kearey and Brooks, 1991; Milsom, 1989). Essentially, the Bouguer anomaly values, g_{ba}, are obtained by first correcting the measured gravity for temporal variations (tidal effects, instrument drift) to give the observed gravity, g_0, and then by correcting for all non-geological factors:

$$g_{ba} = g_0 - g_{th} - g_{fa} + g_b + g_t \qquad \text{mGal } (\text{m.s}^{-2} \times 10^{-5})$$

where:

g_{th} is the theoretical gravity value on the geoid at the given latitude of the gravity station,
g_{fa} is the free-air correction,
g_b is the Bouguer correction,
g_t is the terrain correction.

The accuracy of the final Bouguer Anomalies are dependent on the accuracy with which the corrections are calculated. Excluding instrument malfunction, the largest error would be caused by inaccuracies in station heights. The combined free-air and Bouguer correction is, on average about 0.2 mGal per metre, using an average crustal density of 2.67 t/m^3 for the Bouguer correction. Consequently, a height accuracy of ± 5 m, as might be achieved using barometric levelling techniques, would lead to an average accuracy of ± 1 mGal in the Bouguer anomaly.

Since the amplitude of Bouguer anomalies depends on the size, depth and density contrast of features of interest to the survey, it is not wise to generalise on these values; anomalies between 0.1 to tens of mGals may be important. It is important to consider carefully the 'expected' anomaly amplitude and wavelength. It would not be feasible, for example, to detect sub-mGals in mountainous areas, due to the inaccuracies in terrain correction estimates.

6.7.2 Interpretation of gravity anomalies

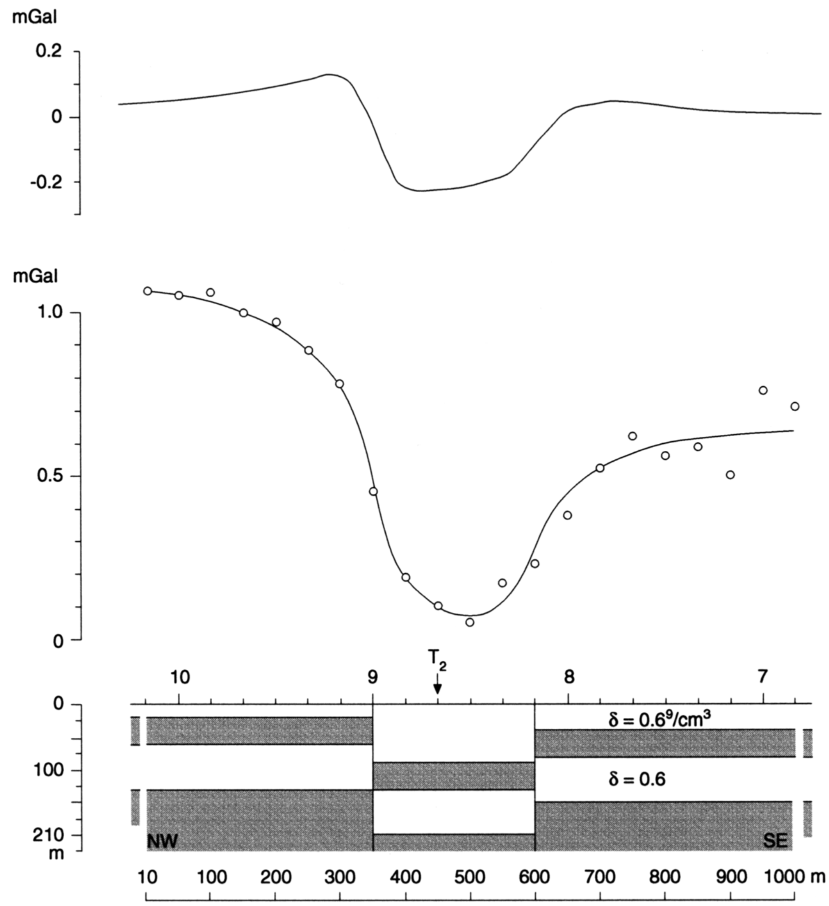
Bouguer anomaly maps may be interpreted qualitatively in terms of structural and lithological features. For example, a strong gravity gradient may represent the faulted juxtaposition of geological units with differing densities, or the edge of an intrusive body.

A gravity low may be caused by a siliceous intrusion or a sedimentary basin, since both granitic rocks and sediments are generally less dense than the remaining commonly occurring rock types. As in all methods based on potential field measurements, the non-uniqueness of possible solutions require *a priori* geological and physical constraints in order to arrive at acceptable model solutions. Also, the cumulative effect of lateral density contrasts at different depths will combine to produce a single observed anomaly. In the case of Figure 6.22 the upper sediment-bedrock interface was resolved by a seismic survey, the lower one was based on borehole information.

The recognition of regional gravity gradients plays an important role in the quantitative interpretation The regional gradient is generally due to a deeper or broader geological feature and removal or simultaneous modelling is essential for an accurate modelling of the target body of interest.

Systematic density measurements of all rock types in a survey area is often not a practical or economic solution. Usually a combination of sample measured values and standard ranges for given rock types is used, even realising that density variation may occur within one geological unit. Complex models to fit the observed gravity anomaly can only be justified if sufficient *a priori* information is available.

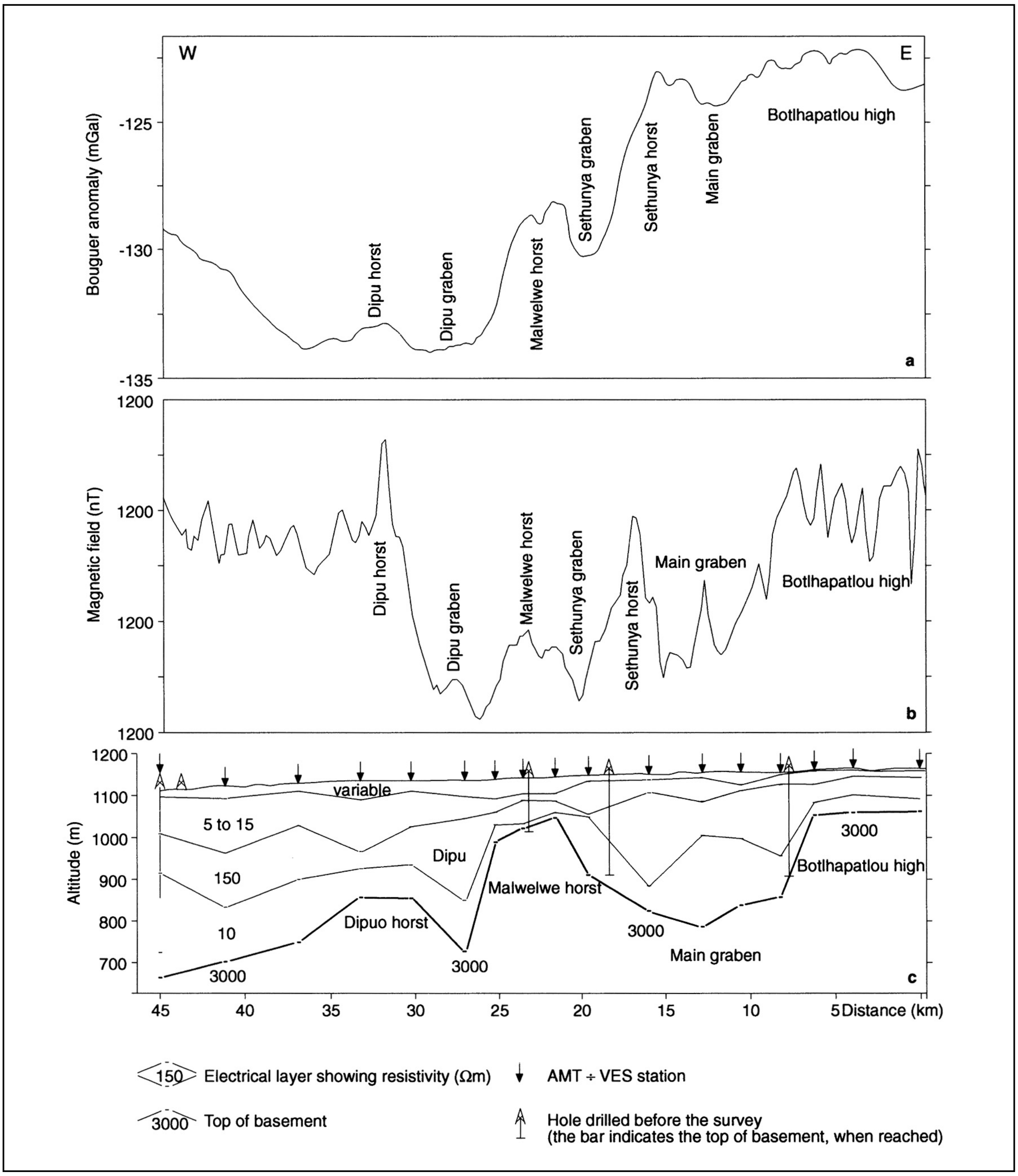
Figure 6.22 Example of the combined effect on a gravity anomaly by a shallow and a deep structure. The circles represent the observed residual gravity anomaly over a buried graben structure. The upper graph represents the calculated anomaly caused only by the displacement of the thin slab of high density material, whereas the lower graph shows the computed total gravity effect from both the shallow and the deep structure; the latter anomaly accords well with the observed data. The contrast in specific density was assumed to be 0.6 g/cm^3 for both the upper and lower contrast (Overmeeren, 1980)



6.7.3 Applications to groundwater exploration

A common application of gravity surveys to groundwater exploration is in the delineation and modelling of regional features such as sedimentary basins and horst-graben structures (Ghazala, 1994; Bourgeois et al., 1994; Alvarez, 1991). Correlation of the Bouguer anomalies with, for example, magnetic, magnetotelluric, VES and borehole data may confirm or refute gravity interpretations as demonstrated in Figure 6.23. In this example the regional gradient was insufficiently defined (too short profile) to calculate sedimentary thickness. Both gravity and magnetic anomalies could be interpreted as basement highs (<100 m) in the extreme west of this profile. AMT (audiomagnetotelluric) and borehole data refute this conclusion. The magnetic and gravity highs are interpreted as probably due to the presence of an intrasedimentary dolerite sill at this location. Although the gravity and magnetic data in this case was not used for quantitative modelling, the fact that the Sethunya horst was not derived from the initial modelling of the AMT and VES results is an example of the value of a multi-method approach. Additional

Figure 6.23 Gravimetry used for a structural study of a sedimentary basin, supported by AMT (audiomagnetotelluric), magnetic and resistivity (VES) surveying. (Bourgeois et al., 1994). Gravity and magnetic anomalies in the western part of the profiles could be interpreted as basement highs. AMT and borehole data refute this conclusion (Bourgeois et al., 1994)



constraints are also required to avoid potential confusion between sediment-basement topography and intra-basement density contrasts (Van Overmeeren, 1981).

A gravity low may be caused by a siliceous intrusion or a sedimentary basin, since both granitic rocks and sediments are generally less dense than the remaining commonly occurring rock types. As in all methods based on potential field measurements, the non-uniqueness of possible solutions require *a priori* geological and physical constraints in order to arrive at acceptable model solutions. Also, the cumulative effect of lateral density contrasts at different depths will combine to produce a single observed anomaly. In the case of Figure 6.22 the upper sediment-bedrock interface was resolved by a seismic survey, the lower one was based on borehole information.

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Figure 6.24 demonstrates how small (sub-mGal) anomalies due to faulted offset of the basement-sediment interface can be detected by careful surveying and data reduction. To achieve the required accuracy it necessitated the levelling of gravity stations to a precision of ± 0.1 m and careful consideration of terrain corrections. The resulting Bouguer anomaly profile showed a noise envelope of 1 mGal. A smoothing filter was applied to remove short-wavelength noise and to highlight the geologically significant anomalies. Correlation with faults derived from aerial photo interpretation and with VES results added confidence to the geological model. Note here that the regional gradient is modelled as the broad N-S slope of the density interface; outcrop evidence is in agreement with this conclusion.

High sensitivity semi-regional gravity surveys have been successfully applied to the direct detection of paleo-channel aquifers (Carmichael and Henry, 1977; Angelito et al., 1991) giving typical anomalies of 0.2 to 2.0 mGal. Secondary porosity in karst terrain can be sometimes detected by microgravity surveys (Butler, 1984), but there is a depth limit, and the expected results should be careful considered before embarking on such a high-resolution survey. Gravity data alone can not distinguish between water-filled and air-filled cavities, but under favourable circumstances a combination with resistivity data may resolve this ambiguity.

6.8 Ground penetrating radar

The ground penetrating radar method (GPR) has been in use for a number of years, mainly for engineering geology applications and for shallow investigations (less than 10 m). Recent developments, again stimulated by advances in electronic and computing technology, have made it an interesting and extremely useful tool for detailed groundwater investigations.

GPR uses electromagnetic waves of frequencies between 25 to 200 MHz, the domain of radar waves. When radar waves transmitted by a radar antenna penetrate the earth, reflection and transmission processes take place, similar to those of reflection seismics. Reflection takes place at the interface between layers with contrasting electric properties. Reflected waves are received by a receiver antenna which registers travel times (in nanoseconds) and amplitudes in the receiver system. The transmitter and receiver antennas can be used in a fixed mode, or in a mode in which one of the antennas is moved away from the other at regular intervals. Figure 6.25 demonstrates the principle of the method.

(i) The di-electric constant

The degree of reflection is determined by the contrast of the dielectric constants of the materials on either side of the interface. The dielectric constant varies from a value of 4 for dry sands to 65

Figure 6.24 Careful surveying allows the detection of faulted offset of the basement-sediment interface. Intersecting faults show up as small gravity anomalies. Location: Moura, Portugal

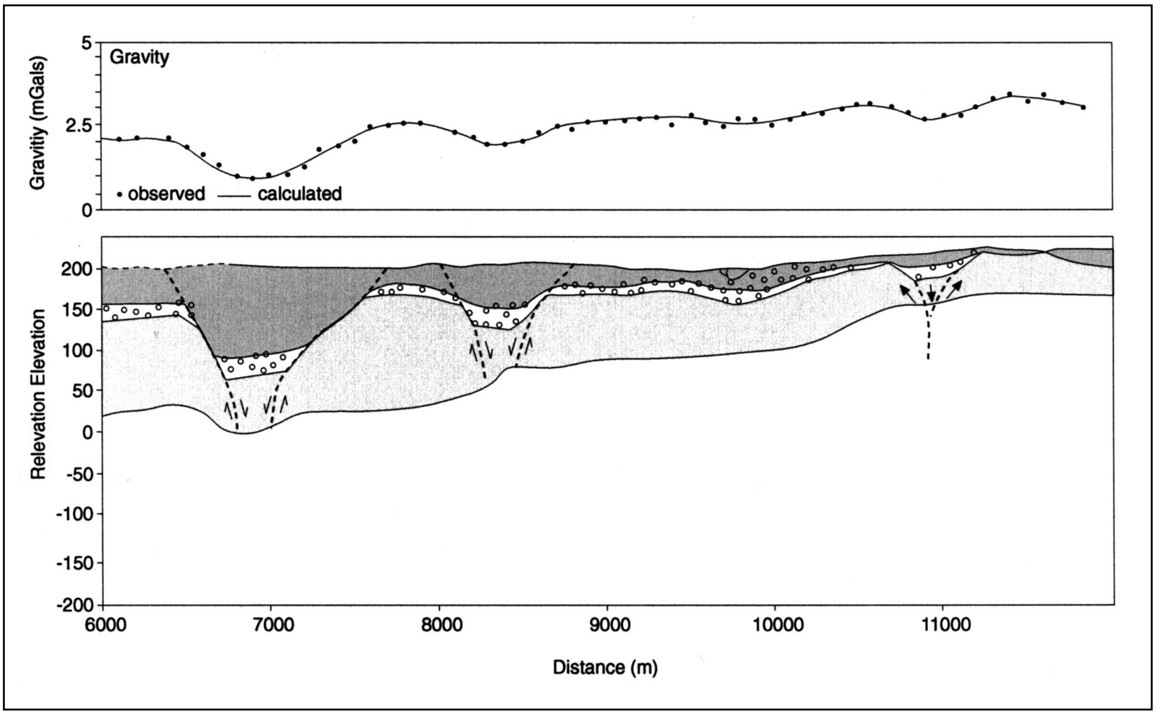
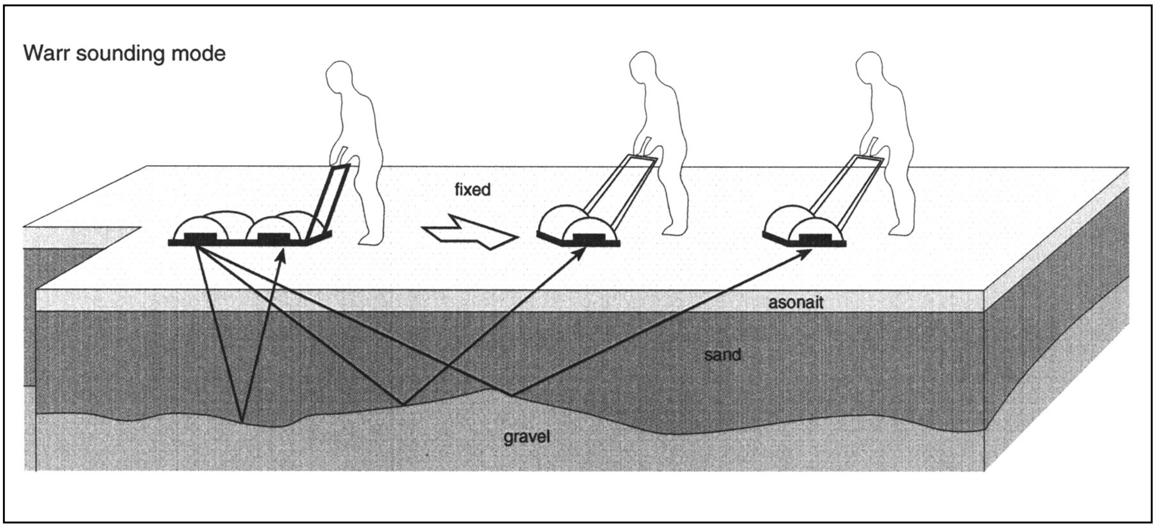


Figure 6.25 The sounding mode for GPR. The principle is similar to the one used for reflection seismics (CMP or CDP technique) and allows for the calculation of wave velocity in each layer and the calculation of depth to each reflecting interface



for peat (Van Overmeeren, 1994). Fresh and saline water have a dielectric constant of 80. The high contrast between dry and water-saturated materials has the consequence that a water table generally acts as a good reflector for radar waves. Crystalline and metamorphic rocks usually show relatively low values of 5 to 15. Only gabbro may have a value as high as 40 (Telford et al., 1990).

(ii) Depth penetration, damping

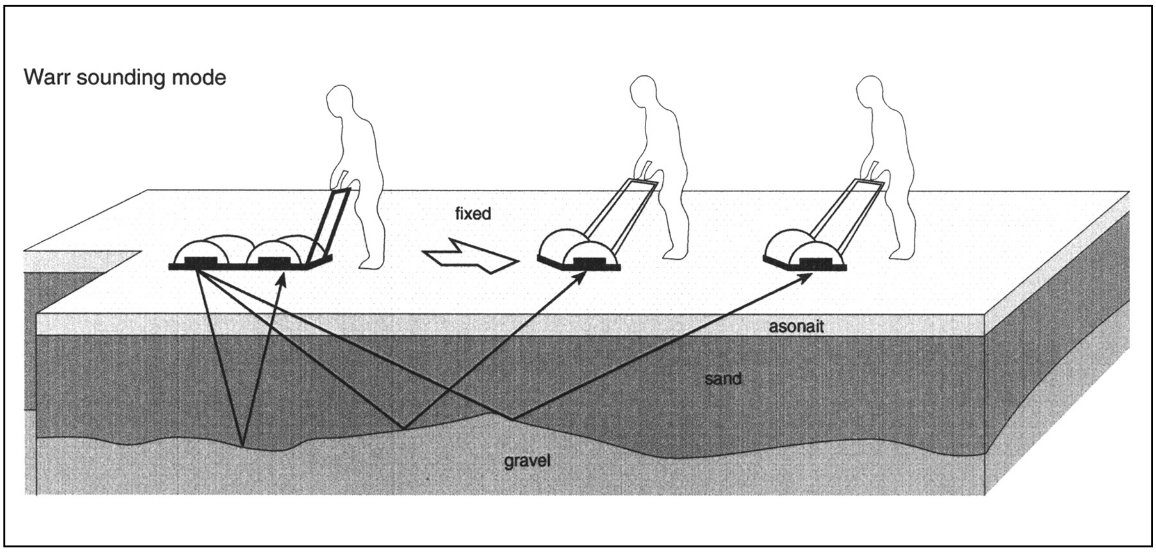
The depth penetration of radar waves depends on the resistivity ρ of the medium and on the frequency. As a consequence, the penetration depth of radar waves is rather limited in a low resistive environment. Van Overmeeren (1994) reports that in a sandy formation the penetration depth was reduced from some 40 m with 25 MHz to around 10 m with 200 MHz. The attenuation coefficient a of radar wave amplitudes is expressed in dB/m.

(iii) Radar wave velocity

Although difficult to measure accurately, it has been found that radar waves show considerable differences in propagation velocities for different earth materials. To present these high velocities in reasonable figures, they are presented in centimetres per nanosecond (cm/ns). In free air this velocity is 30 cm/ns, in earth materials it can range from 15 to 3 cm/ns. The variation in velocity makes it possible to apply seismic reflection techniques to obtain a radar section. Applying the Common Mid Point technique, but in a single coverage mode, results in the registration of hyperbolic reflectors. Using processing techniques similar to those used in seismics, wave velocities can be derived and NMO corrections can be applied to obtain the section. Once the wave velocities are known time sections can be converted into actual depth sections.

Figure 6.26 presents an example of a radar section obtained over a sandy formation (push moraine) in the Netherlands (Van Overmeeren, 1994). The section clearly registers steps in the water table attributable to small faults.

Figure 6.26 Example of a GPR section of push moraine deposits, showing groundwater steps at 286 m and at around 330 m (Overmeeren, 1994)



6.9 Geophysical borehole logging

6.9.1 Principles and instrumentation

As long ago as the 1920s the petroleum exploration industry developed techniques to perform geophysical observations in boreholes. These techniques were called *wireline logging* after the principle of lowering a measuring probe into the borehole, suspended from a cable, but they are generally referred to (geophysical) borehole logging or well logging.

Logging is applied in geohydrology to accurately determine the sequence and thickness of rock formations, their physical properties, and to establish porosity, permeability, water flow and groundwater salinity (Repsold, 1989). In addition, it is used to check and control the interpretation of geophysical data recorded from the surface. Typical parameters, measured in boreholes are: Resistance, Spontaneous Potential (SP), Formation Resistivity (in various modes), Formation Conductivity (Induction Logging), Fluid Conductivity, Nuclear Radiation (in various modes), Formation Temperature, Seismic Wave Velocity (sonic, acoustic), Borehole Diameter (calliper) and Formation Fluid Flow.

(i) Instruments and tools

The set up of borehole logging equipment is shown in Figure 6.27. The basic components are the *measuring probe*, which is suspended in the borehole by the *logging cable*, which is wound onto a *winch*. The logging cable is connected to the actual measuring instrument, which controls the logging speed and direction, signal levels and calibration, and supplies downhill power to the probe, if needed. The measuring instrument also contains a paper and pen recorder, which registers the signal level of the probe and continuously records the depth of the probe. Modern instruments are usually fitted with a digital recording systems, which allows the data to be redisplayed and processed in the office. Modern winches are driven electrically and consequently need a generator to supply the power.

Normally there is a specific probe for every parameter. Therefore the probe can easily be removed from the cable head and exchanged for the appropriate one. For some parameters it is possible to combine several tasks into one probe.

The logging cable is made of steel wiring on the outside to take the weight of the cable and probe in the borehole and has a core of electrical leads to transmit different signals, but also to supply electric power to the probe. The logging cable is partly or completely covered by an insulating sheath, to eliminate unwanted effects on electrical measurements. At the surface a *measuring sheave* records the displacement of the cable via a stepmotor or an optical-electrical device. The displacement information is linked to the recording instrument.

The recorder is fitted with a number of measuring modules, which can be switched on to perform their task. Every module has its specific controls for calibration, signal zeroing, signal level (recorder deflection) sensitivity and time constant. Some logging instruments particularly in use in geohydrology are made very light and compact, to make them portable. Measuring modules can then easily be removed and replaced, thus saving space and weight.

What is important in logging applications is that many logging techniques can only be used in open, uncased boreholes. Most of the electrical techniques fall in this category, as well as the calliper log and the sonic log. Other techniques, such as the nuclear logging, temperature and fluid conductivity logging and flow meter measurements, can be recorded in completed (cased) wells.

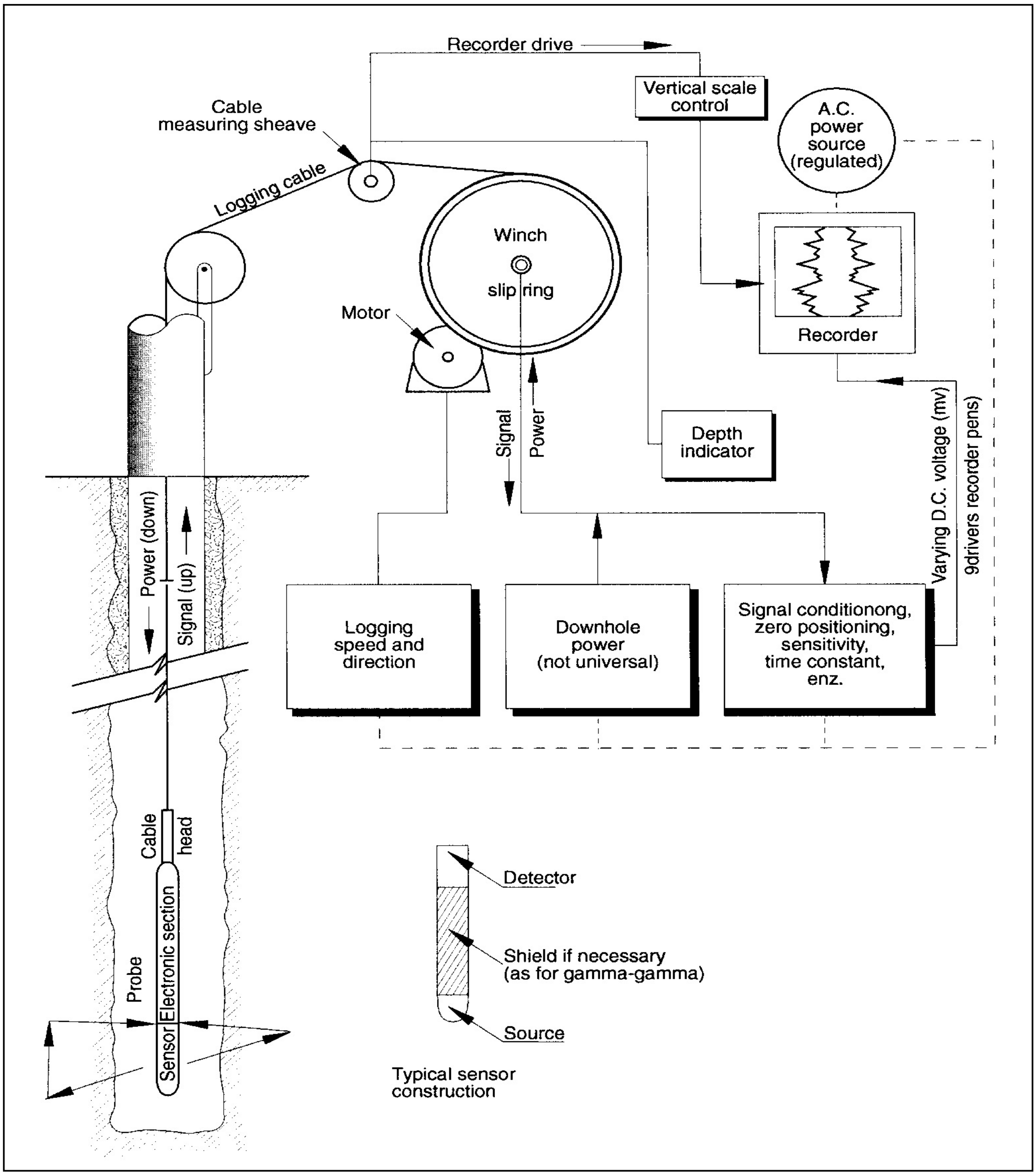
6.9.2 Logging physical parameters

The most common methods in borehole logging are the electrical methods. The probes to be used are rather simple and inexpensive, with the exception of *Induction Logging*. Frequently used in combination with the electrical methods are *Natural Gamma Ray* logging and *Calliper* logging.

(i) Electrical logging

Spontaneous Potential or SP is a parameter generated by the presence of drilling liquid, commonly a mixture of mud and water. The SP is caused by two effects, an electrochemical potential and a

Figure 6.27 Basic components of geophysical borehole logging (Keys and MacCary, 1971)



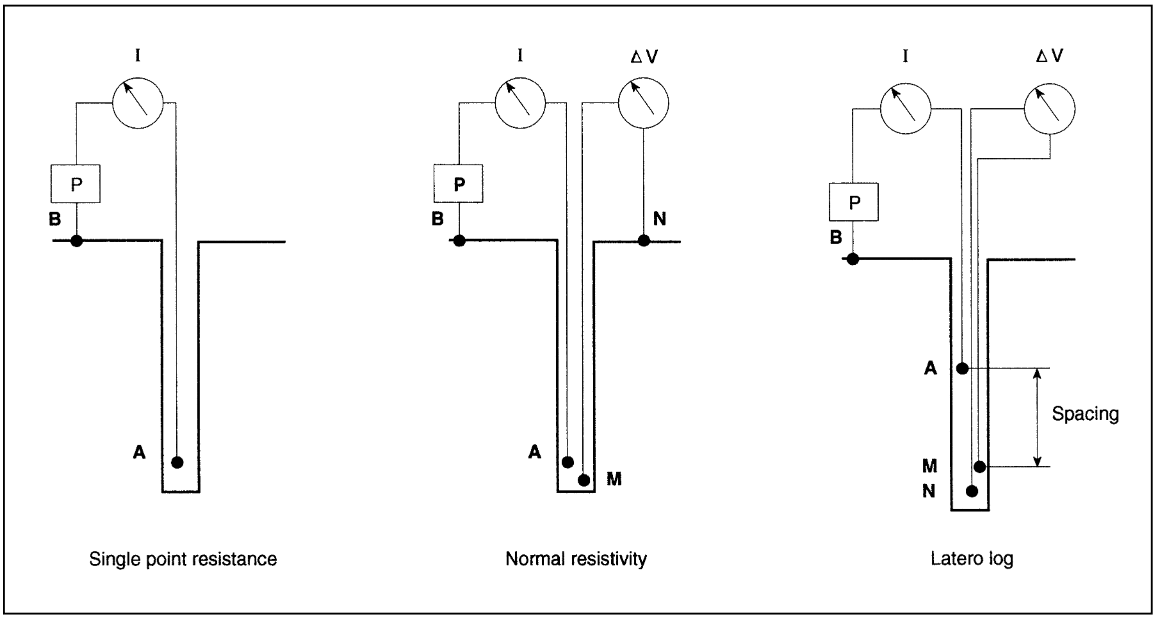
streaming (electrokinetic) potential. The *streaming potential* is generated by the flow of water with ions dissolved, through the narrow pores of a rock formation and can have a maximum value in the order of 10 to 15 mV. The streaming potential across the mudcake on a sand layer is generally negative. The *electrochemical potential* is generated when two fluids with different salt concentrations are in contact with each other, either directly or via a semi-permeable membrane. Lynch (1962) gives a detailed description of the mechanism. The electrochemical potential can be negative or positive, depending on the salinity of the formation fluids.

The SP voltage is a relative value, whose sign depends on the formation fluids, the drilling mud and the nature of the drilled formation, i.e. sand or clay (shale). In the interpretation, the SP level at clay (or shale) formations is taken as a baseline. The SP will be relatively positive across fresh water-bearing sands. The response may shift to a relatively negative deflection across sand containing saline water. In a borehole the baseline can shift in position due to variations in (clay) lithology and groundwater composition. As a consequence, SP measurements are only suitable for qualitative interpretation. Generally it is useful to determine layer boundaries, to identify thin beds, lithology and a quantitative indication of saline groundwater.

(ii) Resistivity logging

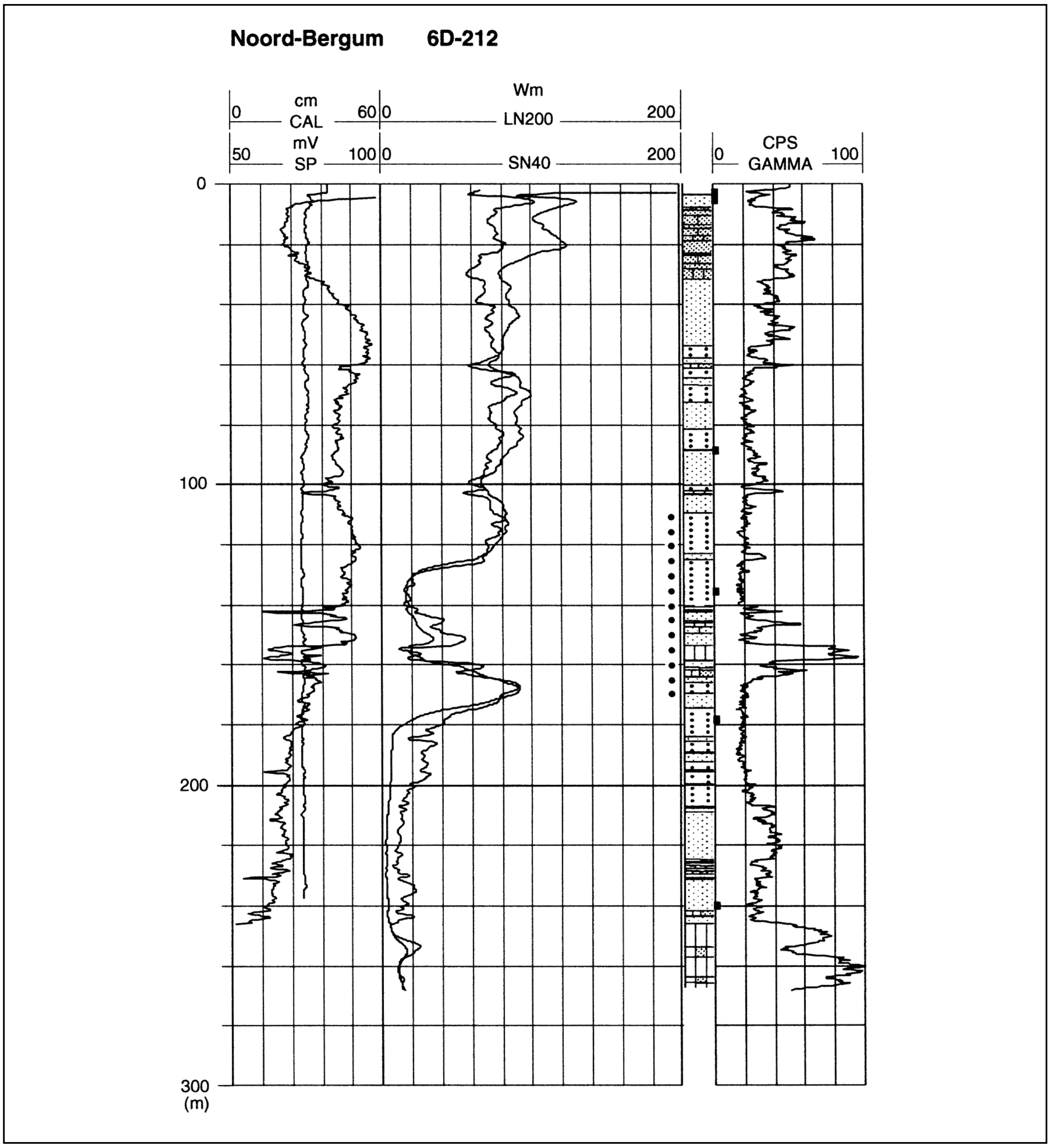
Resistivity logging makes use of a similar arrangement of electrodes as in surface resistivity surveying. One current electrode (B) and one potential electrode (N) are placed at the surface, while the others (A and M) are positioned on the probe in the borehole. A potential field is generated between A and B, while M measures the potential field around A (Figure 6.28). This technique is also referred to as the Normal Device. The resistivity is calculated from : $r = K\Delta V/I$, with the geometrical factor $K = 4\rho AM$.

Figure 6.28 Electrode configurations for single point resistance logging, normal resistivity logging and latero logging. The single point resistance configuration can also be used for SP logging, when no power is supplied to run a current through the formations



The depth investigation of a Normal Device depends on the electrode separation AM. A small separation results in a shallow penetration (approximately 2AM), which is consequently strongly affected by drilling mud in the borehole and by drilling fluid which has penetrated into the borehole wall. A small separation also results in a better resolution of thin beds. A large electrode separation suffers less from the borehole effects and therefore gives a better depiction of the true formation resistivity ρ_f of the formation, with the disadvantage of lesser resolution for thin beds. This entails fitting a resistivity probe with two potential electrodes M_1 and M_2. Records with both a large and a small separation are called *Long Normal* and *Short Normal* resistivities (LN and SN). Figure 6.29 shows an example of SN and LN measurements. Typically a ratio of four is used for the LN and SN separations. In geohydrological applications, separations are often 25 and 100 cm or 50 and 200 cm.

Figure 6.29 Example of a borehole log as used in groundwater surveys. The left graph shows the calliper (borehole diameter) and SP log. The middle graph shows the long normal (LN) and short normal resistivity logs, while the right graph represents the natural gamma radiation log with a lithological interpretation. The resistivity logs show fresh water formations below ca. 130 m. Below this depth the salinity increases until a depth of 180 m, to remain constant at larger depths. The gamma log shows a high level between 155 and 160 m, correlating with resistivity (and SP) lows; this indicates the presence of a clay layer (Courtesy TNO-Applied Geosciences, presently Netherlands Institute for Applied Geoscience NITG)



Since the measurements are affected by the drilling mud, borehole diameter and the invasion zone they must be corrected for these effects so that true formation resistivities can be determined. To this end correction diagrams have been prepared, or computer techniques can be applied.

(iii) Variations on the resistivity probe

The drive to obtain a more reliable value for the formation resistivity resulted in the development of the *Laterolog Device*. In this variation, both M and N electrodes are placed on the probe, with the A current electrode relatively far away (18 feet) from the midpoint of M and N. This distance is called the *spacing*. Because of this array, the current is forced deeper into the formations and the investigation depth is much greater than with a Normal Device.

(iv) Microresistivity Log

This tool, called a *Dipmeter* was developed to investigate the shallow invasion zone of the borehole wall. It has all four electrodes closely spaced on a pad, which is pressed against the borehole wall for good physical contact, but also to limit disturbing effects from the environment. Because of the small electrode separation the tool can also register very thin beds within the formation. The probe can also combine three or four pads, mounted at angles of 120° or 90°. The local dip of the thin layers can be established by correlation of the Microresistivity curves.

A variation on the microresistivity device is the *Microlatero Log*. It combines the specifications for the microlog with the laterolog.

(*v*) *Single Point Resistance*

This is a simplified version of resistivity logging. It uses one electrode at the surface and one in the borehole (Fig. 6.28). Both act as current and as potential electrodes at the same time. The recorded resistance response is not a linear one, therefore the data cannot be used for *quantitative* interpretation. Still, the results are useful for qualitative interpretation, and very resistive, thin layers can be detected very well without the problem of response inversion, as is the case in normal logging. The technique requires relatively cheap instrumentation and a single core cable and is often found in small and basic logging units.

(vi) Induction logging

This is an electric method, based on the electromagnetic principles as used in surface EM methods. Transmitter and receiver coil are mounted in the probe. This tool was developed in order to log in non-conductive drilling fluids (oil and gas). Typically this tool has a large potential for use in geohydrology to log in boreholes above a low water table, since it does not require physical contact between the probe and the borehole wall. It can also be used in PVC cased boreholes. At present, induction loggers specifically designed for use in groundwater exploration are available.

(vii) Calliper logging

As discussed before, it is often necessary to correct recorded data for the borehole diameter effect. Irregularities in borehole diameter can be caused by material from the borehole wall washing out during drilling, and by the drilling system. The well diameter can be measured directly, by three or four ‘feelers’ attached to the probe. The expansion of these feelers is ruled by the borehole diameter and translated into a signal for the recorder. Results of a calliper measurement are shown in Figure 6.29.

(viii) Nuclear logging

Nuclear logging concerns a group of methods, based on recording the natural radiation of formations or the effect of formations on radiation transmitted from a source in the probe.

Natural Gamma Radiation logging is based on the emission of gamma radiation (photons) by radioactive minerals, such as uranium, thorium and the K^{40} isotope of potassium. As well as to gamma radiation, alpha and beta radiation are also generated, but these do not have sufficient power to penetrate the ground. K^{40} is a very common natural component in earth materials and occurs more frequently in clays and shales than in sands or chalks and dolomites. Therefore the level of radiation is an indication of the type of lithology. The gamma radiation is measured with a *scintillometer*, often containing an NaI crystal, which emits light when hit by a photon (scintillation). The emitted light is converted by the photocathode of a photomultiplier tube into emission of electrons. The photomultiplier amplifies the electron flow to a recordable level. The gamma radiation is a statistical process, which means that the intensity varies with time. To compensate for the irregular emission, the radiation counts are averaged over a time interval, the *time constant* TC. Usually the TC is in the order of 2 to 10 seconds. The time constant is set by the operator. A small TC will show larger statistical variations, i.e. the repeatability of a gamma log will be small. A large TC will have less statistical variation, but if the logging speed of the tool is too high, it may not produce the true radiation level of a particular formation. It is clear that thin layers would then not be detected through their radiation level. Logging speed in gamma radiation logging is typically of the order of 1 to 2 metres per minute. Calibration of a gamma radiation tool opens up ways of quantitatively interpreting geohydrological parameters.

Gamma – Gamma logging or *Density Logging* uses a gamma radiation source, which is built into the probe. The emitted radiation penetrates into the surrounding formation and is scattered by it and partly absorbed. The backscattered radiation is recorded by the probe. The degree of absorption is related to the density of the formation. Thus the gamma-gamma log represents variation of density with depth. This tool is particularly useful for identifying relatively dense anhydrite and barite. Again the method can be used for quantitative analysis if the tool is calibrated.

Neutron logging includes two variations: the *Neutron-Gamma log* and the *Neutron-Neutron log*. The neutron-gamma log uses a 'fast' neutron source in the probe, which radiates thermal neutrons (with high energy level) into the formation. When the neutrons collide with nuclei in the rock, gamma radiation is emitted, which in turn is recorded by the probe. This process is particularly sensitive to the presence of hydrogen in the formation, i.e. water and hydrocarbons. It is therefore an effective tool to determine the relative amount of pore water in a formation. The neutron-neutron log also uses a neutron source in the probe, but records the remaining (backscattered) neutron with a neutron counter. The physical effect of hydrogen is the same and the curves obtained are similar. Calibrated tools allow for quantitative analysis of porosity in the saturated zone. Above the water table the moisture content can be determined.

Logging techniques using radiation sources are less commonly used, since the use of these instruments requires more skilled operators than the other techniques. Protection personnel against health hazards and permits for storage and transport of equipment require higher expenditure than for 'normal' equipment.

(ix) Sonic logging, or *velocity logging*

This method is used to measure acoustic wave velocity in the formation around the borehole. The velocity is related to density and the elastic properties of the rock. Measurements are made with a probe, which contains a wave transmitter and two or more receivers placed at different distances from the source. The tool records the travel time of the sonic wave between transmitter and receiver. The transmitter and receivers are acoustically insulated, to prevent the signal from travelling directly through the tool itself. The probe is centred in the borehole, which means that the wave travels partly through the drilling mud. The distance travelled in the borehole is eliminated by the use of at least two receivers. The travel time is usually recorded in msec/foot. Velocity calculations are based on first arrival recording of the compressional wave. The porosity of a formation can be calculated from the empirical *time-average equation*, developed by Wyllie, Gregory and Gardner (1958). These porosity calculations are useful in consolidated rocks, but for unconsolidated formations it is very difficult to make reliable calculations. A compaction factor $1/C_p$ must then be applied to the above expression.

Generally not much use is made of the density and sonic logs in geohydrology, since the logging is rather expensive and the processing is based on computer applications. The sonic logs however, provide additional information for reflection seismic interpretation. Density and velocity parameters can be used to prepare synthetic seismograms, which are used for correlation with the processed field data. The nuclear methods can be used in cased and open boreholes.

(x) Flow meter measurements

Flow meter measurements are made with a probe in the cased borehole to measure the water flow from aquifer formations during production. In principle an impeller is used to monitor the flow, but more sophisticated techniques based on mass movement are also used. This yields an impression of the permeability of different parts of the formation.

(xi) Fluid conductivity

Fluid conductivity can be measured with a water conductivity meter attached to the logging cable. Very often it is combined with the flow meter. It presents a continuous water conductivity profile of the drilled formation.

(xii) Temperature logging

Temperature logging results in a temperature profile of the borehole, in which anomalous high or low segments are indicative of the origin of inflowing water. This recording is usually made after one or two weeks after completion of the well, in order to prevent effects of the drilling activities on the natural thermal balance of the location.

6.9.3 Data processing and interpretation

The main value of borehole logging lies in correlation. Logs from different boreholes in a survey area or wellfield are carefully checked for similarity in response, which can lead to absolute correlation or to a correlation of a certain sequence of layers. Invaluable stratigraphic insight is obtained in this way

In the previous sections it was already mentioned that much of the borehole log information can be used to do a quantitative interpretation, allowing true formation resistivity, porosity, density and saturation to be established. Since the boreholes are logged almost continuously, the enormous amount of data calls for computer processing. In this aspect, the applications in geohydrology are approaching those of petrophysics as applied in oil exploration.

Most important in data processing and interpretation is the correlation of the physical and geophysical parameters with the hydrological parameters of an aquifer. Potentially this allows detail in modelling an aquifer to be improved, which in turn may lead to better groundwater management.

6.10 References and additional reading

ALVAREZ, R., 1991. Geophysical Determination of Buried Geological Structures and Their Influence on Aquifer Characteristics. *Geoexploration,* Vol. 27, pp. 1–24.

ANDERSON, W. L., 1979. Numerical Integration of Related Hankel Transforms of Orders 0 and 1 by Adaptive Digital Filtering. *Geophysics,* Vol. 44, pp. 1287–305.

ANGELITO, V.; CERVERA, G.; CHAPELLIER, D., 1991.La gravimétrie expéditive appliquée à la recherche d'aquifères en zone aride; cas de la nappe alluviale du Teloua (Agadez, Niger). *Geoexploration,* Vol. 27, pp. 179–92.

ARCHIE, G. E., 1942. The Electrical Resistivity Log an Aid in Determining Some Reservoir Characteristics. *Trans. A.I.M.E.,* Vol. 146, pp. 54–64.

ASTIER, J. L.; PATERSON, N. R., 1989. Hydrogeological Interest of Aeromagnetic Maps in Crystalline and Metamorphic Areas. Paper 59. In: Garland, G. D. (ed.), *Proc. Exploration '87: Third Decennial International Conference on Geophysical and Geochemical Exploration for Minerals and Groundwater, Ontario Geological Survey, Special Vol. 3,* 960 pp.

BARKER, R. D., 1981. The Offset System of Electrical Resistivity Sounding and Its Use with a Multicore Cable. *Geophysical Prospecting,* Vol. 29, pp. 128–44.

BOURGEOIS, B.; MATHIEU, F.; VACHETTE, C.; VAUBURG, P., 1994. AMT Measurements Compared with Gravity and Magnetometry for Structural Study of a Sedimentary Basin. Letthakeng-Bothlapatlou Groundwater Project, Botswana. *J. Appl. Geophysics,* Vol. 31, pp. 7–25.

BROMLEY, J.; MANNSTRÖM, B.; NISCA, D.; JAMTLID, A., 1994. Airborne Geophysics – Application to a Ground-Water Study in Botswana. *Groundwater,* Vol. 32, pp. 79–90.

BUTLER, D. K., 1984. Microgravimetric and Gravity Gradient Techniques for Detection of Subsurface Cavities. *Geophysics,* Vol. 49, pp. 1084–96.

CARMICHAEL, R. S.; HENRY, G., 1977. Gravity Exploration for Groundwater and Bedrock Topography in Glaciated Areas. *Geophysics,* Vol. 42, pp. 850–9.

FRISCHKNECHT, F. C., 1971. Electromagnetic Scale Modelling. In: J. R. Wait (ed.), *Electromagnetic Probing in Geophysics.* The Golem Press, Boulder CO, pp. 265–320.

GHAZALA, H. H., 1994. Structural Interpretation of Bouguer and Aeromagnetic Anomalies in Central Sinai. *J. African Earth Sci.,* No. 19, pp. 35–22.

HAGEDOORN, J. G., 1959. The Plus-Minus Method of Interpreting Seismic Refraction Results. *Geophysical Prospecting,* Vol. 7, pp. 158–82.

HAWKINS, S. L. V., 1961. The Reciprocal Method of Routine Seismic Refraction Investigations. *Geophysics,* Vol. 26, pp. 806–19.

HUNTER, J. A.; PULLAN, S. E.; BURNS, R. A.; GAGNE, R. M.; GOOD, R. L., 1984. Shallow Reflection Seismic Mapping of the Overburden-Bedrock Interface with the Engineering Seismograph – Some Simple Techniques. *Geophysics,* Vol. 49, pp. 1381–5.

KELLER, G. V.; FRISCHKNECHT, F. C., 1966. *Electrical Methods in Geophysical Prospecting.* Pergamon Press.

KEYS, W. S.; MACCARY, L. M., 1971. Application of Borehole Geophysics to Water Resources Investigations. *Techniques of Water Resources Investigations of the US Geological Survey, Book 2,* Chapter E1. Washington, D.C., United States Government Printing Office.

KOSINSKY, W.; KELLY, W. E., 1981. Geoelectric Sounding for Predicting Aquifer Properties. *Groundwater,* Vol. 19.

LYNCH, E. J., 1962. *Formation Evaluation*. Harper and Row, New York.

MAZAC, O.; KELLY, W. E.; LANDA, I., 1985. A Hydro-Geophysical Model for Relations between Electrical and Hydraulic Properties of the Aquifers. *J. Hydrol.,* Vol. 79, pp. 1–19.

MCNEILL, J. D., 1990. Use of the Electromagnetic Methods for Groundwater Studies. In: S. H. Ward (ed.), *Geotechnical and Environmental Geophysics, Vol. 1: Review and Tutorial.* Tulsa, Society of Exploration Geophysicists, pp. 191–218.

MEEKES, J. A. C.; SCHEFFERS, B. C.; RIDDER, J., 1990, Optimization of High-Resolution Seismic Reflection Parameters for Hydrogeological Investigations in The Netherlands. *First Break,* Vol. 8, pp. 263–70.

MILSOM, J., 1989. *Field Geophysics*. Geological Society of London Handbook. Publ. Open University Press.

PALMER, D., 1980. The Generalized Reciprocal Method of Seismic Refraction Interpretation. In: K. B. S. Burke (ed.), Soc. Explor. Geophs., Tulsa.

PELLERIN, L.; LABSON, V. F.; PFEIFFER, M. C., 1994. VETEM – A Very Early Time Electromagnetic System (abstract). *The John S. Sumner Memorial International Workshop on Induced Polarization in Mining and the Environment,* Tucson, Arizona.

REEVES, C. V., 1992. New Horizons in Airborne Geophysical Mapping. *Exploration Geophysics,* Vol. 57, pp. 956–9.

REPSOLD, H., 1989. Well Logging in Groundwater Development. *IAH Contrib. to Hydrogeol.,* Vol. 9, Heise, Hannover, 136 pp.

SENGPIEL, K. P., 1986. Groundwater Prospecting by Multifrequency Airborne Electromagnetic Techniques. In: G. J. Palacky (ed.), *Airborne Resistivity Mapping, Geological Survey of Canada Paper 86–22,* pp. 131–8.

SPORRY, R. J.; ARZATE, J. A.; YOUSSIF, M. S., 1991. Geophysical methods in groundwater modelling. *Proc. A. E. G. Seminar,* Vol. 2., Hyderabad.

SHERIFF, R. E.; GELDART, L. P., 1982. *Exploration Seismology, Vol. 1: History, Theory and Data Acquisition; Vol. 2: Data-processing and Interpretation*. Cambridge University Press, Cambridge.

STEEPLES, D. W.; MILLER, R. D., 1990. Seismic Reflection Methods Applied to Engineering, Environmental and Groundwater Problems, In: Stan Ward (ed.), Soc. Explor. Geophs., *Investigations in Geophysics,* No. 5, Volume on Environmental Geophysics, pp. 1–30.

SUMNER, J. S., 1976. *Principles of Induced Polarization for Geophysical Exploration*. Elsevier Science Publ. Co., 277 pp.

TELFORD, W. M.; GELDART, L. P.; SHERIFF, R. E., 1990. *Applied Geophysics*. Second Edition, Cambridge University Press.

VAN DER VELPEN; SPORRY, R. J., 1993. RESIST – A Computer Program to Process Resistivity Sounding Data on PC-Compatibles. *Computers and Geosciences,* Vol. 19, pp. 691–703.

VAN OVERMEEREN, R. A., 1981. A Combination of Electrical Resistivity Seismic Refraction and Gravity Measurements for Groundwater Exploration in Sudan. *Geophysics,* Vol. 46, pp. 1304–13.

——, 1994. Georadar for Hydrogeology. *First Break,* Vol. 12, pp. 401–8.

VAN OVERMEEREN, R. A.; RITSEMA, I. L., 1988. Continuous Vertical Electric Sounding. *First Break,* Vol. 6, pp. 313–24.

WYLLIE, M. R. J., GREGORY, A. R.; GARDNER, G. H. F., 1958. An Experimental Investigation of Factors Affecting Elastic Wave Velocities in Porous Media. *Geophysics,* Vol. 23, pp. 459–93.

YILMAZ, Ö., 1987. *Seismic Data Processing*. Soc. Expl. Geophys., Tulsa.

7 Well drilling and design methods

7.1 Introduction

The first use of groundwater as a source of water supply is lost in antiquity. In fact, as soon as the primitive man learned to domesticate and rear cattle and sheep in the valleys of the Tigris-Euphrates, Yellow, and Indus rivers (10000 BC) the well became his most important possession. Since that time, virtually all homes with their own water supply have wells and use groundwater. Furthermore, groundwater is increasing in importance in world water supplies, in part as a response to the growing costs and other constraints in storing and treating surface water and partly because the advantages of groundwater are now better understood. Today, more than half of the world's population depends on groundwater for survival (UNESCO, 1992).

The use of wells for domestic purposes (human and animal consumption) usually has the highest priority, followed by industrial requirements, and finally agricultural use. Today, wells provide 75 percent of drinking water supplies in Europe and more than 50 percent in the United States (Cohen, 1985; Driscoll, 1989; Rebouças, 1976; Solley et al., 1993). During the last 25 years about 300 million wells have been drilled for groundwater withdrawal all over the world, and in the United States of America alone, 100 million are currently in use (Carpenter, 1994; Fetter, 1993; Rebouças, 1991, Solley et al., 1990; UNESCO, 1992). Most of these are generally 150 mm (6 in) in diameter and less than 150 m deep; only a small proportion are over 250 m in depth. However, in the large sedimentary basins there are many wells deeper than 500 m that reach the confined aquifers; e.g. wells in the huge sedimentary basins, such as Great Australian Artesian Basin, Paraná and Maranão Basins in Brazil (some of more than 2000 m depth and with diameters of the pump chamber casing of 762 mm) yielding from tens to hundreds of thousands of cubic metres a day (Habermehl, 1985; Margat, 1990; Rebouças, 1988; Rebouças, 1991).

Discussion in this chapter is limited to drilled vertical wells for producing or monitoring groundwater. Because it is not possible to describe in detail the numerous drilling and well construction techniques, the reader is urged to consult the major sources cited in this chapter.

7.2 The development of well drilling techniques

Shallow, hand-dug wells and crude water-lifting devices marked the early exploitation of groundwater. The oldest dug-wells known were found in the Middle East and are dated from 8000 BC. The art of drilling wells by lifting and dropping a string of tools suspended on a cable was invented, perfected, and extensively practised by the ancient Chinese 5000 before present. The early Chinese could drill wells to depths ranging between 1,200 and 1,500 metres (Hardcastle, 1987; Meinzer, 1934; Tolman, 1937).

Percussion methods of well drilling were developed much later in Greece and Rome (500 BC–AD 500). Medieval advances (AD 500–1500) were made in Western Europe, largely after the discovery of flowing wells, first in France about 1100 AD, and a few decades later in eastern England and northern Italy. The wells in the region of Artois, discovered in 1126, became so famous that flowing wells are often called artesian wells after the name of the region (Tolman , 1937).

The emergence of Western technology (1500–1750) and the birth of steam power, rapidly improved the development of the cable tool method of drilling, often referred to as the standard method, or percussion method. The Industrial Revolution (1750–1900) brought advances in power technology, such as development of steam power, internal combustion engine, and electric power, which induced the most impressive development of the cable tool method of drilling, as a result in part to knowledge borrowed from the oil and gas industry (Bowman, 1911; Campbell and Lehr, 1973; Driscoll, 1989; Hardcastle, 1987; Lehr et al., 1987; Meinzer, 1934).

The hydraulic rotary method of well drilling was developed in the oil fields of Louisiana (United States of America) about 1890. Since that time it has replaced the cable-tool method of oil well drilling, and even of water well drilling to a great extent (Bowman, 1911; Campbell and Lehr, 1973; Driscoll, 1989; Gordon, 1958; Speedstar Division of Koering, 1967). Later, the development of the combined hydraulic rotary-percussion methods of well drilling and of the electric-driven pumps made possible the recovery of groundwater in large amounts (up to 1,000 m^3 per hour), and at increased depths (up to more than 2,000 metres). At the present time, there are no technological limitations to reaching the deeper confined aquifers in all over the world (Anderson, 1992; Driscoll, 1989; Lehr et al., 1987; Rebouças, 1994; Solly et al., 1993).

If in the early times of civilisation the well was the base for development in the arid and semi-arid regions, recently groundwater has become an important source of water supply for domestic, industrial and agricultural purposes in temperate and even in the humid tropical climatic regions, in view of its long range economic feasibility. Currently, most of the drilled wells are generally 15 cm in diameter and less than 150 m deep; only a small proportion are over 200 m in depth, but in the large sedimentary basins there are many wells deeper than 1,000 m that reach the best aquifers (Driscoll, 1989; Margat, 1990; Rebouças, 1991; 1994, Smith, 1980).

7.3 Water quality protection for wells

In the past, the purity and sanitary quality of groundwater was assumed, and even when groundwater resources were used for drinking water supplies, little or no treatment was thought to be required. But in recent years, there is growing evidence that this resource is being contaminated locally. The main pollution sources are municipal and industrial wastes, sewage treatment and disposal, spills and leaks from storage and transport of liquids, well injection of liquid wastes, agricultural activities, and mining activities.

Groundwater contamination is so severe in certain localities that continued use of the water could lead to serious health problems. Protection of water quality partly depends on the well design, the drilling methods and the materials used to construct the production and monitoring wells. When those wells are badly designed, constructed, and left uncontrolled, they invite groundwater quality degradation (Fetter, 1993; Rebouças, 1991; EPA, 1975, 1980). As a result, more than ever, the drilling contractor is faced with the challenge of complying with new legislation that addresses groundwater quality and the growing number of groundwater contamination problems. These challenges and problems can be handled satisfactory only through greater understanding of the water well construction standards for all new and reconstructed wells intended or used for supplying water for human consumption, preparation of food products, and monitoring groundwater quality.

Collector wells and infiltration galleries are sometimes constructed where special hydrogeological conditions exist (Wright et al., 1987).

7.4 Standards for water and monitoring wells

Many countries, states and organisations of drilling contractors have established minimum standards covering such aspects of water well construction as: location with respect to possible pollution sources, casing types and weights, sanitary seals, the protection of aquifers, and the sterilisation of wells (ASTM, 1992; Driscoll, 1989; Fetter, 1993; Lehr et al., 1987; OECD, 1989; EPA, 1975).

Water well standards should assure that:

- the completed well will be constructed of material that will be compatible with the environment and will give an adequate well life time,
- workmanship will be of an acceptable standard,
- the well will be of adequate size and design to yield the desired discharge with maximum efficiency.

Standards and well specifications should:

- be reasonably flexible,
- foster competitive bidding,
- permit the preparation of cost estimates,
- be a guide to those responsible for supervising the construction of the well.

For monitoring work, many of the objectives of a drilling programme are similar to those for a water well, but some of the construction steps must be made with greater care, to ensure that the water quality is protected and reliable water samples can be obtained (Aller et al., 1989; Driscoll, 1989; Fetter, 1993; EPA, 1975; USDI, 1981).

Groundwater protection policy and strategy should be based on the concept that prevention of pollution is always less expensive than aquifer rehabilitation, which is a costly, time consuming and technically demanding task.

7.5 Well drilling techniques

Shallow low-capacity production wells are usually dug, driven, or jetted, and high-capacity production wells are usually drilled by cable-tool rigs, direct circulation rotary rigs and variations thereof such as reverse circulation rotary drilling, air drilling systems, and inverse drilling methods. In hard rocks down-the-hole-hammer techniques – a combination of percussion and rotary drilling – are used.

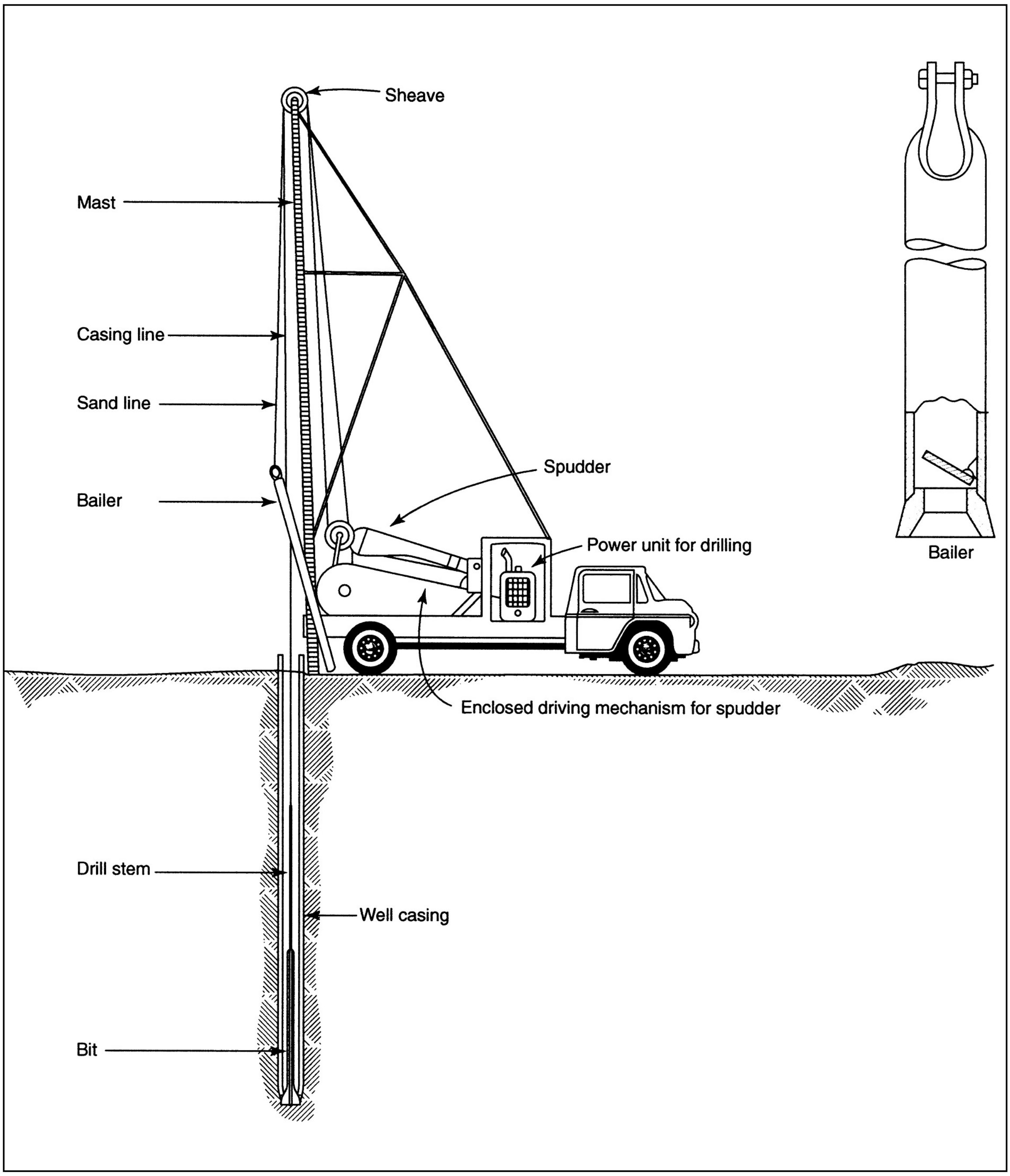
Numerous drilling techniques have been developed to cope with the wide range of geological conditions, from hard rock such as granite and dolomite to completely unconsolidated sediments such as alluvial sand and gravel. Moreover, the drilling contractor may vary the usual drilling procedure depending on the depth and diameter of the well, type of geological formation to be penetrated, sanitation requirements, and principal use of the well, such as production, recharge or monitoring.

In this chapter only the most frequently used well drilling techniques will be described.

7.5.1 Cable tool method and variations

The cable tool method of drilling, often referred to as the standard method, or percussion method, is one of the oldest and simplest drilling methods (Figure 7.1). The cable tool

Figure 7.1 Typical truck-mounted cable tool equipment for drilling wells. Regular bailer with flat valve bottom (After Driscoll, 1989)



equipment is probably the most versatile of all rigs in its ability to drill satisfactorily under a wide range of environmental and hydrogeological conditions. It may be the best, and in some cases the only, method to use in coarse glacial till, boulder deposits, or rock strata that are highly disturbed, broken, fissured, or cavernous.

The cable tool machine is usually compact, and powered with a diesel engine placed at the front end of the frame. It can be easily moved in rough terrain. The simplicity of design, ruggedness, and easy maintenance of the rig and tools are particularly advantageous in isolated areas. Moreover, it generally requires less skilled operators and a smaller crew than other rigs. The low horsepower requirements are reflected in lower fuel consumption, an important aspect where fuel costs are high or sources of fuel are remote. Cable tool rigs are generally limited to drilling maximum hole diameters of 600 to 750 mm (24 to 30 inches) and to depths of less than 600 m (2,000 ft).

The hole is drilled by raising and lowering a heavy bit on the end of a steel cable which is threaded over sheaves at the top of the mast and down to the drill-line drum. For effective penetration the drilling motion of the cable tool machine must be synchronised with the gravity fall of the full string of cable tools. The drill bit breaks consolidated rock into small fragments, but when drilling in unconsolidated sediments it primarily releases the material. The broken and crushed material in the bottom of the hole is removed by means of a bailer (Figure 7.1).

In stable rock, an open hole can be drilled, but in unconsolidated formations, casing must be driven down the hole during the drilling. Above the water table or in otherwise dry formations, water is added to the hole to form a slurry of the cuttings so they may be readily removed by a bailer. However, much less water is required for drilling than with most other commonly used rigs, an important consideration in arid and semi-arid zones. Also, sampling and geological logging are simpler and more accurate with the cable tool rig. Table 7.1 presents the major advantages and disadvantages of the cable tool method (Anderson, 192; Bowman, 1911; Campbel and Lehr, 1973; DAO, 1965; Driscoll, 1989; Lehr et al., 1987; NWWA, 1981, 1985; Roscoe, 1985; USDI, 1981).

Table 7.1 Advantages and disadvantages of cable tool method

Advantages	*Disadvantages*
1. Rigs are relatively cheap, require little maintenance, are readily moved in rugged terrain or where space is limited.	1. Relatively slow rate of progress compared to other rigs of similar capacity.
2. Machines have low energy requirements and can drill in areas where little make-up water is available.	2. Economic and physical limitations to depth and diameter.
3. Reliable for a wide variety of geological conditions, and the samples are not contaminated by drilling mud. of pulling back long strings of casing.	3. Necessity of driving casing in unconsolidated materials and difficulty
4. Generally require less skilled operators and smaller crew than other rigs of similar capacity.	4. Each cable tool driller can complete only a limited number of holes per year.

The major disadvantage, compared to other types of rigs, is its lower rate of progress and its depth limitations. A further disadvantage of the cable tool rig is the necessity of driving casing concomitant with drilling in unconsolidated materials. This precludes the geophysical well-logging, although the logging is desirable in many instances. However, gamma logs may be taken inside the casing. The driving of casing necessitates a heavier wall pipe than would otherwise be required in some installations. Screens often must be set by pullback or bail-down methods. The pullback method in deep or large diameter wells is sometimes extremely difficult, and the bail-down method may give rise to problems of alignment (AWWA, 1984; Bowman, 1911; Driscoll, 1989; Gordon, 1985; Lehr et al., 1987; NWPA, 1981; OECD, 1989; USDI, 1981).

Every cable tool machine has certain interdependent limits to borehole depth and diameter. For example, if a hole is relatively small in diameter, it may be drilled to relatively great depth. In larger-diameter holes, the weight of the drill string and cable may become so excessive that the machine cannot function, thereby limiting well depth to a smaller diameter.

A variation of the cable tool drilling technique, called the full-hole cable tool method, has been used in Brazil since the 1960's. With the open hole full of water or drilling fluid, the string of cable tools is lifted and dropped to cut the borehole. In this case the hole can be drilled open-hole, even in completely soft or unconsolidated sediments, because the hydrostatic water or drilling fluid pressure prevents the borehole walls from caving in. Wells between 150 to 250 mm (6 to 10 inches) in diameter have been drilled in this manner to depths of 100 m in surficial aquifers which occur in Tertiary unconsolidated deposits, in crystalline rocks with deeply weathered mantle or detrital cover, and in shallow confined or semi-confined sandy aquifers (Rebouças, 1988).

Another cable tool drilling technique, used for many years in Japan and more recently in the Western United State, is called the open-hole or reverse cable tool method. With the borehole full of water or drilling fluid, heavy sand pumps or bailers are operated inside the casing to cut the borehole. Using this method wells to about 600 mm (24 inches) in diameter and to about 30 m in depth for irrigation have been drilled and screened within one day (USDI, 1981).

Jet drilling is basically a percussion method combined with a pressure pump. The drill pipe is lifted and dropped, which chops up material at the bottom of the hole. The water helps to jet the broken material loose and caries the cuttings up the hole where they are discharged (Campbell and Lehr, 1973; Driscoll, 1989).

The full-hole cable tool method and jet drilling are readily adapted to drilling holes 50 to 100 mm (2 to 4 inches) in diameter in soft formations such as sand and silt. These holes are

Figure 7.2 Major components of a rotary drilling rig (After Driscoll, 1989)

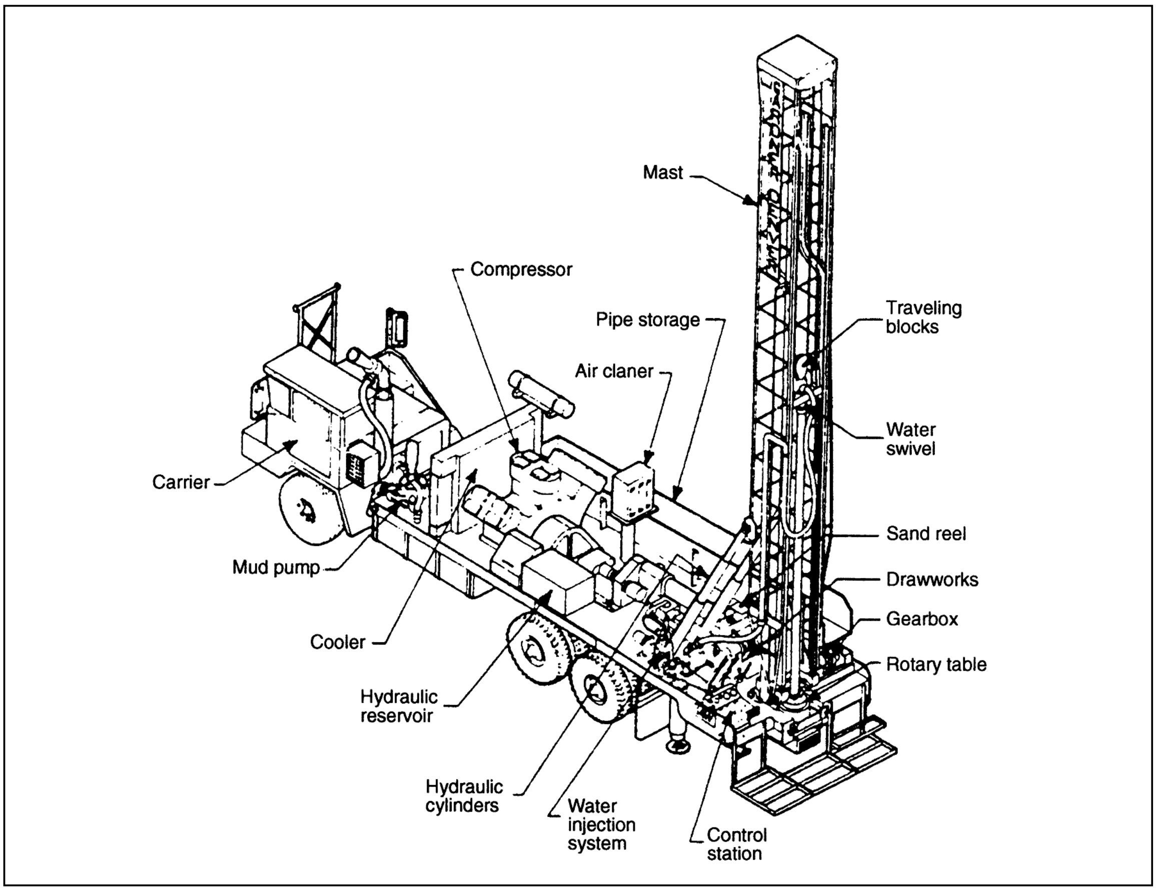
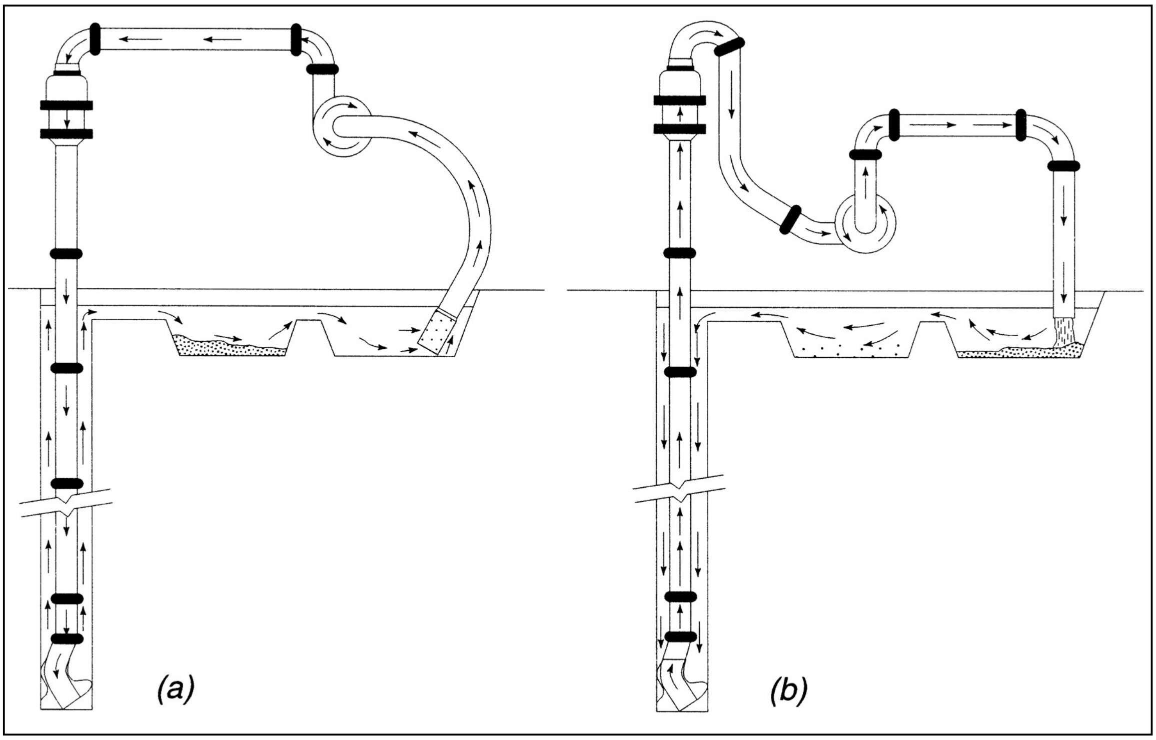


Figure 7.3 Diagrams of a direct (a) and a reverse (b) rotary circulation system; arrows indicate the direction of the mud circulation



useful in installing aquifer test observation wells, monitoring wells and small capacity water wells.

Both the jetting and full-hole cable tool variation methods are disadvantageous in regard to sampling of formations and water during the constructing operation. Also, if the permeability of the penetrated sediments is much greater than about 10 m per day, injected water may disturb the water characteristics at some distance from the wall of the hole (Campbell and Lehr, 1973; Driscoll, 1989; Lehr et al., 1987; Roscoe, 1985).

7.5.2 Direct circulation rotary drilling and variations

The hydraulic direct circulation rotary method of drilling was first used in the oil fields of Louisiana (United States of America) about 1890. Since that time it has become important for the drilling of water wells, especially in deep sedimentary confined aquifer systems (Anderson, 1992; Driscoll, 1989; Lehr et al., 1987; Roscoe, 1985; USDI, 1981). The direct circulation rotary rig drills by rotating a bit at the lower end of a string of drill pipe, which transmits the rotating action from the rig to the bit (Figure 7.2). As the bit is turned, water-based drilling fluid (mud) is pumped down the pipe to lubricate and cool the bit, to pick up material from the bottom of the hole, and to clean the hole by transporting the cuttings to the surface in the annular space between the uncased hole wall and the drill pipe. The water based drilling fluid also forms a thin layer of mud on the wall of the hole which reduces seepage losses and, together with the hydrostatic head exerted by the mud column, holds the hole open. At the surface, the fluid is channelled into a settling pit or pits where most of the cuttings drop out. Clean fluid is then picked up by the pump at the far end of the pit or from the second pit and is recirculated down the hole (Figure 7.3) Drilling fluids used in the water well industry include water-based and

Table 7.2 Major types of drilling fluids used in the water well industry

Water-based	*Air-based*
1. Clean, freshwater	1. Dry air
2. Water with clay additives	2. Mist: Droplets of water entrained in the air stream
3. Water with polymeric additives	3. Foam: Air bubbles surrounded by a film of water containing a foam stabilising surfactant
4. Water with clay and polymeric additives	4. Stiff foam: Foam containing film strengthening materials such as polymers and bentonite

air-based systems; the major types are shown in Table 7.2 (Anderson, 1992; Driscoll, 1989; NWWA, 1985; USDI, 1981).

The type of drilling fluid selected will depend principally on the rock formation expected to be penetrated, the equipment available, and occasionally on environmental regulations, and the experience of the drilling crew. For example, water-based drilling fluid with clay or polymeric additives is commonly used in unconsolidated formations, while air-based fluid is used in consolidated rocks and sediment; and clean water is used by reverse rotary drilling equipment for drilling large diameter wells in unconsolidated or semi-consolidated sediments. Regardless of which drilling fluid type is used, its effectiveness will depend upon the crew's ability to anticipate the chemical and physical changes taking place during drilling and to modify the fluid as required.

At a minimum, all managing staff on a rotary rig should be able to measure drilling fluid density and viscosity, and understand the relationship of these properties to hole stability, cuttings removal, and fluid-loss control (Driscoll, 1989; NWWA, 1981; USDI, 1981).

Because of limitations in pump capacity and therefore effective cuttings removal, most direct rotary equipment used to drill water wells is limited to boreholes with maximum diameters of 550 to 600 mm (22 to 24 inches).

These sizes may not be sufficient for high-capacity wells, especially those that must be filter packed. As hole diameter increases past 600 mm (24 inches), the rate of penetration becomes less satisfactory. Under these conditions, telescopic well designs have been adopted, as shown in Section 7.7. Direct rotary drilling, the most common method, offers advantages and disadvantages, as presented in Table 7.3 (AWWA, 1984; Driscoll, 1989; Lehr et al., 1987; USDI, 1981).

7.5.3 Reverse circulation rotary drills

Reverse circulation drills have been designed to overcome the limitation on hole diameter and drilling rate. Originally they were used only in unconsolidated formations. Recently, the reverse circulation rotary drills have been used both in fairly consolidated rocks such as sandstone and even in hard rocks, using both water and air as the drilling 'fluid'. This machine is probably the most rapid drilling equipment available for unconsolidated formations, but requires a large volume of water which must be constantly replenished since drilling mud is rarely used.

In Table 7.4 are presented the advantages and disadvantages of the reverse circulation rotary method (AWWA, 1984; Driscoll, 1989: Lehr et al., 1987; NWWA, 1981; EPA, 1975; USDI, 1981).

Table 7.3 Advantages and disadvantages of the direct circulation rotary drilling method

Advantages	Disadvantages
1. Penetration rates are relatively higher in all types of materials, and drilling depth capacities are greater	1. The cost of a rotary rig is much higher than a cable tool rig of equal capacity, and maintenance and repair are more complex
2. Minimal casing is required during the drilling operation, and permits use of most geophysical well-logging equipment	2. Requires much more skilled and larger crew with drilling fluid knowledge and experience
3. Rig mobilisation and demobilisation are rapid	3. Mobility of the rigs may be limited by the land surface, mostly the slope or wetness conditions
4. Well screen can be set easily as part of the casing installation	4. Collection of accurate samples requires special procedures and use of drilling fluids may cause plugging of certain formations

Table 7.4 Advantages and disadvantages of the reverse circulation rotary method

Advantages	Disadvantages
1. Large-diameter holes (400 to 1,800 mm) quickly and economically	1. It requires a large volume of water can be drilled which must be constantly replenished
2. The porosity and permeability of the penetrated formations are relatively undisturbed near the borehole	2. The rigs and components are usually larger maintenance more complex, and thus more expensive
3. Well screens can be installed easily as part of the casing installation	3. More larger and skilled crew are generally required
4. Most sedimentary formations can be easily drilled	4. Larger mud pits are required than for other drilling methods
5. This drilling method is probably the most rapid drilling equipment available for unconsolidated formations	5. Igneous and metamorphic rocks can not be drilled

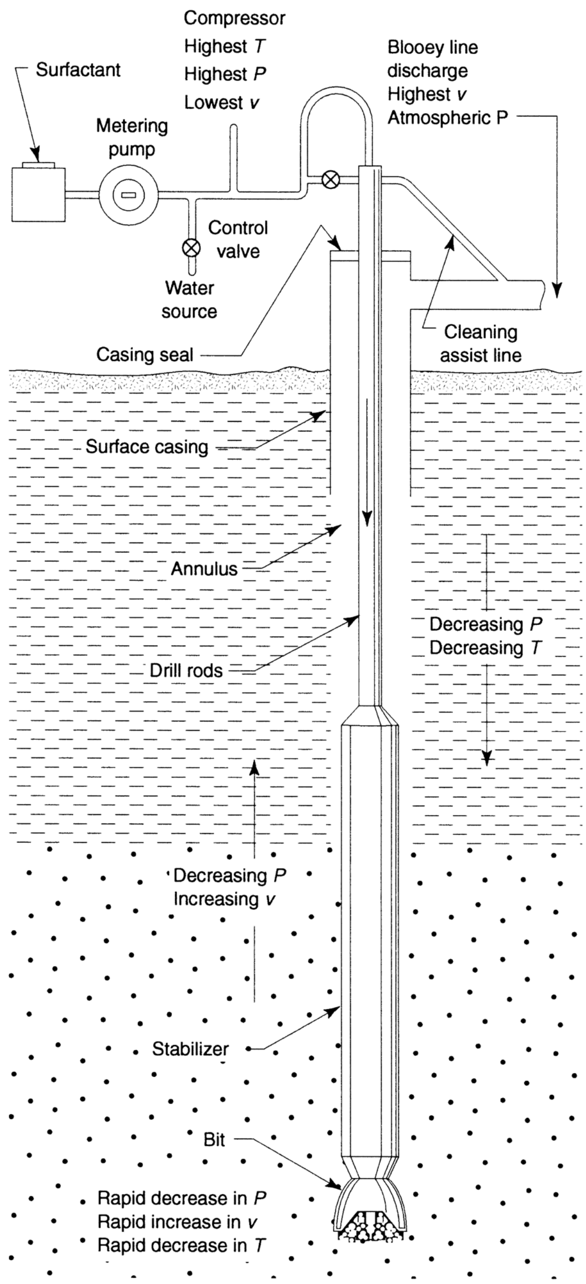
The reverse circulation drill has essentially the same design as the direct rotary drill except that most pieces of equipment are larger. Also, the operation is essentially the same as a direct circulation rotary rig, except that the water is pumped up through the drill pipe rather than down through it (Figure 7.3). The fluid returns to the borehole by gravity flow, and moves down the annular space between the drill pipe and borehole wall to the bottom of the hole, picks up the cuttings, and re-enters the drill pipe through ports in the drill bit. The discharge is directed into a large pit in which the cuttings settle out. To prevent the hole from caving in, the fluid level must be kept at ground level at all times, even when drilling operation is suspended temporarily. Much reverse drilling equipment is equipped with air compressors to aid in circulating the drilling fluid. The hydrostatic pressure of water column plus the velocity head outside the drill pipe support the borehole wall (AWWA, 1984; Driscoll, 1989; Lehr et al., 1987; NWWA, 1981; USDI, 1981).

In the reverse circulation rotary method, the drilling fluid can best be described as muddy water rather than drilling fluid, and additives are rarely mixed with the water to make a viscous fluid. Suspended clay and silt that recirculate with the fluid are mostly fine materials picked up from the penetrated geological formations.

7.5.4 Air drilling systems

The air rotary drilling method was developed primarily in response to the need for a rapid drilling technique in mining exploration and geotechnical surveys and more recently in hard

Figure 7.4 Basic components of an air rotary drilling system (After Driscoll, 1989)



rock hydrogeology. Two different drilling methods use air as the primary drilling fluid: direct air rotary method and down-the-hole air hammer. The direct air rotary equipment is essentially the same as for direct circulation rotary drilling, except the fluid channels in the bit are of uniform diameter rather than jets, and the mud pump is replaced by an air compressor (Figure 7.4). Air is circulated down the drill string to cool the bit and to blow the cuttings to the surface (Driscoll, 1989; Lehr et al., 1987; Roscoe, 1985).

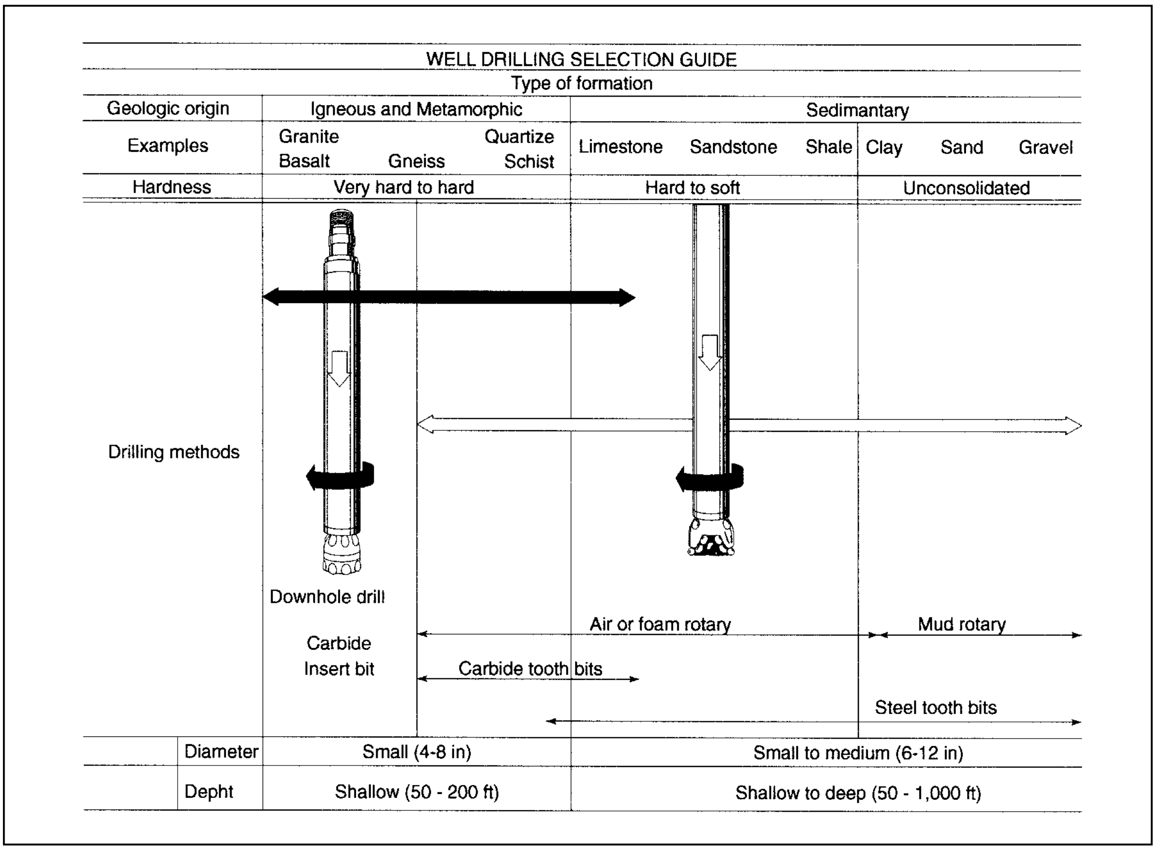
Initially, the air rotary method was used for relatively small diameter such as 50 to 150 mm (2 to 6 inches) boreholes in hard rocks. Larger holes have become possible through use of foams and other air additives, and diameters up to 800 mm (36 inches) have been drilled successfully. The equipment is mostly used in hard rock terrain where water is encountered in fractures or similar openings and wells are completed as open holes. Penetration rates are often faster and the bit life longer when using air as compared with water-based drilling fluids. Better bottom-hole cleaning is partly responsible for this difference in performance. If too much water from the formation comes into the hole during drilling, however, the penetration rate is no better than when drilling with water-based drilling fluids (Driscoll, 1989; USDI, 1981).

Tricone rock bits, up to about 305 mm (12 inches) diameter, similar to those designed for drilling with water-based fluids, are commonly used. In many areas bottom bits, with sintered tungsten-carbide inserts in the perimeter of the steel rollers, are used successfully (Driscoll, 1989; Lehr et al., 1987; USDI, 1981). Figure 7.5 shows the guide for the use of bit types in air drilling systems (Driscoll, 1989).

Because of the time lag before cuttings arrive at the surface, it is not advisable to base well design on samples of cuttings.

Shortly after the development of air rotary drilling, the down-the-hole hammer bit was

Figure 7.5 Guide for the use of bit types in air drilling system (After Driscoll, 1989)



developed. This second direct rotary method using air is called the 'down-the-hole' drilling system. A pneumatic drill operated at the end of the drill pipe rapidly hammers the rock while the drill pipe is slowly rotated. The rate of penetration in several rock types is higher than that obtained by other drilling methods or other types of tools. Cuttings are removed continuously by the air used to drive the hammer. This arrangement efficiently combines some of the advantages of the cable tool and rotary rig. Advantages and disadvantages of the air rotary drilling systems are presented in Table 7.5 (Driscoll, 1989; Lehr et al., 1987; USDI, 1981).

Table 7.5 Advantages and disadvantages of the air rotary drilling systems

Advantages	*Disadvantages*
1. Penetration rates are high, especially down-the-hole hammers in high resistant rocks such as granite, dolomite	1. Restricted to consolidated or with semi-consolidated rocks
2. Cuttings removal is extremely rapid	2. Maintenance costs are high
3. Estimates of the yield from a particular aquifer feature penetrated can be made during the drilling	3. Initial cost of large air compressor is high

Drilling depth of rotary air and down-the-hole hammer systems in saturated materials is limited by the available air pressure which must be greater than that exerted by the column of water in the hole, otherwise the rig stops working if the rig is to function.

7.6 Auger hole drilling

Although geotechnical and geological exploration has traditionally used augers, their use in the water industry has been quite limited until recently. Currently, many well contractors bore with earth augers when installing of groundwater monitoring wells and aquifer clean-up wells. Although monitoring wells can be drilled by virtually any drilling method, auger techniques are used to install the overwhelming majority of monitoring wells in the world, because of the availability and relative ease of collecting formation samples (ASTM, 1992; Driscoll, 1989; NGWA, 1965; EPA, 1975).

Three principal types are used commonly: (1) hand auger, (2) solid stem auger and (3) hollow-stem auger (Figures 7.6, 7.7 and 7.8). These earth augers have the important advantage that drilling fluids and mud are normally not required. Another advantage is the ability to provide accurate information on all geological materials being drilled, either by cuttings, split-spoon or barrel samples, and side wall samples.

7.6.1 Hand augers

Hand augers have generally been used to install shallow production or monitoring wells up to 10 metres in depth and with casing diameters of 100 mm or less. A typical hand auger advances by turning into the soil until the auger is filled (Figure 7.6). The device is then removed and the sample is dumped from the auger. Motorised units for one or two operators are available.

Generally, the borehole cannot be advanced below the water table for more than one or two metres, because of the risk of borehole collapse. It is often possible to avert borehole collapse below the water table by driving a wellpoint into the augered hole, thereby advancing

Figure 7.6 Diagram of a hand auger (After Aller et al., 1989)

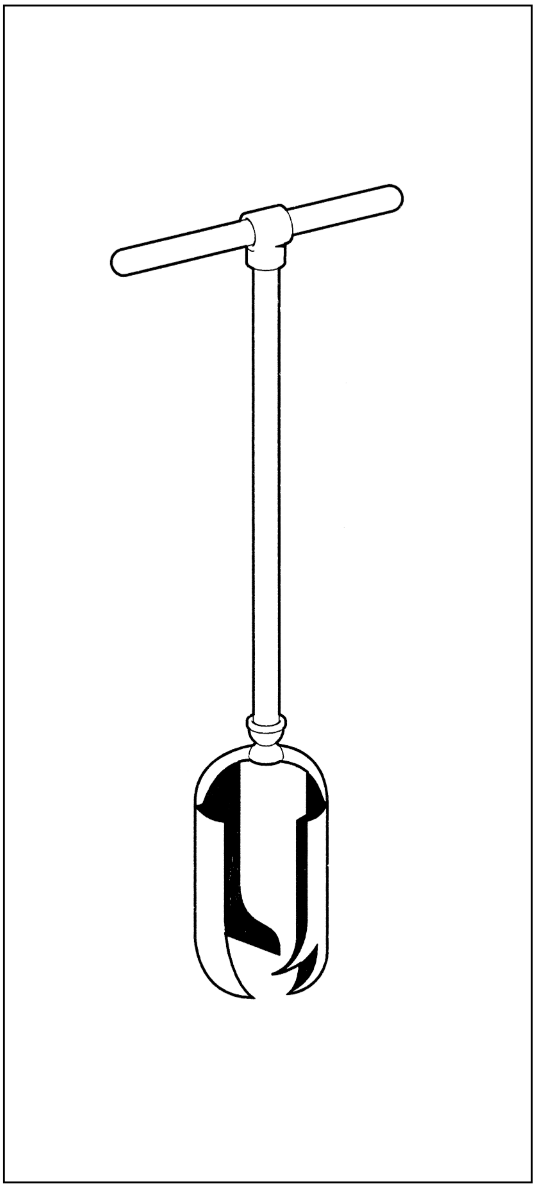
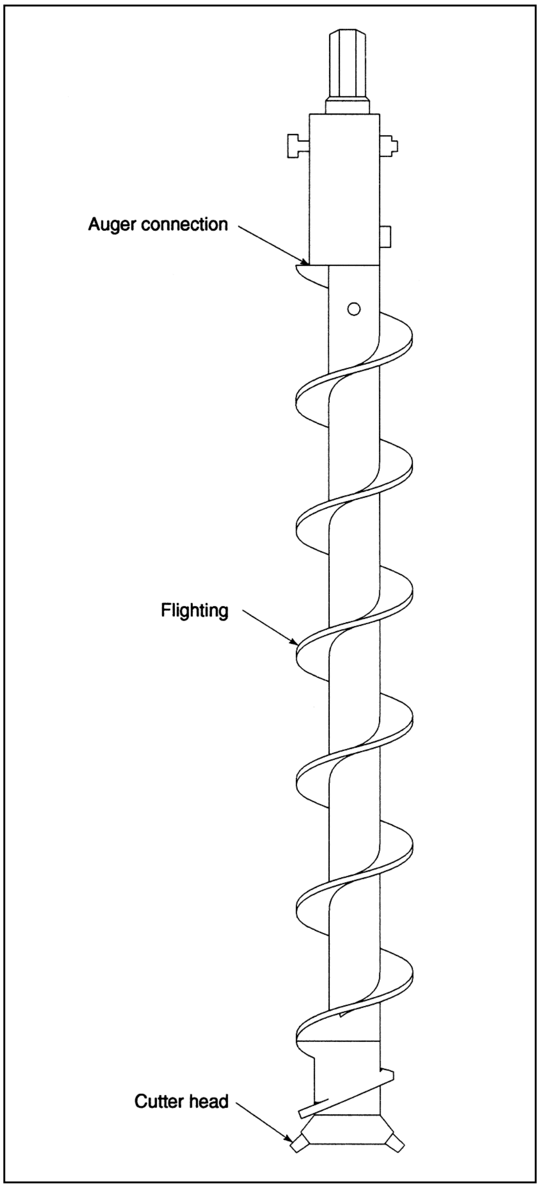


Figure 7.7 Diagram of a solid-stem auger (After Aller et al.,1989)



the wellpoint below the water table. The wellpoint can then be used to pump water or to measure water levels and to provide access for water quality samples

Table 7.6 lists the advantages and disadvantages of the auger methods used to install production and monitoring wells (Aller et al., 1989; ASTM, 1992; Driscoll, 1989; Fetter, 1993; EPA, 1975; USDI, 1981).

7.6.2 Solid-stem augers

The solid-stem auger is advanced by a rotary drive at the surface and forced downward by a hydraulic pulldown device (Figure 7.7). Augers are available in diameters up to about 600 mm (24 in). Where used for monitoring well installation, available auger diameters typically range from 150 to 350 mm (6 to 14 in) in external diameter.

Table 7.6 Major advantages and disadvantages of the auger techniques

Type	*Advantages*	*Disadvantages*
Hand augers	1. Good soil sampling 2. Aquifer zones identification 3. Easy installation of shallow water wells or monitoring wells	1. Limited to very shallow depths 2. Unable to penetrate below the water table 3. Unable to drill very dense formations
Solid-stem augers	1. Good soil sampling with split-spoon or thin-wall samplers 2. Able to penetrate below the water table 3. Fast and mobile	1. Soil samples limited to areas and depth of predominant stable formations 2. Unable to work in most unconsolidated formations 3. Depth capacity decreases quickly as diameter of auger increases
Hollow-stem augers	1. Good soil sampling with split-spoon or thin-wall samplers 2. Easy drilling and installation in all unconsolidated formations 3. Water-quality sampling during drilling 4. No drilling fluid is used 5. Usually less expensive than rotary or cable tool drilling	1. Can be used only in unconsolidated formations 2. Limited to depths of 30 to 50 m. 3. Possible problem in controlling heaving sands 4. Limited diameter of augers limits casing size 5. Cost of hollow-stem augers is relatively high

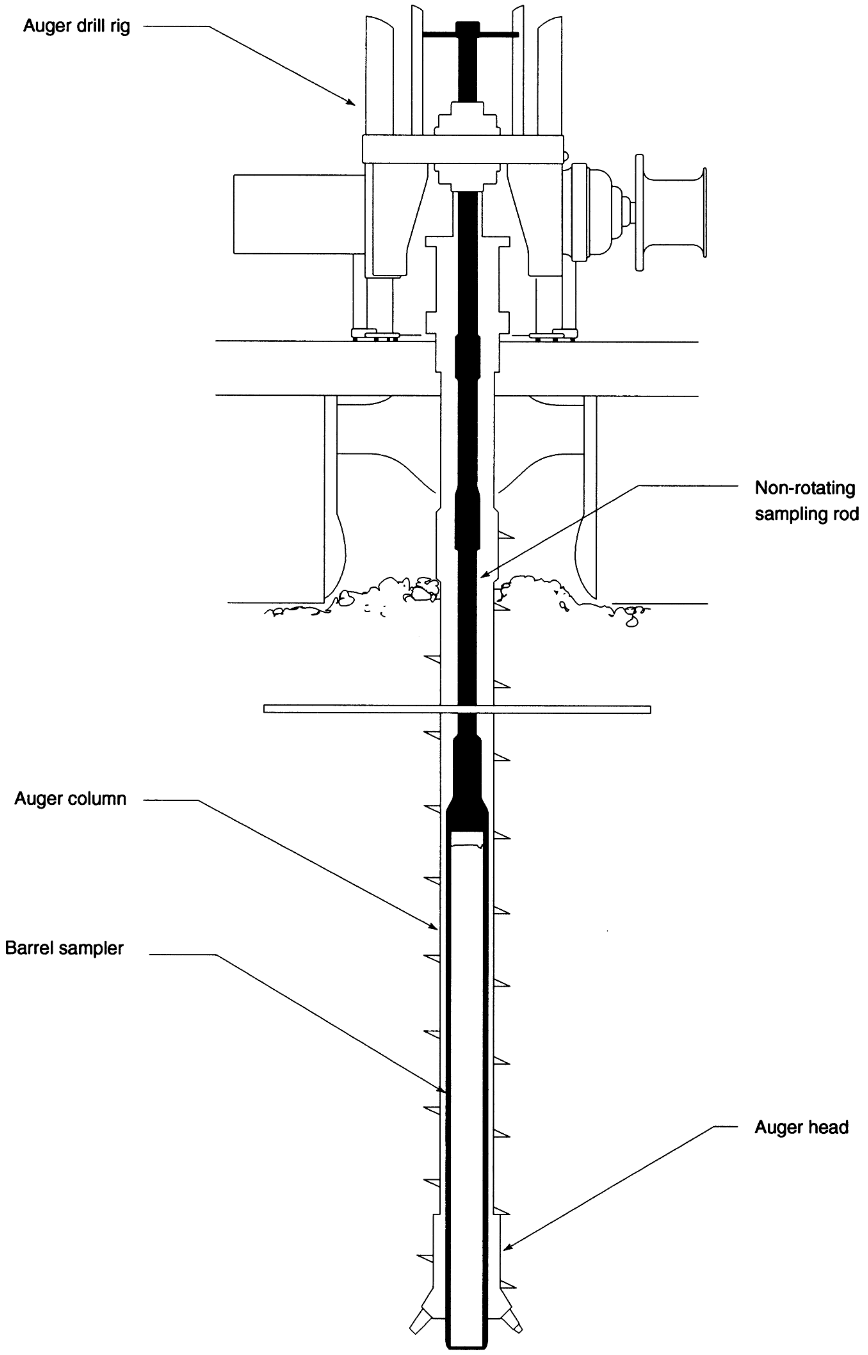
Many of the drilling rigs used for installing monitoring wells in stable unconsolidated material can reach depths of approximately 40 metres with 350 mm augers and approximately 50 metres with 150 mm augers. As the auger column is rotated into soil, cuttings are retained on the flights. The auger is removed from the hole and rock formation samples are taken from the retained earth. (Aller et al., 1989; ASTM, 1992; Campbell and Lehr, 1973; Driscoll, 1989; EPA, 1975).

7.6.3 Hollow-stem augers

The hollow-stem augers are more effective than solid-stem augers because they can be used as temporary casing to prevent caving-in and abandonment of the borehole (Figure 7.8). The hollow-stem method is a fast and efficient techniques of drilling and completing small-diameter wells to moderate depths. Casing pipe and screens can be installed and gravel-packed without using temporary casing or drilling fluids.

The most widely available hollow-stem augers are 158.8 mm (6.25 in) external diameter auger flights with hollow stems of 81.25 mm (3,25 in) internal diameter The equipment most frequently available to power the augers can reach depths of 45 to 50 m in clayey or silty sand formations. Because of their availability and the relative ease of formation sample collection, hollow-stem augering techniques are used for installing the overwhelming majority of monitoring wells in the United States of America.

Figure 7.8 Diagram of a hollow-stem auger (After Aller et al., 1989)



7.7. Well design and construction

7.7.1 General

The controlling factors in permanent well installations are usually sanitation, stability, and an estimated minimum useable well life of 25 years. Selection of the design features of a well and a particular method of construction depends upon the well's objective, the quantity of water desired, economic factors, and hydrogeological conditions. Furthermore, experience has shown that well design features and construction practices have measurable effects on well performance and operating life and on the economic utilisation of the well.

It is obviously not good engineering to design a 200 m^3/h well to serve an industry when 20 m^3/h will suffice. It is equally poor practice to choose materials of inferior quality, merely to cut initial costs. This would only reduce the useful life of the well and increase operation and maintenance costs (Campbell and Lehr, 1973; Driscoll, 1989; Lehr et al., 1987). Nevertheless, despite these relationships, the engineering and scientific aspects of well design have received little attention. As a result of this and other factors, water well design has commonly been based on the experience, observations, and judgement of the designer and driller.

A number of commonly used design standards that may be of interest to users have been published by governmental, private or professional organisations, but there are so many that it is impossible to describe them all here (Ahrens, 1987; Anderson, 1992; ASTM, 1992; Bowman, 1911; Campbell and Lehr, 1973; Driscoll, 1989; EPA, 1975; USDI, 1981).The guidelines presented below are affected by governmental regulations and by site-specific geotechnical, hydrogeological, and subsurface chemistry conditions.

7.7.2 Particulars of design

Well design is the process of specifying the drilling techniques, materials and physical dimensions for a production or monitoring well. The design guidelines presented below focus mainly on the design of municipal, industrial and irrigation wells (Ahrens, 1970; Aller at al., 1989; Anderson, 1992; ASTM, 1992; Lehr et al., 1987; NGWA, 1965; EPA, 1975).

The principal objectives of good design should ensure the following:

- durable and reliable construction,
- production of the desired yield or the highest yield available from the aquifer with the highest efficiency,
- accurate water level measurement and/or extraction of representative groundwater quality samples,
- efficient hydrogeological site characterisations.

The basic information desired for the design of efficient production well includes:

- aquifer types (water table or confined aquifers),
- current depth and trends of the water levels,
- stability characteristics and rock formations of the unsaturated zone of the water table aquifer or at the top of a confined aquifer,
- thickness, nature of the porosity (interstitial or resulting from secondary voids), and the degree of confinement of the aquifer,
- grain size and sorting of the water-bearing formations,
- available casing and screen materials,
- desired yield and pump size patterns,
- design and construction features of wells previously drilled in the area,
- state and local statutes and regulations,
- operation and maintenance history of previously constructed wells (control of corrosion, encrustation and contamination),

Unfortunately, the information desired is rarely all available, even in a developed area. In an undeveloped area, all that may be known initially is the approximate location of the proposed well and the desired yield.

Most frequently, for major wells (yields greater than 20 m^3 per hour) the available data should be supplemented by a hydrogeological investigation and the drilling of a pilot hole. The pilot hole will provide an accurate lithological log, sediment samples for mechanical analyses, water samples for chemical analyses, and information on static water levels and the type of aquifer present.

If the site is found inadequate for any reason, the pilot hole can be abandoned without the major cost of drilling a production well. If the site is found to be satisfactory, a good well design can be prepared and specifications defined. Furthermore, the contractor can foresee many problems, minimising the unknown factors, the risks during construction, and even saving the cost of the pilot hole.

For minor wells, the well drilling costs may be about the same as for a pilot hole; thus, the latter may be uneconomic. As a result, the recommended procedure for the design of such wells is to make a preliminary design based on the available information. Additions or changes to the preliminary design such as in casing diameters and screen slot size and setting can be made on the basis of information obtained during the drilling (Ahrens, 1970; Campbell and Lehr, 1973; Driscoll, 1989; Lehr et al., 1987; EPA, 1975).

The particular design of monitoring wells will depend on:

- how the well is to be used, whether for taking water samples, for measuring the elevation of the water table, or for recovering contaminants,
- the hydrogeological context,
- the chemical reactivates/inertness of the contaminants,
- cost of casing and screen materials.

The terminology used for various components of a well is not standardised. Consequently, various terms are used for similar components in different publications and in different parts of each country. In the following sections, the terminology is based on the cited bibliography.

7.7.3 Main components of a water well design

Standard design procedures involve choosing the casing diameter and material, estimating well depth, selecting the length, diameter and material for the screen, determining the screen slot size, and choosing the completion method. Whether a well is naturally developed or filter packed depends on the grain-size distribution curve of the water-bearing formation. Coarse-grained non-homogeneous material can be developed naturally, whereas fine-grained homogeneous materials are best developed using a filter pack. Well screen slot openings for either method are selected from a study of sieve-analysis data for samples representing the aquifer materials.

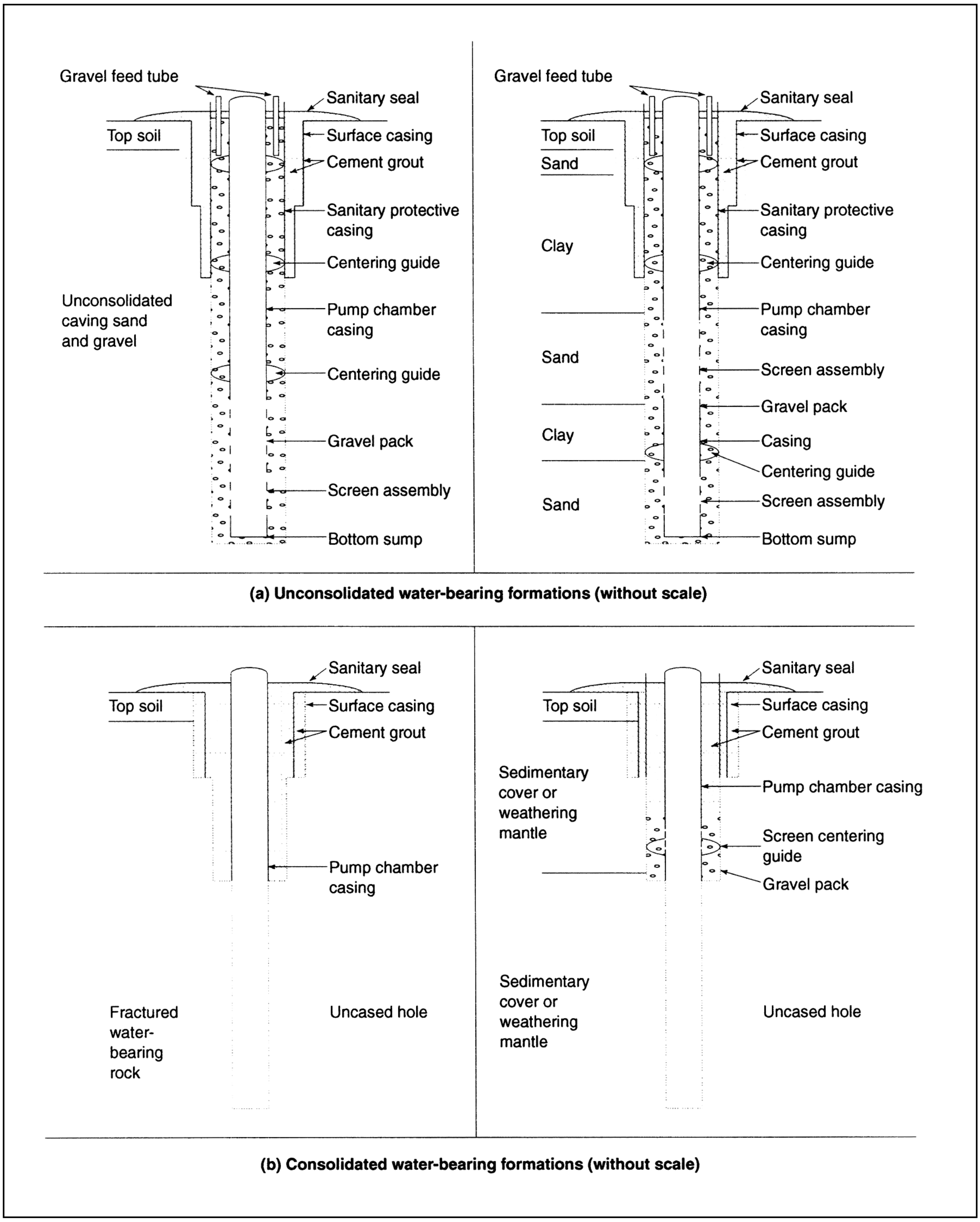
Every well consists of four main parts: surface casing, the pump chamber casing, the riser pipe to conduit water upward from the aquifer intake portion to the pump intake, and the screen assembly (Figure 7.9 and 7.10).

In consolidated rock aquifers, the borehole length serving as the water conduit and the intake portion of the well may be left uncased. Some consolidated water-bearing formations, however, such as sandstone, may deteriorate over time because high flow rates remove the cement that holds sand grains together, thus causing a slow collapse of the borehole wall. Therefore, screens are often used to protect pumps from loosened formation particles, and to stabilise the water-bearing formation in many consolidated rocks, especially sandstone, limestone, and sometimes granite (Anderson, 1992; Campbell and Lehr, 1973; Driscoll, 1989; Lehr et al., 1987; Roscoe, 1985; USDI, 1980).

(i) Surface casing

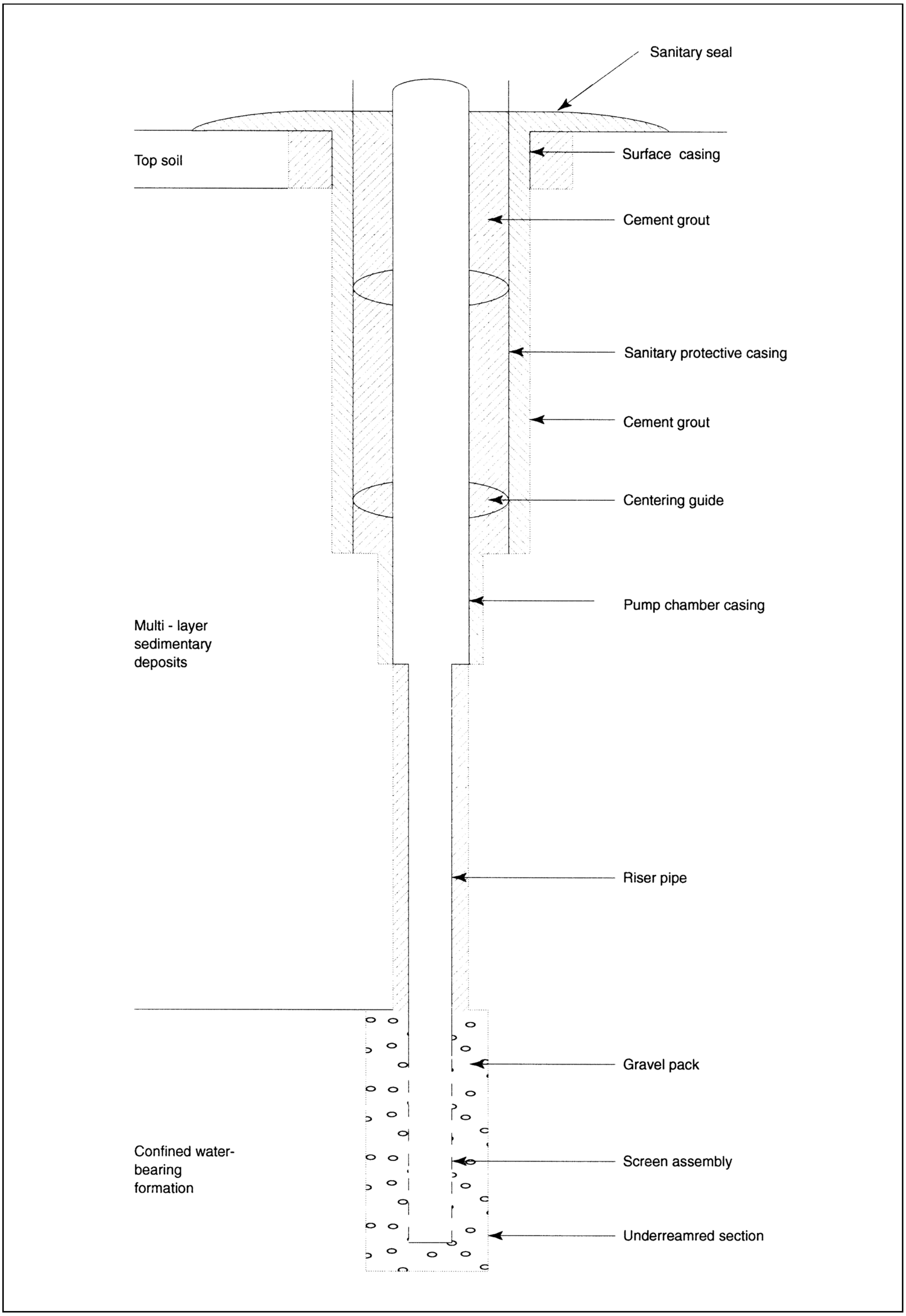
Surface casing is generally installed to facilitate drilling a well by supporting unstable materials so they will not cave in and fall into the hole, to reduce loss of drilling fluids, and to facilitate

Figure 7.9 Examples of well construction in unconsolidated (a) and consolidated (b) aquifer formations



installing or pulling back other casing. It is installed from near the ground surface through unstable, unconsolidated, or fractured rocks and up to a short distance into a firm, stable, or massive and, if possible, relatively impermeable layer. The surface casing may be temporary and removed when completing the well, or it may be a permanent part of the structure. When permanent surface casing is used, the first operation in construction of the well is to drill

Figure 7.10 Diagram showing the main parts of a deep water well design



an oversized hole and to install, centre, and grout the casing pipe. Permanent surface casing is frequently used to seal out undesirable surface or shallow groundwater. In fact, it provides a better degree of protection against pollution infiltration, and facilitates placing a sanitary seal.

Table 7.7 gives the recommended minimum and optimum diameters of permanent surface casing for various well yields, taking into account the two main alternatives, a naturally developed well or a gravel packed well (AWWA, 1984; Campbell and Lehr, 1973; Driscoll, 1989; Lehr et al., 1987; USDI, 1981).

Table 7.7 Recommended diameters for pump chamber and permanent surface casing

Desired well yield	*Pump chamber casing diameter*			*Surface casing diameter for naturally packed wells developed wells*		*Surface casing diameter for gravel*	
(m^3/h)	*mm*	*(inches)**		*mm*	*(inches)*	*mm*	*(inches)*
Up to 20	127–152	(5–6)	ID	152–203	(6–8)	406	(16)
20 to 40	152–203	(6–8)	ID	203–254	(8–10)	457	(18)
30 to 80	203–254	(8–10)	ID	254–305	(10–12)	508	(20)
70 to 160	254–305	(10–12)	ID	305–356	(12–14)	559	(22)
100 to 230	305–356	(12–14)	ID	356–406	(14–16)	610	(24)
180 to 400	356–406	(14–16)	OD	406–457	(16–18)	660	(26)
300 to 700	406–508	(16–20)	OD	457–559	(18–22)	762	(30)
450 to 860	508–610	(20–24)	OD	559–460	(22–26)	864	(34)
700 to 1,400	610–762	(24–30)	OD	660–813	(26–32)	1,016	(40)
900 to 1,800	711–762	(28–32)	OD	762–864	(30–34)	1,067	(42)

* The size of the pump chamber casing is based on the outer diameter of the bowls for vertical turbine pumps and on the diameter of either the pump bowls or the motor for submersible pumps. The increasingly sophisticated technology of the pumping equipment, makes it advisable for the well designer to contact a pump supplier, about the desired yield, the head conditions and the required pump efficiency.

For nominal diameters ranging between around 203 mm (8 in) and 457 mm (18 in) and to maximum depths between 10 and 50 metres, the ASA schedule number or class 10 weight pipe and wall thickness of 0.25 inches is usually adequate for permanent surface casing with welded joints. For setting depths up to 150 m, and for diameter up to around 813 mm (32 in) the class 20 pipe with wall thickness ranging between 0.35 and 0.5 inches, is generally used. In gravel-packed wells of the depth and diameter of most water wells, the gravel pack is usually placed through nominal 51 mm (2 in) to 102 mm (4 in) coupled pipes. To permit insertion of these pipes, there must be sufficient annular space between the surface casing or the wall of the hole and pump chamber casing to allow the pipe couplings to pass through (Anderson, 1992; Driscoll, 1989; Lehr et al., 1987; Koering, 1967; USDI, 1981).

After the surface casing has been installed and the grout has set, the well is drilled deeper through the bottom of the cased hole. The borehole drilled is usually about 50 mm (2 in) smaller in diameter than the outside diameter of the surface casing.

(ii) Pump chamber casing

The pump chamber casing furnishes a direct connection between the ground surface and the aquifer intake part of the well, and when permanent surface casing is not used, it seals out undesirable surface or shallow groundwater and supports the sides of the hole. The design of the pump chamber casing, the water conduit casing and the screen assembly requires careful consideration of the hydraulic factors that influence well performance (Ahrens, 1970; Driscoll, 1989; Lehr et al., 1987; Roscoe, 1985; EPA, 1975).

Depth of setting of a pump, hence depth of the pump chamber casing, is determined by estimating projected pumping levels and considering the following factors:

- present static water level and long-period fluctuations of levels. Probable drawdown at desired yield,
- possible interference by other wells or boundary conditions. Required pump submergence,
- the presence of any telescoping overlap.

Pump chamber casing should be always installed plumb and straight. Deviation from the vertical should not exceed two-thirds of the inside diameter of the casing per 30 m of depth approximately.

The diameter of the pump chamber casing must be large enough to accommodate the pump required for the desired yield, with enough clearance for installation and efficient operation. Sometimes, technological and economical constraints may impose several reductions in size of the pump chamber casing, so that the desired diameter can be set at the required depth by telescoping. Unless the inner casings must be cut off to accommodate the pump bowls, or the value of the casing makes retrieval worthwhile, it is a good idea to leave all casing strings in the hole. Under these conditions one eliminates the need for much of the grouting, the risk of leaks between casing strings, and the chance that overlapping inner casings will be damaged during cut-off. If casing is cut off, this should always be done after development of the well (Driscoll, 1989; Lehr et al., 1987; EPA, 1975).

Pump chamber casing should be grouted except when it is run inside a grouted surface casing. Grout must then be placed in the annulus between casings if the aquifer material will expand or if the two casings are set in water-bearing formations that have different static water levels.

(iii) Riser pipe

The riser pipe comprises all casing for water flowing upward from the aquifer portion intake to the pump intake. In single string construction of uniform diameter, blank pipe may extend from the top of the screen up to the pump chamber casing to which it may be attached by welding, a coupling, or a reducer, or the riser pipe may be telescoped into the pump chamber casing for one or several meters. Where the extension is relatively short, it is sometimes referred to as a flush tube extension or overlap pipe .

Under these conditions, the diameter of the riser pipe must be sufficient to assure that the uphole velocity is 1.5 m/sec (5 ft/s) or less. If the pump is set below any screened section, there will be sufficient area around the bowls to allow water to pass downward with minimum head loss to the pump intake. However, heat build-up can be a problem for a submersible pump set in a blank casing section beneath the screen, because the intake portion of the submersible pump is located above the motor.

In deep wells that have both high static and high pumping water levels, the casing diameter can be reduced at a depth below the lowest anticipated pump setting, to reduce drilling and material costs. Furthermore, drilling conditions, drilling methods, or economic factors sometimes make it necessary to complete the lower portion of the well with smaller diameter casing or screen. More than one inner casing can be telescoped depending on well depth. In that case, the casing diameters for individual segments can be chosen from the bottom of the well upward, to accommodate the driving conditions, the overlap requirements, and the annular space necessary for grouting or filter packing. Ideally, each casing string should be anchored in a poorly permeable bed. If the telescoped casing section ends in sand, water should be run continuously into the annulus between the two strings. If the casing segments are anchored in thick clay beds, no grout may be necessary. In the design process, casing diameters for individual segments must be chosen taking into account that grouting regulations usually require at least a 51 mm (2 in) annular space (Ahrens, 1970; Driscoll, 1989; Lehr et al., 1987; EPA, 1975; USDI, 1981). This is done in many wells completed in confined aquifers where pressure is relatively high, such as the case in the huge sedimentary basins in Australia, Brazil, and Russia (Habermehl, 1985; Margat, 1990; Rebouças, 1976, 1994b). Upward velocity in any smaller diameter casing beneath the pump bowls should be 1.5 m/sec (5 ft/sec) or less. Table 7.8 lists maximum discharge rates for various water conduit casing sizes that produce moderate friction losses (Driscoll, 1989; Lehr et al., 1987; USDI, 1981).

Table 7.8 Maximum discharge rates for optimum diameter of riser pipe; based on an upward velocity of 1.5 m/sec

Desired well yield	*Optimum diameter of pump chamber casing*		*Optimum diameter of riser pipe*	
(m^3/h)	*mm*	*in*	*mm*	*in*
Up to 50	203	8	102	4
70	254	10	127	5
100	305	12	152	6
180	356	14	203	8
280	406	16	254	10
400	438	18	305	12
500	508	20	337	14
650	508	20	387	16
800	610	24	438	18
1,000	711	28	489	20
1,500	762	30	591	24

To meet certain pressure requirements, pipe is manufactured in three general weight classes: standard, extra heavy, and double extra heavy. Since 1935 the American National Standards Institute (ANSI) has assigned schedule numbers to classify wall thickness for different pressure applications. Depending on the wall thickness, however, the inside diameter may be less than or greater than the number indicated. To complicate matters further, pipe manufacturers have developed their own products with specifications that only approximately follow the design guidelines established by ANSI (Driscoll, 1989; Lehr et al., 1987; USDI, 1981). Under these conditions, it is difficult to present a simple analysis of the various types of casing used in the water well industry.

The term 'nominal' is used to designate the inside diameter, because the actual sizes vary somewhat above or below the standard size. Data presented in Table 7.8 show that in terms of hydraulic factors, string construction of uniform diameter is usually possible for minor wells.

At the present time, most water well drolleries use pipe constructed to ASTM standards rather than to API standards. In the huge sedimentary basins, however, pipe used for water wells is generally constructed to API standards.

Pipe used for constructing water wells is built conforming to major standard specifications: ASTM standards, and API standards. In general, ASTM standards steel pipe A-53 or

A-120 are recommended for most drilling situations and typical water quality conditions. API standard 5L, either grade A or B, is commonly used for exceptionally corrosive water, for deep wells, and for difficult drilling conditions, which are characterised by very dense water-bearing formations and deep boreholes having a large diameter. In every case, the wall thickness must be sufficient to support full hydraulic loading if the casing is pumped dry (Anderson, 1992; Driscoll, 1989; Lehr et al., 1987; USDI, 1981).

Western European manufacturers sometimes produce pipe according to other specifications that is similar, but not identical; in few cases European pipe is made to metric measurements rather than to dimensions in the Imperial system.

Aggressive corrosion, heavy encrustation, or both, often require the use of non-metallic pipes. In most cases, thermoplastic pipe such as polyvinyl (PVC), polyethylene, acrylonitrile-butadiene-styrene (ABS), or rubber modified polystyrene (SR) is the most satisfactory and economical. Standardisation of thermoplastic well casing is covered under ASTM Standard F480 (ASTM, 1992; AWWA, 1984; NGWA, 1965; EPA, 1975). The major advantages are light weight, ease of installation, corrosion resistance, and low price. In water wells drilled in consolidated rock to depths in excess of 250 m, thermoplastic pipes of diameters up to 203 mm (8 in) have been used successfully to supply industrial plants and domestic water supply systems. However, these materials lack the tensile and impact strengths. Thermoplastic pipe of suitable wall thickness and in diameters up to 254 mm (10 in) or 305 mm (12 in) is manufactured for a setting depth of about 100 metres in unconsolidated water-bearing formations. Fibreglass reinforced thermoplastic pipe up to 254 mm (10 in) in diameter with 4.6 mm (0.18 in) to 5.1 mm (0.2 in) wall thickness has been used extensively in some areas in water wells to depths of about 150 m. However, the conditions are exceptional, and collapse under normal development procedures has been reported to be a problem (Anderson, 1992; Lehr et al., 1987; NWWA, 1981; USDI 1981).

(iv) Screen assembly

Well screens are required in all unconsolidated and most semi-consolidated formations and occasionally in consolidated aquifer rock. Three main factors govern the selection of material used to fabricate well screens: (1) encrustation/corrosion factors, (2) strength requirements of the screen, and (3) cost factors. The least expensive and most commonly available screens are made of low carbon steel. Those made from non-ferrous metals and alloys, thermoplastics, and exotic materials are used in areas of aggressive corrosion and encrustation, to prolong well life and efficiency (Anderson, 1992; Driscoll, 1989; Lehr et al., 1987; EPA, 1975).

The screen type may range from pipe perforated in place to carefully fabricated cage-type wire-wound screen with precisely sized openings. Perforated casing made by sawing, machining, or torch cutting were commonly used until recently. Slot openings usually range from about 0.25 to 5.0 mm; the maximum percentage of open area is about 10 percent for the larger slots. Another screen type of low efficiency has been manufactured with stamped perforations. Slots range from 1.5 to 5.0 mm and maximum percentage of open area is about 20 percent. However, the use of low efficient screen types is rapidly because more efficient screens consisting of a continuous winding of round or specially shaped wire mounted on a cage of vertical rods are becoming cheaper. Almost any slot size is readily available from 0.1 to 6.0 mm, usually in increments of about 0.1 mm. Open areas are the largest obtainable and slot sizes can be closely matched to aquifer gradations. Although such screens are more expensive initially than other types of screen, they are usually more economical, especially when used in thin but highly productive aquifers, and in the deep, confined aquifers (Anderson, 1992; Driscoll, 1989).

The main purposes of a screen assembly are to (1): stabilise the wall of the hole, (2) control sand entrance into the well, and (3) facilitate flow into and within the well. The screen section may consist of only a screen length, or of a screen assembly with a combination of slot sizes desired, or associating blank casing sections. It is advisable to install a bottom sump consisting of 2 to 5 metres of blank casing set below the bottom of the lowest screened section. The sump provides a storage space for sand brought into the well which settles to the bottom and prolongs the effective operation of the total screen length. Screen assemblies set in unconsolidated materials should have a bottom seal or a concrete plug. The seal may consist of a steel plate welded or coupled to the bottom. This seal not only precludes materials heaving up into the well under certain circumstances but also provides a bearing area for support of the screen assembly (Ahrens, 1970; Driscoll, 1989; Lehr et al., 1987; EPA, 1975; USDI, 1981).

Currently, most screens are made in lengths ranging from 1.5 m to 6.0 m which can be jointed by welding or couplings to give the desired length of screen. To avoid electrolytic corrosion, couplings and welding materials should be composed of the same materials as the screen.

The optimum length of well screen is based on the thickness of the aquifer, available drawdown, and nature of the stratification of the water-bearing formation. In virtually all aquifers, certain zones will be more productive than others. Thus, the screened sections of the well must be placed in front of the zones that have the highest hydraulic conductivity. These layers can be selected on the basis of an interpretation of the following information:

- the lithological well log and driller comments on drilling operation such as fluid loss, penetration rate, etc.;
- the geophysical borehole logs such as calliper, resistively, spontaneous potential, natural gamma ray radiation, gamma-gamma ray radiation, and neutron logging;
- grain size distribution curve of the samples of water-bearing formations;
- hydrogeological site evaluation.

In uniform unconfined aquifers a continuous length of screen is usually installed. Theoretical considerations and experience have shown that screening of the bottom one-third to one-half of an aquifer provides the optimum design for aquifers less than 50 m thick. Where aquifers are deeper and thicker, judgement should be used to determine the most economical combination of penetration depth and screened section length. In this case the well screen is usually placed at the bottom of the well and its length should cover not less than 35 percent of the estimated thickness of the penetrated part of the aquifer (Driscoll, 1989).

For non-homogeneous unconfined aquifers, the screen sections must be positioned in front of the most permeable layers of the lower portions of the water-bearing formation, separated by blank pipes opposite the poor aquifer materials. If possible, the total screen length should cover not less than 35 percent of the aquifer layers penetrated by the well, so that maximum drawdown is available.

In confined aquifer conditions, theoretical considerations and experience have shown that full penetration and maximum percentage of screen are recommended where aquifer depth and thickness make such construction economically feasible. Usually, 80 percent of the thickness of the homogeneous and uniform water-bearing sediment, or 80 percent of the most permeable layers of the non-homogeneous sediment should be screened to obtain between 90 and 95 percent of the total specific capacity of the well that could be obtained by screening the entire thickness of the aquifer. An exception to this rule is made when poor-quality water is found in part of an aquifer. In this case, the well should be completed to a depth that will avoid the undesirable water. Any part of the hole drilled into a portion of the aquifer containing poorer quality water should be isolated.

Pump chamber casing sometimes has perforated or screened sections in it, above the pump bowls. This practice should be avoided, because if drawdown increases to depths below the screened section, cascading results in air getting into the water which induces cavitation and may result in other adverse effects such as the inflow of undesirable shallow groundwater. Also, periodic exposure of a screen to the atmosphere may contribute to corrosion and/or to growth of organisms such as iron bacteria.

Screens are available in diameters ranging from 32 mm (1 1/4 in) to 1,524 mm (60 in). Screen diameter should be selected on the basis of the desired yield from the well and thickness of the aquifer. Taking into account that the entrance velocity of the water generally must not exceed the design standard of 0.03 m/sec (0.1 ft/sec), the screen diameter can be adjusted within narrow limits after length of the screen and size of the screen openings have been selected in accordance with the thickness of the aquifer and with the gradation of the sediments or the size of the filter pack, respectively.

In a naturally developed well, the usual approach is to select a slot screen which will allow 50–40 percent of the material to pass, retaining 50–40 percent. This practice results in creating a zone of graded grain distribution extending over a distance of 0.3 to 0.6 m near the borehole. The finer formation materials brought into the screen are pumped from the well during development.

In filter-packed wells, the annular space between the screen and the borehole wall is filled with specially graded material. The principal functions of a gravel pack are: (1) to stabilise the aquifer formation and minimise sand pumping, (2) to permit use of the largest possible screen slot with resultant maximum open area, and (3) assure good hydraulic conductivity, thus increasing the effective radius of the well and the yield (Anderson, 1992; Driscoll, 1989; Lehr et al., 1987; Roscoe, 1985; USDI, 1981).

The average entrance velocity is calculated by dividing the desired well yield by the total area of the screen openings. If the velocity is greater than 0.03 m/sec, the screen length and/or diameter should be increased to provide enough open area so that the entrance velocity is 0.03 m/sec or less. Table 7.9 gives open areas in square inches per foot of length of some representative screens of various slot sizes and diameter (Driscoll, 1989; Lehr et al., 1987; USDI, 1981).

The transmitting capacity of a well screen section, expressed as US gallons per minute (gpm)/ft of screen (0.7 m^3/h per metre screen) at the recommended velocity of 0.1 ft/s (0.03 m/s),

Table 7.9 Open areas for some representative screens

*Screen diameter**	*Slot size*	*Continuous slot*		*Bridge slot*		*Mill slotted*		*Plastic continuous*		*Plastic slotted*	
mm (in)	*10^{-3} in*	*in/ft*	*%*	*in/ft*	*%*	*in/ft*	*%*	*in/ft*	*%*	*in/ft*	*%*
102 ID	20	44	25	–	–	–	–	22	13	–	–
(4)	60	90	52	19	12	8	5	52	30	18	11
203 ID	30	80	25	–	–	–	–	57	18	26	8
(8)	60	135	41	17	6	15	5	93	29	47	14
	95	165	51	–	–	23	7	–	–	–	–
305 ID	30	77	16	12	3	–	–	–	–	–	–
(12)	60	135	28	33	7	21	5	–	–	52	11
	95	182	38	–	–	32	7	–	–	–	–
	125	214	45	68	14	43	9	–	–	–	–
406 OD	30	97	16	16	3	–	–	–	–	52	9
(16)	60	169	28	35	6	27	5	–	–	–	–
	95	228	38	–	–	41	7	–	–	–	–
	125	268	45	78	13	55	9	–	–	–	–

* For example, an 8-in continuous-slot well screen has slot openings of 0.030, 0.060 or 0.095 in.

is calculated readily from the open area figures shown in Table 7.9. Multiplying the number of square inches of open area by a conversion factor of 0.31 gives the transmitting capacity at a velocity of 0.03 m/s (0.1 ft/s). The unit conversion factor of 0.31 results from specifying an entrance velocity, V, of 0.03 m/s (0.1 ft/s) in the equation Q = VA. For example, the transmitting velocity of a 12-in diameter screen may vary from 77 x 0.31 = 24 gpm/ft up to 214 x 0.31 = 66 gpm/ft, or 16 m^3/h per metre up to 45 m^3/h per metre.

7.8 Well design in deep confined aquifer systems

Special well designs have been developed in certain areas because of particular hydrogeological conditions. These advanced methods include the type of drilling equipment used, the filter pack injection techniques, and economic aspects of the wells. The technologies and procedures have been developed because more water is needed than could obtained with standard design criteria and because favourable cost/benefit ratios have been realised. Although developed locally, these technologies can be used successfully in areas with similar social, economic and hydrogeological conditions. Several examples of alternative well designs are described below.

7.8.1 Hydrogeological conditions

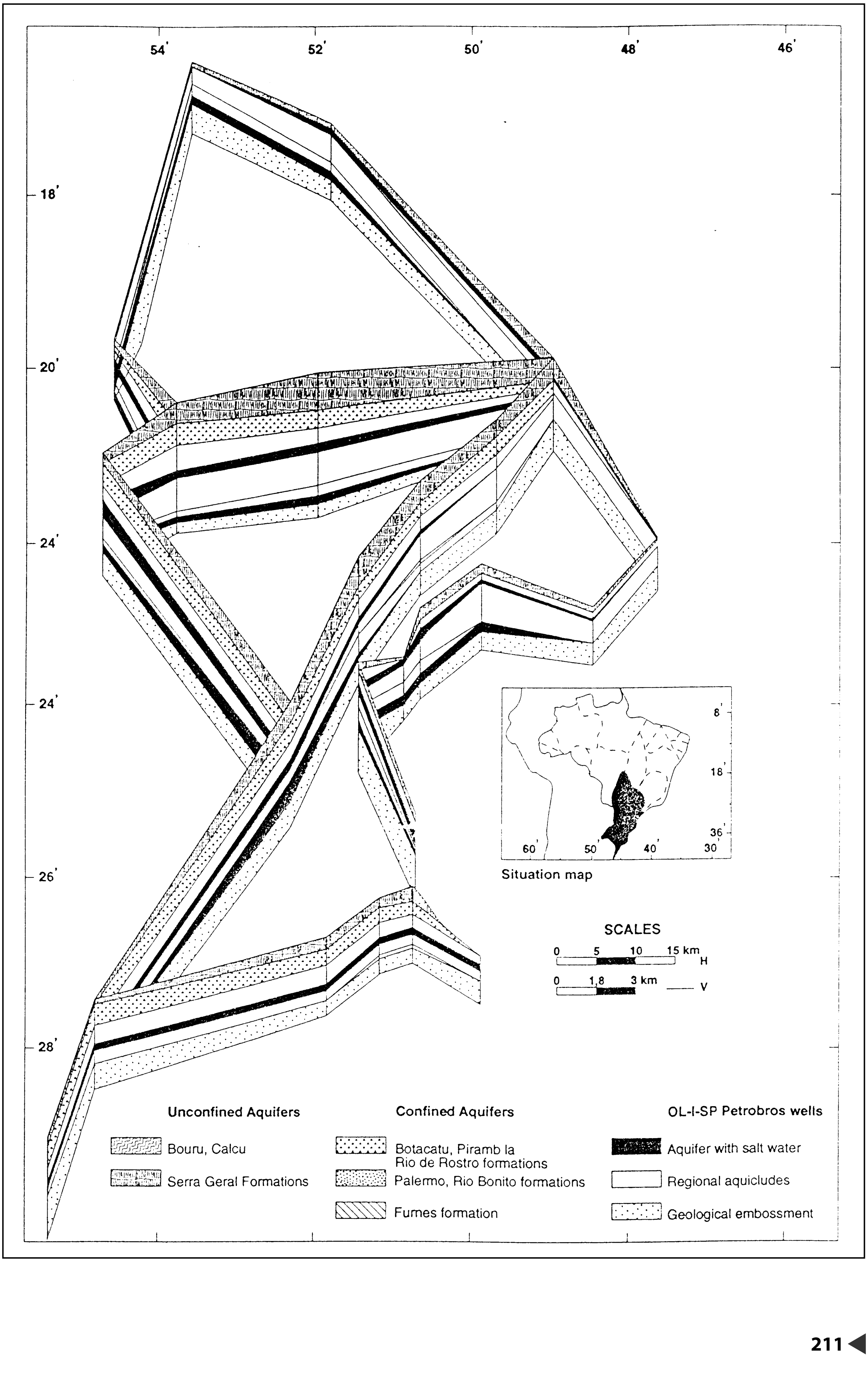
Underlying most of the developed region of South America, covering about 1,000,000 km^2 in Brazil, 100,000 km^2 in eastern Paraguay, 100,000 km^2 in north-western Uruguay and 400,000 km^2 in northeastern Argentina is the Paraná Sedimentary Basin (Rebouças, 1976, 1994b). The basin is up to 5,000 m thick, and forms a large intracratonic structure uplifted and exposed along its eastern margin and tilted southwest (Figure 7. 11).

The sedimentary sequence from the Silurian to Cretaceous is almost undisturbed, comprising a multi-layered aquifer system with gentle dips towards the centre of the basin. Regional and local faults served as channels for extruding basalt flows during the Cretaceous period. Currently the basaltic package extends over about 1,000,000 km^2 and has a maximum thickness of up to 2,000 m in the centre of the sedimentary basin. Groundwater occurs within the interflow zones and along cooling joints; interflow sediment deposits greatly increase these local and occasional basalt water-bearing conditions. Most commonly the vertical permeability is very small in comparison to the horizontal permeability; indeed the basalt flow package is the hydrogeological substratum for groundwater stored in the overlying Cretaceous sandstone formations.

In the Paleozoic sequence, with aquifers occurring in quartzose sandstones of continental and marine origin, and confining beds of siltstone and marine argillaceous sediments, groundwater usually has a poor quality.

More important as a source of good quality groundwater is the Botucatu aquifer system which includes the Triassic deposits (Piramboia and Rosario do Sul formations in Brazil and Buena Vista in Uruguay) and Jurassic sandstones accumulated by aeolian processes under desert conditions (Botucatu formation in Brazil, Misiones formation in Paraguay, and Tacuarembó formation in Uruguay and Argentina). This aquifer system covers about 1,200,000 km^2 (840,000 km^2 in Brazil, around 70,000 km^2 in Paraguay, 230,000 km^2 in Argentina, and 60,000 km^2 in Uruguay). As a result of the differential subsidence, the total thickness of both the fluvial-lacustrine and aeolian deposits varies widely; from more than 800 m near the southwestern Brazilian border to complete absence in some limited areas in the subsurface. But in general terms, thicknesses of more than 500 m tend to prevail along a NNE-SSW axis which is followed by the Paraná and Uruguay rivers today. The Botucatu aquifer system is confined by the Cretaceous basalts of the Serra Geral Formation and by underlying Permo-Triassic deposits of

Figure 7.11 Fence diagram of the Paraná sedimentary basin (Rebouças, 1994b)



low permeability resulting in artesian conditions over about 70 per cent of the total area it underlies. Jurassic aeolian sandstones form the best part of the aquifer system (hydraulic conductivity ranging between 0.3 to 5.0 m/day), whereas more argillaceous fluvial-lacustrine Triassic deposits are notably inferior, with hydraulic conductivity values an order of magnitude less. Not surprisingly in view of the great depths it reaches (up to 3,000 m) and the thick confining basaltic cover, the water stored in the Botucatu aquifer has temperatures exceeding 55°C in its deepest parts.

At the end of the Cretaceous the present structural configuration was largely established and, since then, the aquifer has almost everywhere been flushed by infiltrated rain water. Natural direct recharge occurs in the narrow eastern and western uplifted marginal unconfined zone. In the confined context indirect recharges and discharges may occur through the basaltic package, in view of the hydraulic head balances. Groundwater flow patterns and the continuing recharge from geological to modern times have been confirmed by environmental isotope and hydrochemical studies.

Throughout the vast area of this aquifer system, most of its water is potable. Few exceptions occur where water quality is affected by enrichment with fluorine. The highest values reaching 12 mg/l are related to areas of stagnant water or to the lacustrine deposits of the Piramboia formation. As a result, it is frequently recommended that wells penetrate only partially, because high fluorine contents may occur in the lowest sections of the aquifer system, and because there may be some flow of low quality water from the basal Permo-Triassic sequence into the Botucatu aquifer system, which may be increased by the pumping effects (Rebouças, 1976, 1991, 1994a, 1994b).

7.8.2 Typical well drilling procedures

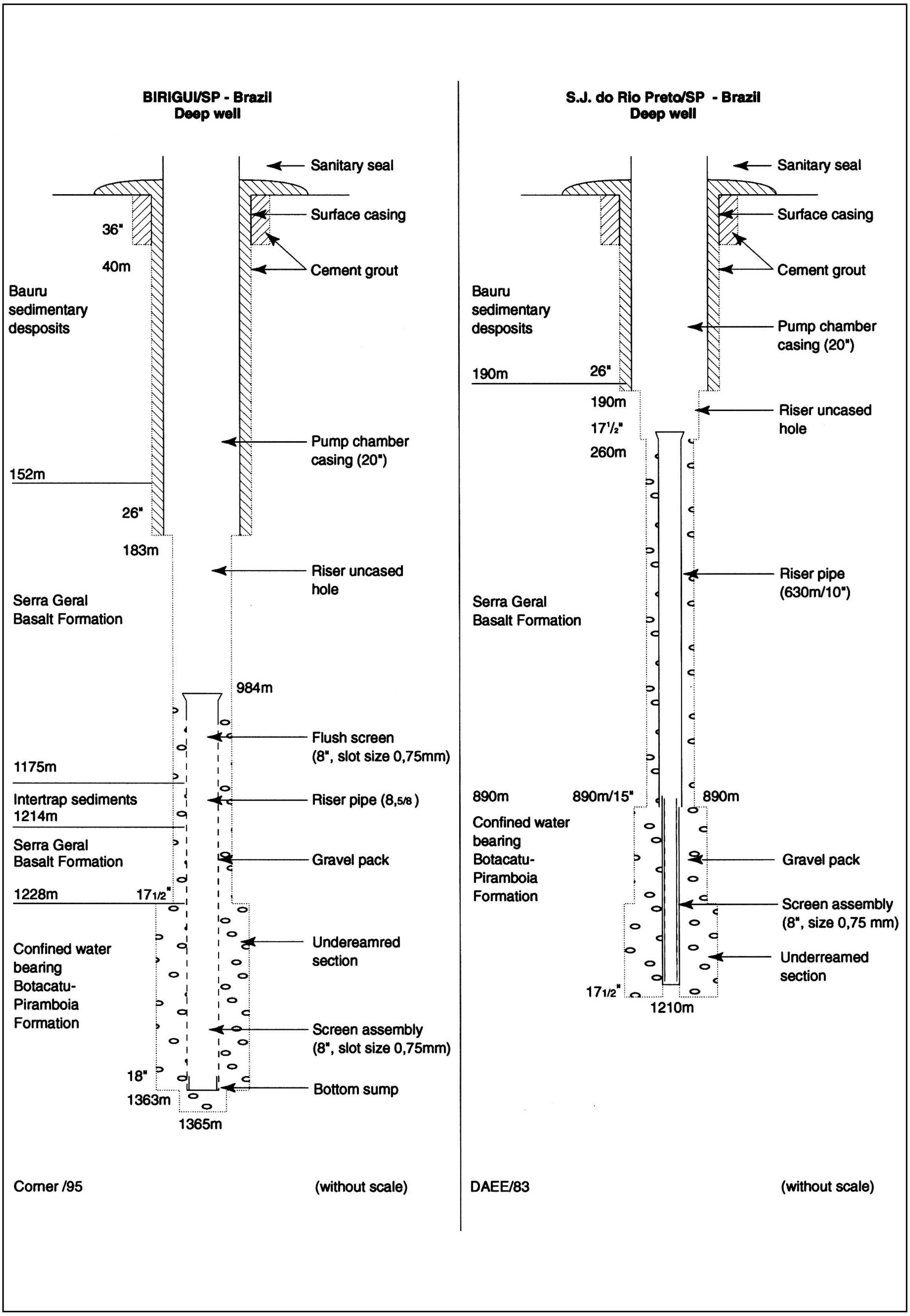
Despite being situated in a region with abundant surface water resources, the Botucatu aquifer is increasing in importance as a source of regional water supplies, partly as a response to the growing costs and other constraints in storing and treating surface water and partly because the advantages of groundwater are now better understood. Currently, about 1,000 wells pump groundwater from the Botucatu aquifer system for domestic water supply, industries and thermal water recreational clubs.

The typical drilling problems result from the accumulation of particles in the fluid system and at the bottom of the borehole, slumping or expansion of active clays which occurs at the interflow zones of the thick basaltic package, and caving-in of the inter-basalt sand deposits. Drilling efficiency depends largely on the driller's experience, drilling fluid control, and well completion steps that are taken during installation of well screens, or immediately thereafter, including filter pack installation, grouting the casing, and developing the well. A good filter packing installation is especially important because the aeolian sediments of the Botucatu Formation are very uniform and fine grained, and because the small slot size dictated by natural development limits the transmitting capacity of the screen, so that the desired yield cannot be obtained. In view of the depths of the wells (up to 2,000 m) and the high artesian pressure, filter packing installation is a laborious and highly costly operation. Figure 7.12 shows various well designs.

A typical procedure for constructing a high capacity well in the Botucatu aquifer system includes the following steps:

- drill a pilot hole at the chosen site, keeping a detailed drilling log,
- conduct geophysical logs (typically SP, resistivity, natural gamma ray, and calliper) in the pilot hole to determine the physical features of the basaltic package, the presence of interflow sandstones, and the hydrogeological characteristics of the Botucatu sandstone. Depending on the physical features of the basalt section between the bottom of

Figure 7.12 Typical water well designs to tap the confined Botucatu aquifer system in Paraná basin, Brazil



the recommended pump chamber casing and the top of the aquifer the well will be either left uncased or lined,

- after the well design has been chosen from analysis of the driller's logs, and geophysical logs, an open hole is drilled to receive the surface casing which is placed from near the ground surface, through the unstable Bauru sandstone deposits and a short distance into a firm and impermeable basalt, and grouted,
- at least 24 hours thereafter the drilling should be continued with a diameter suitable to receive the pump chamber casing which is set in the hole and grouted,
- the borehole section between the bottom of the pump chamber casing and the top of the aquifer is then drilled according the recommended diameters, to reach the desired well yield and/or to accommodate the proper riser pipes,
- to accommodate a proper thickness of filter pack between the screen and formation, the drilled thickness of the aquifer is under-reamed to obtain a 76 to 203 mm (3 to 8 in) annulus space,
- occasionally, a riser pipe assembly is telescoped into the open hole, between the bottom of the pump chamber casing and the top of the aquifer, and then grouted,
- the screen assembly is then telescoped through the casing into the under-reamed aquifer section, and the filter pack is introduced into the well by reverse circulation methods,
- after the filter pack has been installed, development work is continued to remove fine sediment from the filter pack and to clean the contact surface between the filter pack and the water-bearing formation.

In view of the varied diagenetic processes affecting the Botucatu water-bearing formation, the specific capacity of the wells ranges from 5 to 25 m^3/h per metre of water-level drawdown. Currently, the production cost per cubic metre of water varies from US$ 0.10 to US$ 0.50, assuming a discharge of about 500 m^3/hour and a pumping regime of 16 hours per day.

7.9 References and additional reading

AHRENS, T. P., 1970. Basic Considerations of Well Design. *Water Well Journal* (United States of America), Vol. 24, Nos. 4, 5, 6 and 8.

ALLER, L.; BENNETT; T. W.; HACKETT; G.; PETTY, R. J.; LEHR, J. H.; SEDORIS, H.; NIELSEN, D. M., 1989. *Handbook of Suggested Practices for the Design and Installation of Ground-Water Monitoring Wells*. National Water Well Association, Ohio, 380 pp.

ANDERSON, K., 1992. *Ground Water Handbook*. 401 pp.

ASTM, 1992. *Standards on Ground Water and Vadose Zone Investigations*. American Society for Testing and Materials, 166 pp.

AWWA, 1984. *Standard for Water Wells*. American Water Works Association, Denver, CO., A 100–84, 75 pp.

BOWMAN, I., 1911. Well-drilling methods. *U.S. Geol. Survey*, Water Supply Paper 257, pp. 23–30.

CAMPBELL, M. D.; LEHR, J. H., 1973. *Water Well Technology.* McGraw-Hill, New York, 681 pp.

CARPENTER, D., 1994. *Wellhead Protection – Anthology*. NGIC, 13 papers.

COHEN, PH., 1985. Ground-Water Development in the United States of America. *Proc. IAH 18th Congress, Vol. XVIII, Part 1, Cambridge, United Kingdom*, pp. 17–30.

DAO, 1965. *Technical Manual 5–297, Well Drilling*. Department of the Army Operations, Washington, D.C.

DRISCOLL, F. G., 1989. *Ground Water and Wells*. Johnson Filtration Systems Inc., Second Edition, United States of America, 1089 pp.

EPA, 1975. *Manual of Water Well Construction Practices*. EPA 570/9-75-001, U.S. Environmental Protection Agency, Office of Water Supply, Washington, D.C., 156 pp.

——, 1980. *Proposed Ground Water Protection Strategy*. U.S. Environmental Protection Agency, Office of Drinking Water, Washington, D.C., 540 pp.

FETTER, C. V., 1993. *Contaminant Hydrogeology*. Macmillan, United States of America, 458 pp.

GORDON, W. R., 1958. *Water Well Drilling with Cable Tools*. Bucyrus-Erie Co., United States of America, 256 pp.

HABERMEHL, M. A., 1985. Groundwater in Australia. *Proc. IAH 18th Congress, Vol. XVIII, Part 1. Cambridge, United Kingdom,* pp. 31– 52.

HARDCASTLE, B. J., 1987. Wells Ancient and Modern; An Historical Review. *Quarterly Jour. of Eng. Geol.,* pp. 231–238.

KOERING, 1967. *Well Drilling Manual*. Speed Star Division of Koering Co., Enid, Okla.

LEHR, J.; HURLBURT, S.; GALLAGHER, B.; VOYTEK, J., 1987. *Design and Construction of Water Wells; A Guide for Engineers*, NGWA, United States of America, 256 pp.

MARGAT, J., 1990. Les gisements d'eau souterraine. *La Recherche,* No. 221, Paris, pp. 590–6.

MEINZER, O. E., 1934. The History and Development of Groundwater Hydrology. *Jour. Washington Acad. Sci.,* Vol. 24, pp. 6–23.

NGWA, 1965. *Well Drilling Operations by U.S. Dept. of Army/Air Force*. National Groundwater Association, United States of America, 249 pp.

——, 1985. *Water Well Drillers, Beginners Training Manual*. National Water Well Association, Columbus, Ohio, 245 pp.

NWWA AND PPI, 1981. *Manual on the Selection and Installation of Thermoplastic Water Well Casing.* National Water Wells Association and Plastics Pipe Institute, Worthington, Ohio, 64 pp.

OECD, 1989. *Water Resource Management, Integrated Policies*. Organisation for Economic Co-operation and Development, Paris, 199 pp.

REBOUÇAS, A. DA C., 1976. *Recursos hídricos subterrâneos da Bacia do Paraná; análise de pré-viabilidade*. USPO Tese de Livre Docência, São Paulo, 143 pp.

——, 1988. *Groundwater in Brazil, Episodes,* Vol. 11, No. 3, pp. 209–14.

——, 1991. *Desarrollo y tendencias de la hidrogeología en América Latina*. 25th Anniversary of the Curso Internacional de Hidrología Subterránea, Barcelona, pp. 429–53.

——, 1994a. Gestão sustentável dos grandes aquíferos. *Anais 8° Cong. Bras. Àguas Subterrâneas, ABAS-Recife, Brazil*, pp. 131–9.

——, 1994b. Sistema aquífero Botucatu no Brazil. *Anais 8° Cong. Bras. Àguas Subterrâneas, ABAS–Recife, Brazil*, pp. 500–9.

ROSCOE, M., 1985. *The Engineer's Manual for Water Well Design*. NGWA, United States of America, 242 pp.

SMITH, A. J., 1980. Design as the Key to a Good Recharge Well. *Johnson Drillers's Jour. Johnson Division,* MN, United States of America.

SOLLEY, W. B.; PIERCE, R. R.; PERLMAN, H. A., 1993. *Estimated Use of Water in the United States in 1990*. U. S. Geological Survey Circular 1081.

TOLMAN, C. F., 1937. *Ground Water*. McGraw-Hill Book Co., New York, 593 pp.

UNESCO, 1992. *Ground Water*. United Nations Educational, Scientific and Cultural Organization, Paris.

USDI, 1981. *Ground Water Manual*. US Departement of the Interior, Water and Power Resources Service, A Wiley-lnterscience Pub., 480 pp.

WRIGHT, E. P.; HERBERT, R.; KITCHING, R.; MURRAY, K. H., 1987. Collector Wells in Crystalline Basement Aquifers; Review of Results of Recent Research. *Proc. XXl Congress of IAH, Rome, Italy,* pp. 12–7.

8 Determining hydrodynamic and contaminant transfer parameters of groundwater flow

8.1 Introduction

This chapter briefly describes of the methods to determine and quantify the hydraulic characteristics of homogeneous and isotropic aquifers by pumping tests (section 8.3 and 8.4) and by other methods (section 8.5) and to determine the hydraulic characteristics of fractured aquifers, volcanic rocks and limestones (sections 8.6, 8.7, and 8.8). Diffusion, dispersion and macrodispersion are discussed in section 8.9.

8.2 Parameters of flow in homogeneous aquifers

In the following sections a broad distinction will be made between the aquifers with primary porosity and those with secondary porosity. Primary porosity is porosity related to the sedimentation process and hence related to the diameter of the pores and their distribution in the original sedimentary deposits, or it is related to the formation of igneous and volcanic rocks. Secondary porosity is related to post sedimentary or post formation processes, like folding, fracturing, faulting and shrinkage; a particular group of hard rock aquifers are those in karstic limestone in which solution porosity may play a major role.

The commonly used parameters of fresh water flow (see also Section 1.4) are:

- the hydraulic conductivity K $[L^1 T^{-1}]$;
- the transmissivity T = KM $[L^2T^{-1}]$, where M is the saturated thickness of the aquifer;
- the transmissivity for vertical flow through an aquitard $T_v' = K_v'/M'$ $[T^{-1}]$; the reciprocal of this value is known as the hydraulic resistance of the aquitard: c [T];
- the storage coefficient of a confined aquifer S_A [dimensionless],
- the specific yield of an unconfined aquifer S_Y [dimensionless].

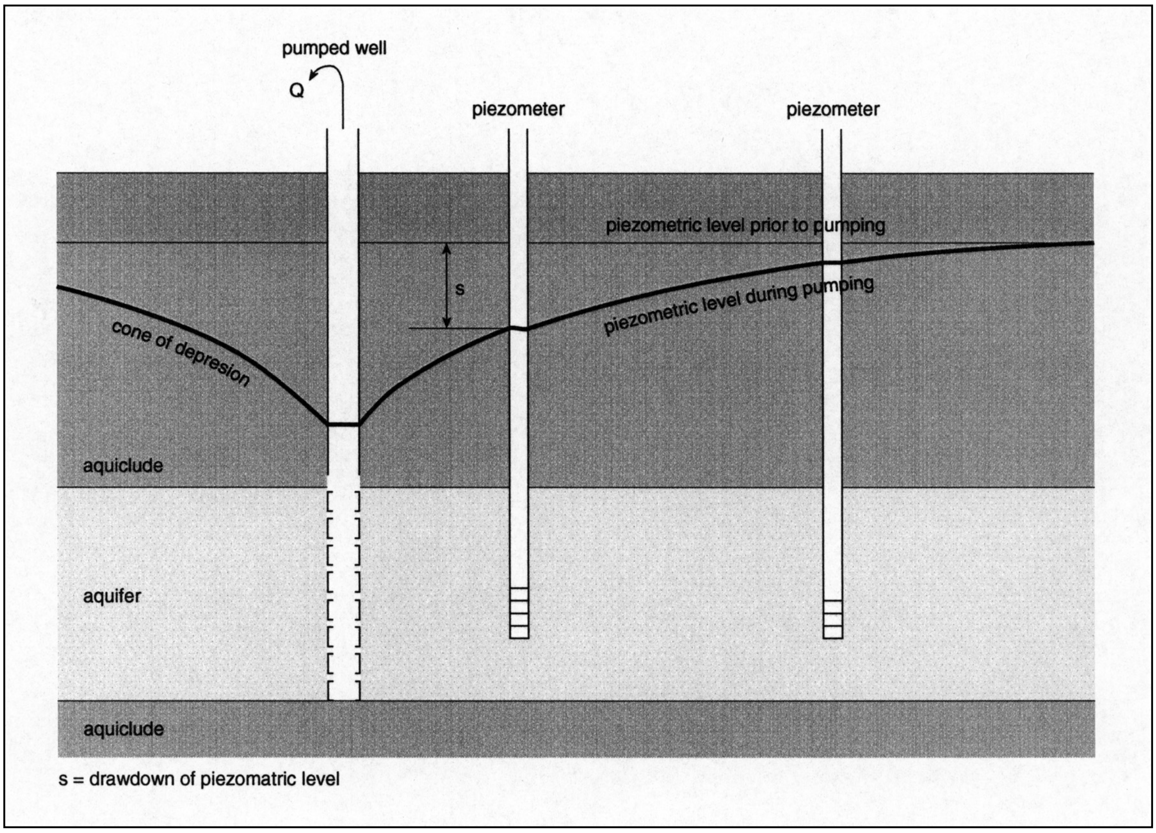
8.3 Determination of hydraulic characteristics by pumping tests

8.3.1 The pumping test

Pumping tests are a commonly used method to determine the transmissivity (and, if the thickness of the aquifer is known, the hydraulic conductivity) and the storage coefficient of an aquifer. The hydraulic resistance of a semi-confining layer of a leaky aquifer can also be calculated from pumping test data.

The principle of a pumping test is that if we pump water from a well and measure the discharge of the well and the resulting drawdown in the well and in piezometers at known distances from the well and at a specific time, we can substitute these measurements into an appropriate well flow equation and calculate the hydraulic characteristics of the aquifer (Fig. 8.1).

Figure 8.1 Drawdown in a pumped confined aquifer



Before a pumping test is conducted, the following geological and hydrological information should be known:

- the geological characteristics of the subsurface (i.e. all the lithological, stratigraphical and structural features that may influence the flow of groundwater);
- the type of aquifer and the presence or absence of horizontal recharge boundaries, e.g. where percolating rain or irrigation water causes the water table of an unconfined aquifer to rise, or where water passes through an aquitard and recharges a semi-confined (leaky) aquifer;
- the thickness and lateral extent of the aquifer and confining beds as found by geophysical surveys or from borehole data;
- the boundary conditions. The aquifer may be bounded laterally by barrier boundaries of impermeable material (e.g. the bedrock sides of a buried valley, a fault, or simply lateral changes in the lithology of the aquifer material). Analogous effects may occur when the aquifer has one or more recharge boundaries (e.g. a deeply incised perennial river or canal, a lake, or the ocean). These conditions have to be taken into account if the drawdown cone around a pumping test reaches one or more lateral boundaries;
- data on the groundwater flow system (e.g. water table gradients and regional trends in groundwater levels);

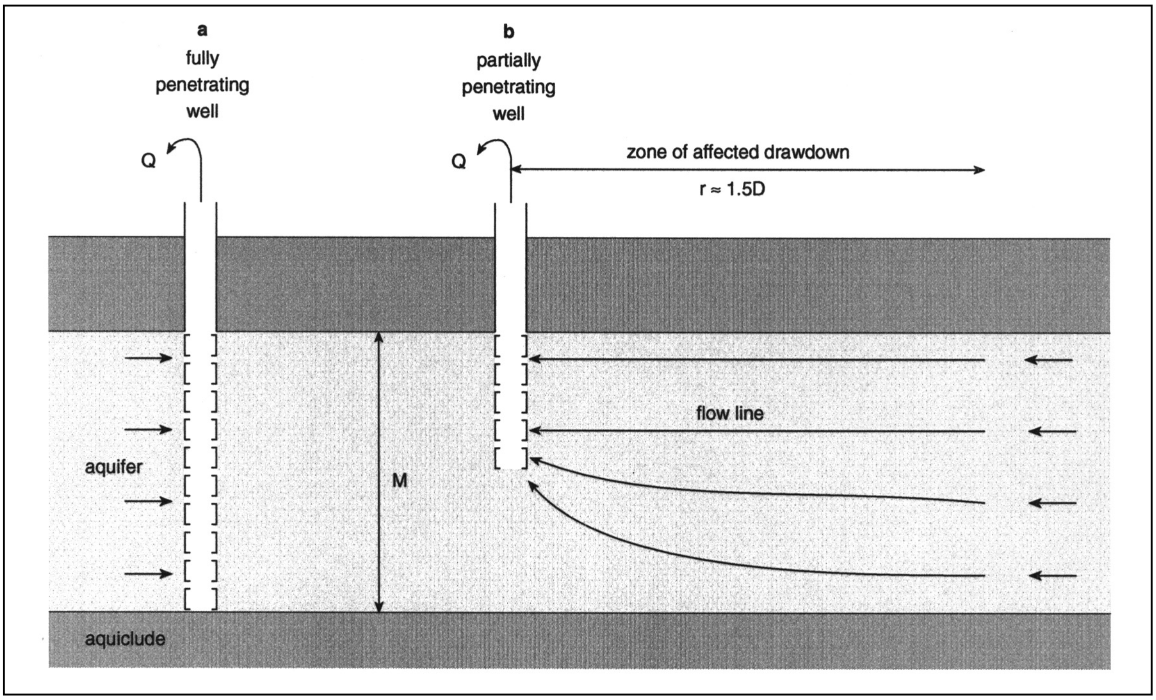
- Any existing wells in the area. From the logs of these wells, it may be possible to derive approximate values of the aquifer's transmissivity and storativity and their spatial variation. It may even be possible to use one of these wells for the test, thereby reducing the cost of fieldwork. Often, however, such a well produces unreliable results because no details are available on its construction and condition.

8.3.2 The well

The pumping test site should be selected taking into account the following points:

- The hydrogeological conditions should be representative of most, if not all of the area;
- The site should be away from railways and motorways where passing trains or heavy traffic might produce measurable fluctuations in the hydraulic head of a confined aquifer;
- The site should not be in the vicinity of existing discharging wells;
- The water produced during the test should be discharged in a way that prevents its return to the aquifer;
- The gradient of the water table or piezometric surface should be small;
- Manpower and equipment must be able to reach the site easily;
- The well should in principle be drilled to the bottom of the aquifer and be screened over 80 percent of the aquifer thickness This makes it possible to obtain about 90 percent of the maximum yield. Moreover with this screen length the groundwater flow to the well can be assumed to be horizontal, an assumption that underlies almost all well flow equations (Fig. 8.2). Exceptions to these rules are:
  - In unconfined aquifers, it is common practice to screen only the lower half or lower one third of the aquifer because, if appreciable drawdown occur, the upper part of a longer well screen would fall dry;

Figure 8.2 Well penetration: a) a fully penetrating well; b) a partially penetrating well



- In a very thick aquifer it will be obvious that the length of the screen will have to be less than 80 per cent of the aquifer thickness. Such a well is said to be partially penetrating. It induces vertical-flow components, which can extend outwards from the well to distances roughly equal to 1.5 times the thickness of the aquifer (Fig. 8.2).

After the well has been drilled, screened and gravel-packed, as necessary, a pump and power unit are installed. They should be capable of operating continuously at a constant discharge for a period of at least a few days. An even longer period may be required for unconfined or leaky aquifers and especially for fractured aquifers. The same applies if drawdown data from piezometers at great distances from the well are to be analysed. The capacity of the pump and the rate of discharge should be high enough to produce good measurable drawdown in piezometers as far away as, say, 100 or 200 m from the well, depending on the aquifer conditions.

8.3.3 The piezometers

A piezometer is an open-ended pipe installed in a borehole that has been drilled to the desired depth below surface. The bottom tip of the piezometer is fitted with a perforated or slotted screen of 0.5 to 1 m length to allow the inflow of water. The annular space around the screen should be filled with a gravel or uniform coarse sand to facilitate the inflow of water. The rest of the annular space can be filled with any material available, except where the presence of aquitards requires a seal of bentonite clay, very fine sandy clay or cement grouting to prevent leakage along the pipe. The water level measured in the piezometer represents the average head at the screen of the piezometer.

The number of piezometers to be installed depends on the amount of information needed, and especially on its required accuracy, but also on the funds available. The distance at which piezometers should be installed depends on the type of aquifer, its transmissivity, the duration of pumping, the discharge rate, the penetration rate of the screen, and whether the aquifer is stratified (Table 8.1).

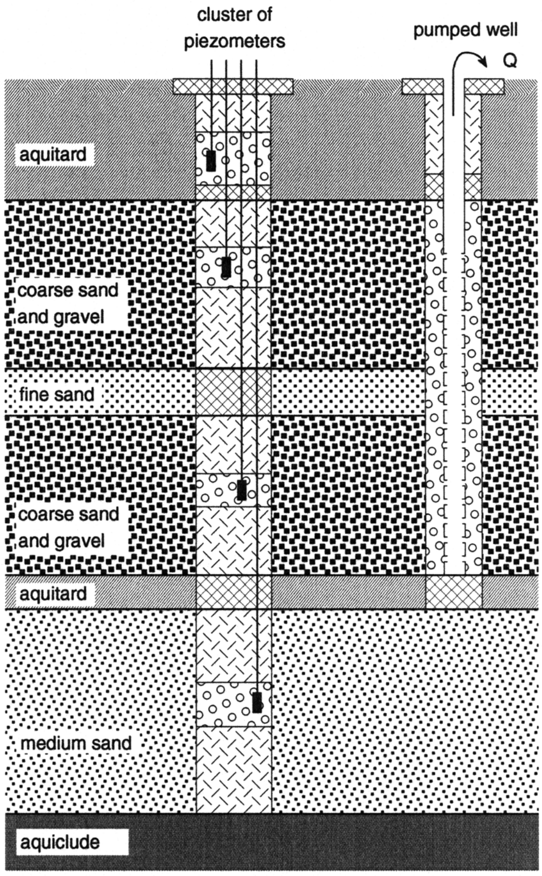
The depth at which a piezometer is installed depends on the aquifer type and the homogeneity of the aquifer. In an isotropic and homogeneous aquifer the piezometer should be

Table 8.1 Preferred distance between pumping well and piezometer in relation to characteristics of the pumping test

Distance	*Near* <——————————	—————> *Far*
Type of aquifer	Unconfined	Confined
Transmissivity	Low	High
Pumping time	Short	Long
Discharge rate	Low	High
Penetration rate	Full	Partial[1]
Stratification	Little[2]	Strong

Notes:
1. The drawdown measured at a distance less than 1.5 times the thickness of the aquifer must be corrected for the influence of the vertical flow components close to the well.
2. As a consequence of the differences in transmissivity at different depths, the drawdown observed close to the well may differ at different depths within the aquifer. With increasing pumping time and increasing distance to the well, the effect of stratification upon the drawdown diminishes.

Figure 8.3 Cluster of piezometers in a heterogeneous aquifer intercalated with aquitards



installed at a depth that is half the length of the well screen. For aquifers made up of sandy deposits with intercalations of less pervious layers it is recommended to install a cluster of piezometers (Fig. 8.3). If the aquifer is overlain by a partly saturated aquitard (leaky aquifer) piezometers should also be installed in the aquitard to check whether the water table is affected when the underlying aquifer is pumped.

8.3.4 The measurements

During a pumping test, measurements are taken of the water levels in the well and the piezometers, of the discharge rate, and of the quality (e.g. the Electrical Conductivity) of the pumped water.

Measurements of the water levels (e.g. twice a day) should start a few days before the start of the pumping test, to establish the natural changes in hydraulic head in the aquifer. This applies to both long term trends as well as to short term variations (e.g. tidal movements in coastal aquifers, or day and night variations). When a test is expected to last one or more days, measurements should also be made of the atmospheric pressure, the levels of nearby surface waters, and of any precipitation or irrigation. Try to avoid the disturbing influence of nearby production wells.

During the test, the water levels in the well and in the piezometers must be measured many times. Because the water levels fall rapidly during the first one or two hours of the test, the readings should be made at frequent intervals. Tables 8.2 and 8.3 give ranges of readings in the well and in the piezometers close to the well, respectively.

For piezometers further from the well and for those in confining layers above or below the aquifer, the intervals in the first minutes need not be so short.

After the pump has been shut down, the water levels in the well and the piezometers will start to rise; rapidly in the first hour, but more slowly afterwards. The rising water levels are measured as residual drawdown during a period that is as long as the pumping period. These data can be used to check the calculations made on the basis of the date collected during the pumping period. The schedule for the recovery measurements should be the same as that adhered to during the pumping period (Tables 8.2 and 8.3). The most accurate recordings of water level changes are made with fully-automatic microcomputer-controlled systems that use pressure transducers or acoustic transducers for digitised recordings; the data are automatically stored on micromemory from which they can be retrieved by a laptop computer. A good alternative is the conventional mechanical recorder, which produces a continuous recording of water-level changes. Fairly accurate water-level measurements can be taken by hand, using e.g. an electrical sounder. For detailed descriptions of automatic recorders see e.g. Driscoll (1986), Genetier (1984), Groundwater Manual (1981).

Table 8.2 Frequency of water level measurements in the pumped well

Time since start of pumping	*Time intervals*
0–5 minutes	0.5 minutes
5–60 minutes	5 minutes
60–120 minutes	20 minutes
120 minutes – shutdown of the pump	60 minutes

Table 8.3 Frequency of water level measurements in piezometers

Time since start of pumping	*Time intervals*
0–2 minutes	approx. 10 seconds
2–5 minutes	30 seconds
5–15 minutes	1 minute
15–50 minutes	5 minutes
30–100 minutes	10 minutes
100 minutes–5 hours	30 minutes
5–48 hours	60 minutes
48 hours–shutdown of the pump	3 times a day

The discharge rate can be measured with a commercial water meter of appropriate capacity. If the water is discharged through a small ditch, a flume can be used to measure the discharge. Other fairly simple methods are: container of known capacity in combination with a chronometer, orifice, orifice bucket or jet-stream method. For details and tables see Driscoll (1986).

8.4 Analysis of pumping test data on homogeneous and isotropic aquifers

8.4.1 Data analysis

It is relatively easy to calculate hydraulic characteristics if the aquifer system (i.e. aquifer plus well) are precisely known. This is generally not the case, so interpreting a pumping test is primarily a matter of identifying an unknown system. System identification includes the construction of diagnostic plots and specialised plots. Diagnostic plots are log-log plots of the drawdown versus time since pumping started. Specialised plots are semi-log plots of drawdown versus time, or drawdown versus distance to the well; they are specific for a given flow regime.

Both plots must be constructed, because the diagnostic value lies in the typical combination of the log-log and semi-log plots (Fig. 8.4). The choice of a theoretical model is a crucial step in the interpretation of pumping tests. If the wrong model is chosen, the hydraulic characteristics calculated for the real aquifer will not be correct. Unfortunately the theoretical solutions of well flow problems are not unique. Some models developed for different aquifer systems, yield similar responses when required to handle a given stress. This means that besides the log-log and semi-log plots of the drawdown versus time, all other relevant hydrogeological information, e.g. lithology, boundary conditions, should be taken into account.

8.4.2 The well flow formula for confined aquifers

Theis (1935) noted that when a well penetrating an extensive confined aquifer is pumped at a constant rate, the influence of the discharge extends outward with time. The rate of decline in head, multiplied by the storativity and summed over the area of influence, equals the discharge. Theis' equation, which was derived from the analogy between the flow of groundwater and the conduction of heat, is written as:

$$s = \frac{Q}{4\pi KM} W(u) \qquad (8\text{-}1)$$

where:

- s = the drawdown in m measured in a piezometer at a distance r in m from the well;
- Q = the constant well discharge in m^3/d;
- KM = the transmissivity of the aquifer in m^2/d;
- W(u) = is a particular exponential integral, with u as the argument, which in this usage is read as the well function of u.

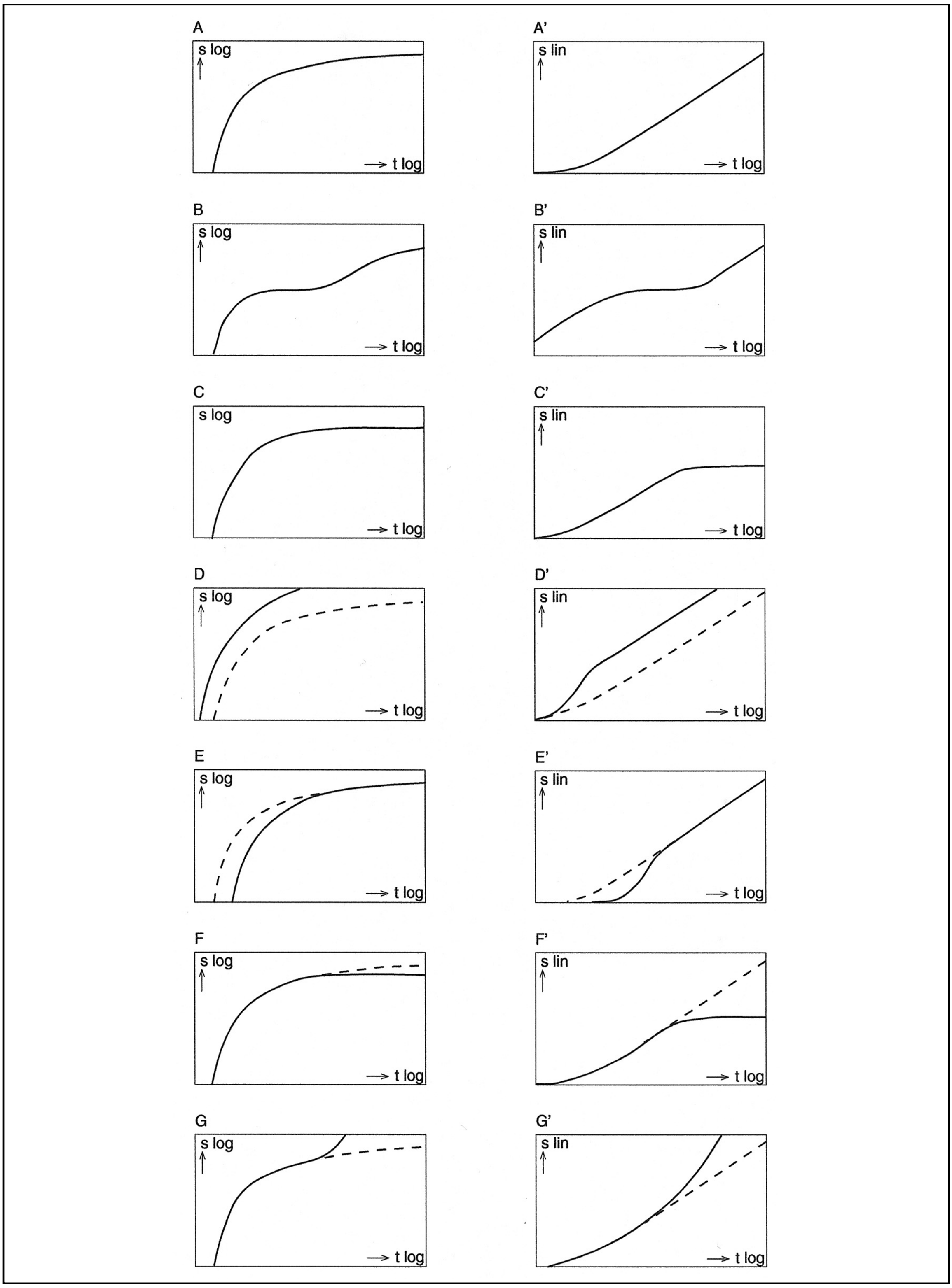
$$u = \frac{r^2 S}{4KMt} \quad \text{and consequently} \quad S = \frac{4KMtu}{r^2} \qquad (8\text{-}2)$$

S = the storage coefficient;
t = the time since pumping started.

W(u) can be written as a series:

$$W(u) = -0.5772 - \ln u + u - \frac{u^2}{2.2!} + \frac{u^3}{3.3!} - \frac{u^4}{4.4!} + \ldots \qquad (8\text{-}3)$$

Figure 8.4 Log-log and semi-log curves of drawdown versus time; A and A': confined aquifer, B and B': unconfined aquifer, C and C': leaky (or semi-confined) aquifer; D and D': effect of partial penetration, E and E': effect of well-bore storage (large diameter well), F and F': effect of recharge boundary, G and G': effects of an impervious boundary



In fact the equation of Theis describes the flow to a well in an aquifer system with the following characteristics:

- a confined aquifer of seemingly infinite areal extent that is homogeneous and isotropic, and uniformly thick over the area influenced by the test;
- a piezometric surface that is horizontal (or nearly so) prior to pumping, over the area that will be influenced by the test;
- the aquifer is pumped at a constant discharge rate;
- the well penetrates the entire thickness of the aquifer and thus receives water by horizontal flow;
- the water removed from storage is discharged instantaneously with decline of head;
- the diameter of the well is small, i.e. the storage in the well can be ignored.

From Eq. (8-1), it will be seen that, if s can be measured for one or more values of r and for several values of t, and if the well discharge Q is known, S and KM can be determined.

8.4.3 Theis's curve-fitting method

The presence of the two unknowns and the nature of the exponential integral in Theis' equation makes it impossible to effect an explicit solution; however, a solution can be found by a curve fitting method. Eq. (8-1) can be written as:

$$\log s = \log (Q/4\pi KM) + \log (W(u))$$

and Eq. (8-2) as

$$\log (t/r^2) = \log 4KM/S + \log (1/u)$$

Since $Q/4\pi KM$ and $4KM/S$ are constants, the relation between log s and log (t/r^2) must be similar to the relation between log W(u) and log (1/u). Values for W(u) as 1/u varies have been tabulated and are widely published, e.g. by Kruseman et al. (1991).

The curve fitting method is based on the fact that if s is plotted versus t/r^2 and W(u) versus 1/u on the same log-log paper the resulting curves (the data curve and the type curve, respectively) will be of the same shape, but will be horizontally and vertically offset by the constants $Q/4\pi KM$ and $4KM/S$. The two curves can be made to match, and the coordinates of an arbitrary matching point A are the related values of s, t/r^2, 1/u, and W(u), which can be used to calculate KM and S with Eqs 8-1 and 8-2, respectively (Fig. 8.5).

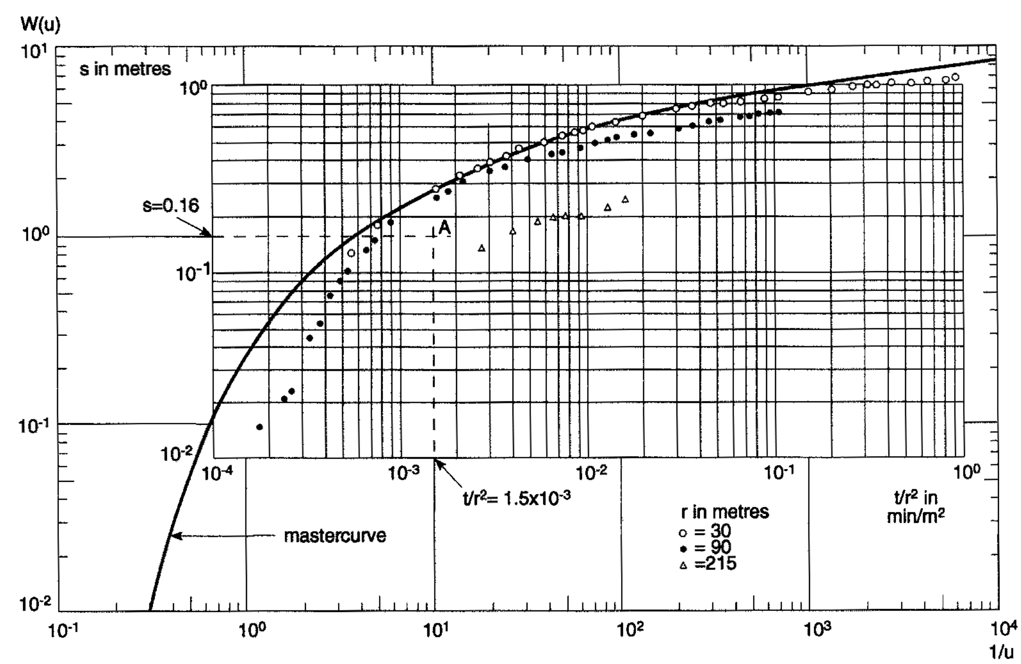
8.4.4 Jacob's straight line method

Jacob's method is based on the Theis formula (Eq. 8-3):

$$s = \frac{Q}{4\pi KM}\left(-0.5772 - \ln u + u - \frac{u^2}{2.2!} + \frac{u^3}{3.3!} - ....\right)$$

From $u = r^2S/4KMt$ it follows that u decreases as the time increases and the distance from the well r decreases. Accordingly, for drawdown observations made in the near vicinity of the well after a sufficiently long pumping time, the terms beyond ln u in the series become so small that they can be ignored. So, for small values of u ($u<0.01$), the drawdown can be approximated by:

Figure 8.5 Analysis of data from pumping test ($Q = 788\ m^3/d$) in Oude Korendijk (The Netherlands) with Theis's curve-fitting method



Note:

For point A: W(u) = 1 and i/u = 10 while s = 0.16 m and $t/r^2 = 1.5 \times 10^{-3}\ min/m^2 = 1.5 \times 10^{-3}/1440\ d/m^2$. Introducing these values and the value of Q into equations 8-1 and 8-2 yields: KM = (Q/4 s)W(u) = = 788/(4 x 3.14 x 0.16) = 392 m²/day.

$$s = \frac{Q}{4\pi KM}\ (-0.5772\ -\ \ln (r^2S/4KMt)$$

or

$$s = (2.30Q/4\pi KM)\log(2.25KMt/r^2S) \qquad (8\text{-}4)$$

As Q, KM and S are constant, a plot of drawdown s versus the logarithm of t forms a straight line (Fig. 8.6). If this line is extended until it intercepts the time axis where s = 0, the interception point has the coordinates s = 0 and $t = t_0$. Substituting these values into Equation 8-3 gives:

$$0 = (2.30Q/4\pi KM)\ \log(2.25\ KMt_0/r^2S)$$

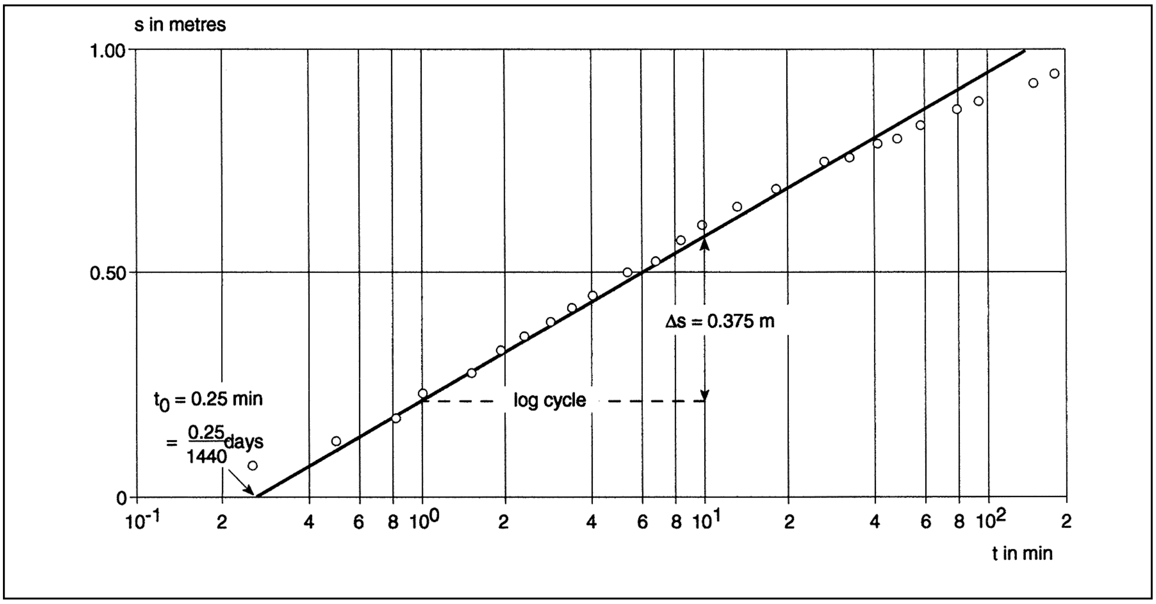
and because $2.30Q/4\pi KM \neq 0$, it follows that $2.25KMt_0/r^2S = 1$ or

$$S = 2.25KMt_0/r^2 \qquad (8\text{-}5)$$

The slope of the straight line in figure 8.6, i.e. the drawdown difference Δs per log cycle of time $\log t/t_0 = 1$, is equal to $2.30Q/4\pi KM$. Hence

$$KM = 2.30Q/4\pi\Delta s \qquad (8\text{-}6)$$

Figure 8.6 Analysis of piezometer (r = 30 m) data from pumping test (Q = 788 m^3/d) in Oude Korendijk (The Netherlands) with Jacob straight line method. Note that t_0 = 0.25 minutes or 0.25/1,440 days and that Δ_s = 0.375 m



Introducing these values into the Eqs 8-6 and 8-5 yields:

$$KM = (2.30 \times 788)/(4 \times 3.14 \times 0.375) = 385 \text{ m}^2\text{/d, and}$$

$$S = (2.25 \times 385 \times (0.25/1440))/30^2 = 1.7 \times 10^{-4}.$$

This method can be applied if the pumping test satisfies the assumptions and conditions mentioned above for Theis's method plus the condition:

- The values of u are small (u<0.01), i.e. r is small and t is sufficiently large.

8.4.5 Recovery analysis

The analysis of the data collected during the pumping period of single well tests (section 8.5.1) and pumping tests can be checked by the analysis of the recovery data collected after the pumping has stopped. The analysis is based on the principle of superposition, i.e. it is assumed that after the pump has been shut down, the well continues to be pumped at the same discharge as before, and that from the time pumping ceased, the well receives an imaginary recharge equal to the discharge (Fig. 8.7). The recharge and the discharge thus cancel each other out, resulting in an idle well, as is required for the recovery period.

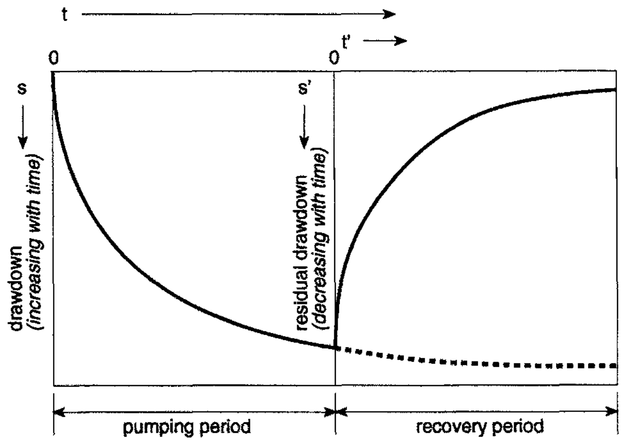
The residual drawdown, s′, after a pumping test with a constant discharge is:

$$s' = \frac{Q}{4\pi KM}\{W(u) - W(u')\}$$

where W(u′) is the well function for the assumed recharge.

When u and u′ are sufficiently small and the storage coefficient during the recovery

Figure 8.7 Time drawdown and residual drawdown



period can be assumed to be equal to the storage coefficient during the pumping period, this equation can be written as:

$$s' = \frac{Q}{4\pi KM}\left(\ln\frac{4KMt}{r^2S} - \ln\frac{4KMt'}{r^2S}\right) \qquad \text{or} \qquad s' = \frac{2.30Q}{4\pi KM}\log\left(\frac{t}{t'}\right)$$

where t is time since the start of pumping and t′ is the time since the cessation of pumping.

A plot of s′ versus t/t′ on single log paper (t/t′ on the logarithmic scale) will yield a straight line with a slope:

$$\Delta s' = \frac{2.30Q}{4\pi KM}, \text{ and hence } \qquad KM = \frac{2.30Q}{4\pi\Delta s'}$$

The recovery data does not allow the calculation of the value of S.

8.4.6 Well flow formula for other conditions

For other aquifer conditions (e.g. leaky aquifers, unconfined aquifers, anisotropic aquifers, aquifers close to hydraulic boundaries, wedge-shaped aquifers), or pumping conditions (e.g. large-diameter wells, variable discharge, partially penetrating wells) well flow equations have been developed that can be solved in an analogous manner, provided that the well function has been tabulated. As a matter of course the well functions become more complicated when more parameters are required to describe the flow to the well under the particular well flow conditions. Kruseman and De Ridder compiled the pumping test analysis methods for a wide variety of conditions (Kruseman et al., 1991; Genetier, 1984; see also Kruseman and De Ridder 1973, 1974 and 1975 that are translations in German, French and Spanish of Kruseman and De Ridder, 1970, the predecessor of Kruseman et al., 1990).

Specific yield from pumping tests

A remark must be made concerning the calculation of the specific yield, S_Y, of an unconfined aquifer. In principle it can be calculated from pumping test data, but in general the results are very unreliable. S_Y is usually estimated from lithological data (Section 8.5.5).

8.5 Determination of hydraulic characteristics by other methods

8.5.1 Single-well tests

In many instances the hydraulic parameters have to be determined when there are no piezometers and the water-level changes are measured only in the pumped well. The drawdown in a pumped well, however, is influenced by well losses and well bore storage. In the hydraulics of well flow, the well is assumed to have an infinitesimal radius so that the well bore storage can be ignored. In reality, well bore storage is large by comparison with the storage in an equal volume of aquifer material, therefore in a single-well test, well bore storage must be considered when analysing the drawdown data. This is especially true when pumping tests are carried out in large-diameter dug wells. On the other hand, the influence of well bore storage on the drawdown in a well decreases with time and becomes negligible at $t > 25\ r_c^2/KM$, where r_c is the radius of the blank casing in which the water level is changing.

To determine whether the early time drawdown data are dominated by well bore storage, a log-log plot of drawdown s_w versus pumping time t should be made. If the early time drawdown plot as a unit-slope straight line, we can conclude that well bore storage exists. Curve-fitting methods and straight line methods have been developed to analyse single-well tests, even when the early time data are affected by well bore storage (Kruseman et al., 1990).

8.5.2 Flowing well tests

When a confined aquifer whose piezometric level is well above the land surface is tapped by a well, this well will be free-flowing or artesian. That means that the well has a constant draw-down, i.e. the difference between the static piezometric head after shutdown and the outflow opening of the well. When the well is opened up after a period of shut-in, the piezometric level in the well drops instantaneously to the outflow level of the well, establishing a constant drawdown, while the well starts flowing with a decreasing discharge rate. When the changes of the discharge rate with time are measured, a curve-fitting method developed by Hantush (Kruseman et al., 1990) can be used to calculate the diffusivity of the well (KM/S).

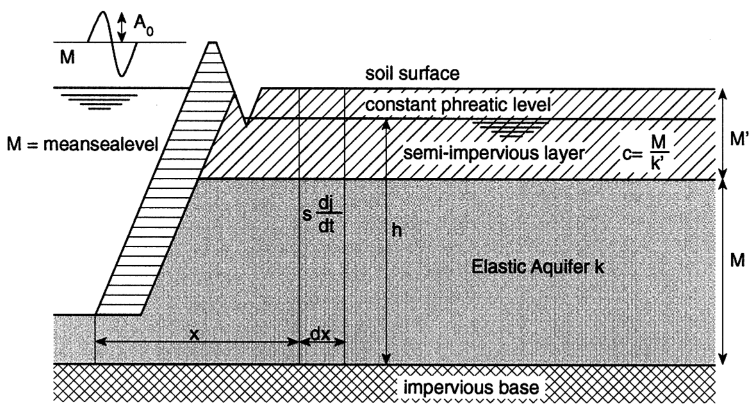
8.5.3 Slug tests

In a slug test, a small volume (or slug) of water is suddenly removed from a well, after which the rise of the water level in the well is measured. Alternatively, a small slug of water is poured into the well and the rise and subsequent fall of the water level are measured. From these measurements, the aquifer's transmissivity can be determined. If the aquifer's transmissivity is more than, say, 250 m^2/d, the water level will recover too quickly for accurate manual measurements and an automatic recording device will be needed. For conventional slug tests curve-fitting methods have been developed (Kruseman et al., 1990).

8.5.4 Tidal movements

In estuaries and coastal deltas where the inland surface water level fluctuates under influence of the tidal movements of the sea, the hydraulic parameters of semi-confined aquifers can be calculated from data on the propagation of the tidal wave in the aquifer. This requires piezometers in a line perpendicular to an open water course as well as a surface water level recorder on the same line. Bosch (1951), as quoted by Van Eyden et al., (1963), studied the situation depicted in Fig. 8.8.

Figure 8.8 Geohydrological conditions to which the formula of Bosch may be applied



They derived the following formula:

$$\Phi_{x,t} = h + A_0.e^{-\alpha x} \,.\, \sin(\omega t - \beta x) \qquad (8\text{-}7)$$

$$\alpha^2 - \beta^2 = \frac{1}{KM \,.\, c} \qquad (8\text{-}8)$$

$$\alpha \,.\, \beta = \frac{\omega \,.\, S}{2KM} \qquad (8\text{-}9)$$

where:

Φ : Piezometric level in the aquifer [L]
x : Distance to the open water [L]
t : time since observations started [T]
h : constant level of the phreatic surface in the semi-pervious confining layer [L]
A_0: amplitude of the tide in the estuary [L]
α : attenuation factor [L^{-1}]
β : retardation factor [L^{-1}]
ω : $2\pi/T$ [radians.L^{-1}]
T : duration of a tidal period [T]
K : hydraulic conductivity of the semi-confined aquifer [LT^{-1}]
M : thickness of the semi-confined aquifer [L]
c : hydraulic resistance of the confining layer [T]
S : storage coefficient [-]

To calculate the geohydrological characteristics the attenuation A_x/A_0 for each piezometer is plotted versus the distance x from the open water on semi-logarithmic paper (A_x/A_0 on the logarithmic scale). This gives a straight line. The slope of this line corresponds to the attenuation factor α. Subsequently, the retardation of the minimum and maximum values of the piezometric values with reference to the moments of Low Tide and High Tide, respectively, is plotted on graph paper versus the distance x from the open water. The slopes of the straight lines give the values of the retardation factor β for LT and HT, respectively. The average of the two values is used in the calculations.

With the values of α and β known, the values of KMc and S/KM are calculated from the equations (8-8) and (8-9).

8.5.5 Lithology and grain size analysis

There have been numerous attempts to determine the hydraulic conductivity from the lithology, or more precisely from the grain size analysis. The formulas developed for this purpose have only local significance and are only applicable to very uniform sand strata. Such formulas have to be calibrated by other methods, e.g. pumping tests.

The range of variation of the specific yield, S_Y, of an unconfined aquifer is rather small; from 0.05 for clayey fine sand to 0.30 for well rounded, perfectly sorted very coarse sand. The specific yield is therefore usually estimated from the lithological data, using tables such as presented by Driscoll (Table 8.4).

Table 8.4 Range of variation of the specific yield (Driscoll, 1986)

Material	S_y	*Material*	S_y
Coarse gravel	23	Limestone	14
Medium gravel	24	Dune sand	38
Fine gravel	25	Loess	18
Coarse sand	27	Peat	44
Medium sand	28	Schist	26
Fine sand	23	Siltstone	12
Silt	8	Silty till	6
Clay	3	Sandy till	16
Fine-grained sandstone	21	Gravelly till	16
Medium-grained sandstone	27	Tuff	21

We consider the values mentioned in Table 8.4 as maximum values for perfectly sorted and perfectly rounded material. Values half these may be encountered in the field.

8.5.6 Laboratory tests

Laboratory tests are generally considered as completely unreliable for determining the hydraulic conductivity of an aquifer.

8.5.7 Water balance and groundwater flow simulation model

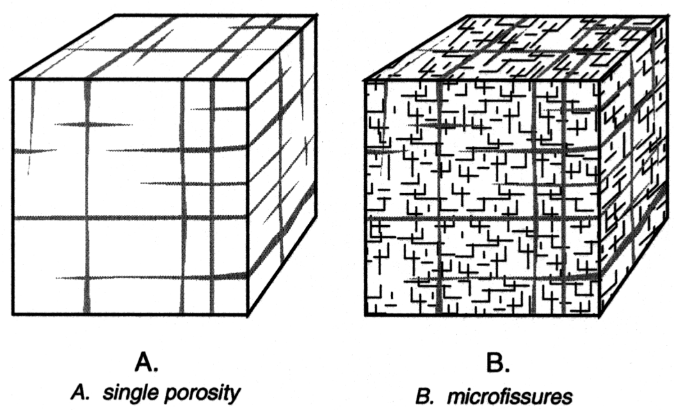
It is only possible to calculate the transmissivity from a water budget if the groundwater flow term is known accurately. This is only so if the groundwater recharge and discharge terms are known accurately, but these are terms that are notoriously difficult to determine.

The groundwater flow simulation model is little else than the integration of the simultaneous water balance calculation in each of the nodal areas in which the study area has been discretised (Anderson and Woessner, 1991). This has the advantage that all available knowledge on the aquifer and groundwater conditions can be integrated and that anomalies are clearly visible in the form of aberrant calculated groundwater levels. During the calibration process all parameters are adjusted within physically acceptable limits, to reproduce the historical groundwater levels. The accuracy of the calibrated transmissivity values depends on how accurate the other parameters are.

8.6 Determination of hydraulic characteristics of aquifers in fractured rocks

Fractures in a rock formation (Fig. 8.9) strongly influence the fluid flow in that formation. Conventional well flow equations were developed primarily for homogeneous and isotropic aquifers (e.g. uniformly well sorted sand layers), therefore they do not adequately describe the flow in fractured rocks. An exception occurs in hard rocks of very low permeability if the fractures are numerous enough and are evenly distributed throughout the rock; then the fluid flow will only occur through the fractures and will be similar to that in an unconsolidated homogeneous aquifer (Fig. 8.9-B).

Figure 8.9 Porosity systems: a) single porosity, b) microfissures, c) double porosity



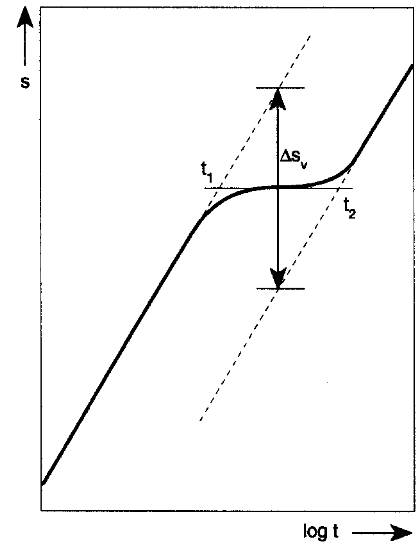
A complicating factor is the fracture pattern, which is seldom known accurately. This means that based on the geological data, a fracture model must be assumed . In recent years, many theoretical models have been developed, but only the well functions for the double porosity model (Figure 8.10), the single vertical fracture and the single vertical dike model have been tabulated (Kruseman et al., 1990).

8.6.1 Double-porosity model

The double-porosity concept regards fractured rocks as consisting of matrix blocks with a primary porosity and low hydraulic conductivity, separated by fractures with a low storage capacity but a high hydraulic conductivity. This concept assumes that there is no variation of head in the matrix blocks, i.e. the interporosity flow is in pseudo-steady state. The flow trough the fractures will be radial and in no-steady state when the aquifer is pumped by a well (Figure 8.11).

Curve fitting methods and straight line methods have been developed to analyse pumping tests in double porosity aquifers.

Figure 8.10 Semi-log time-drawdown plot for an observation well in a fractured rock formation of the double-porosity type



8.6.2 Single vertical fractures

If a well intersects a single vertical fracture, the aquifer's unsteady drawdown response to pumping differs significantly from that predicted by the Theis solution. This well flow problem has long been a subject of research in the petroleum industry, and various solutions have been proposed; most have produced erroneous results.

A major step forward was taken when the fracture was assumed to be a plane, vertical fracture of zero width (zero storage), relatively short length and infinite hydraulic conductivity (Figure 8.11).

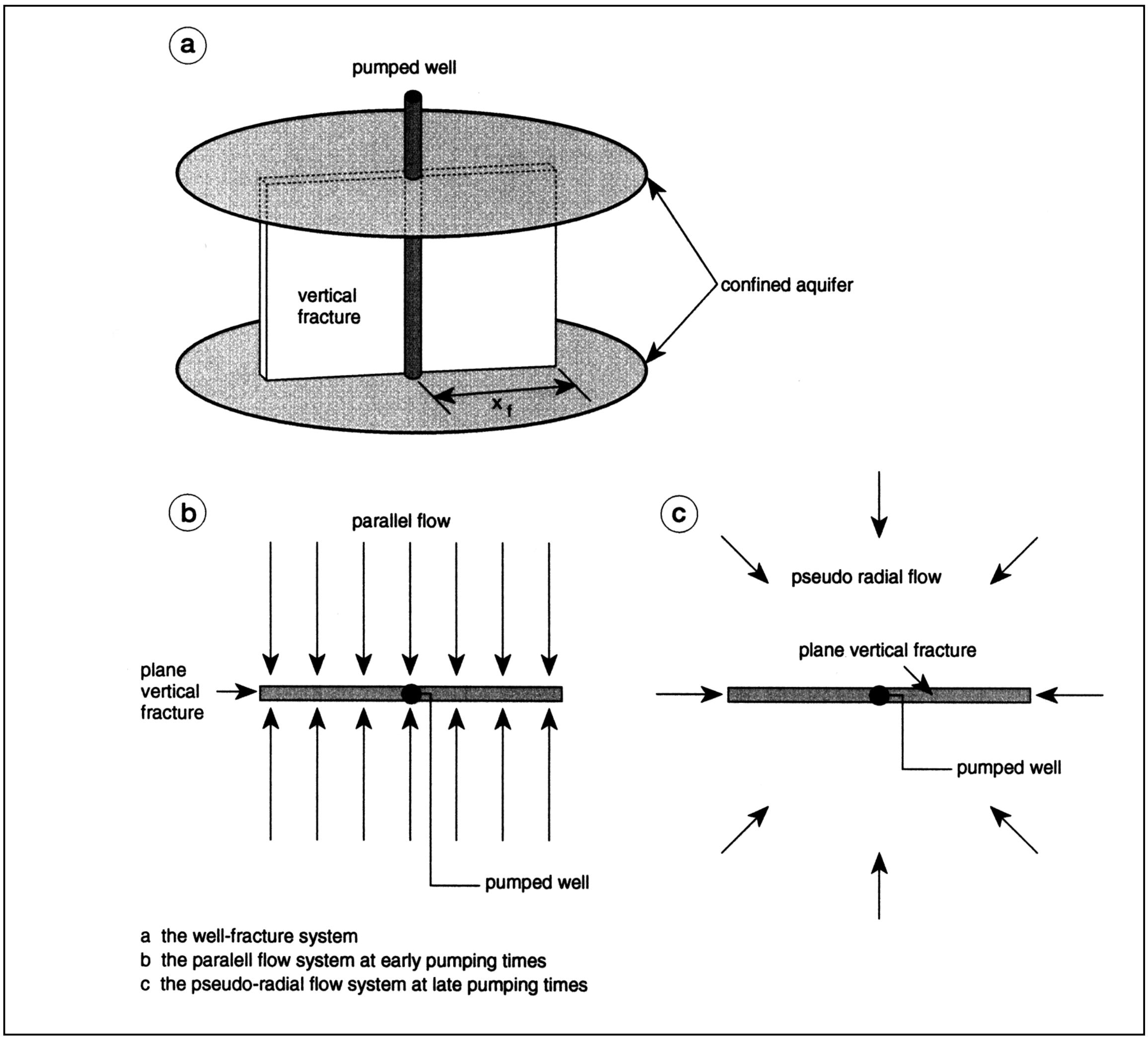
If an aquifer that satisfies this model is pumped by a well that intersects the plane fracture midway, the assumption of infinite (or very high) hydraulic conductivity means that the drawdown in the fracture is uniform over its entire length (i.e. there is no hydraulic gradient in the fracture). The rock around the fracture is poorly permeable, and when the well is pumped the drawdown in the fracture induces a flow from the rock matrix into the fracture. At early pumping times this flow is one-dimensional (i.e. it is horizontal, parallel and perpendicular to the fracture). In unconsolidated sediments this flow pattern occurs when there is a constant groundwater discharge into an open channel with constant head that fully penetrates the aquifer.

As pumping continues, the flow pattern changes from parallel flow to pseudo-radial flow, regardless of the fracture's hydraulic conductivity, and the aquifer reacts as an isotropic aquifer. The time required to attain pseudo-radial flow may be excessively long. Curve-fitting methods have been compiled to analyse pumping tests performed under the above mentioned conditions.

8.6.3 Single vertical dikes

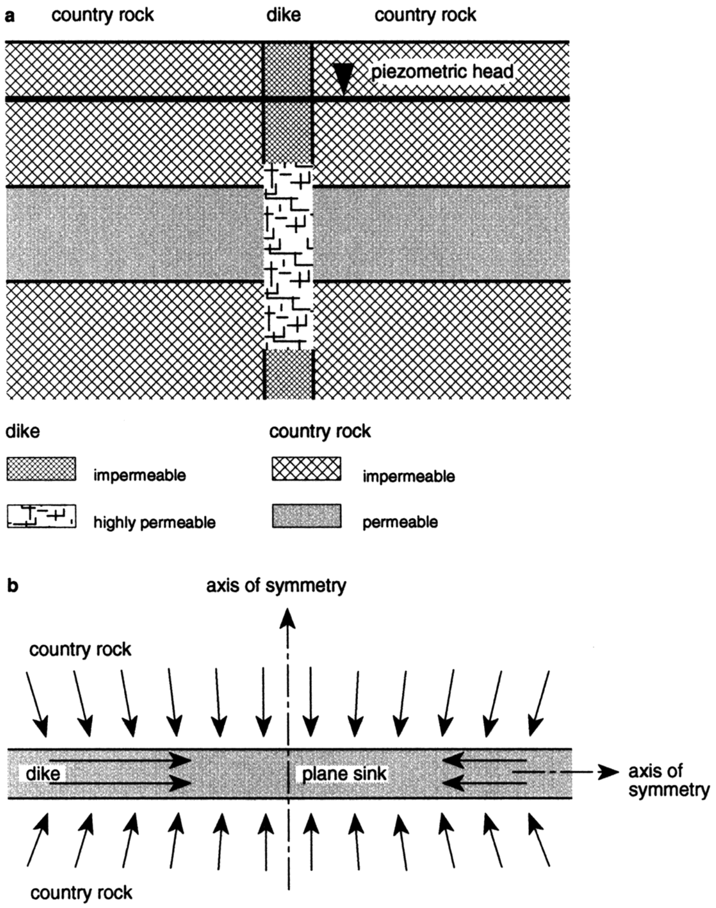
Dikes used to be regarded as impermeable, but they can become highly permeable as a result of jointing when the magma cooled, or by fracturing as a result of shearing, or by weathering. If a single, permeable, vertical dike bisects a country-rock aquifer whose transmissivity is several

Figure 8.11 A well that intersects a single plane fracture of finite length and infinite hydraulic conductivity: a) the well-fracture system, b) the parallel flow system at early pumping times, c) the pseudo-radial flow system at late pumping times



times less than that of the dike, a specific flow pattern will be created if the well is pumped. The dike is assumed to be infinitely long and to have a finite width and a finite hydraulic conductivity. The dike's permeability stems from a system of uniformly distributed fractures, extending downward and dying out with depth. Below the fractured zone, the dike rock is massive and impermeable. The upper part of the dike is also impermeable because of intensive weathering or a top clay layer (Figure 8.12). The water in the fractured part of the dike and in the aquifer in the country rock is thus confined. When the well is pumped at a constant rate, a trough of depression will be formed, instead of a cone of depression. The time drawdown curve shows three characteristic time periods: at early times all the water pumped originates from storage in the dike and none is contributed by the country-rock aquifer. At medium times, all the water pumped is supplied by the aquifer and none is contributed from storage in the dike. At late times, the flow in the aquifer is pseudo-radial. The publication of Kruseman et al. (1990) includes curve-fitting methods developed by Boehmer and Boonstra.

Figure 8.12 Composite dike/aquifer system: a) cross section showing an aquifer of low permeability in hydraulic contact with the highly permeable, fractured part of a vertical dike, b) plan view of flow in a pumped dike at early and late pumping times



8.7 Determination of hydraulic characteristics of volcanic rocks

Volcanic rocks show the characteristics both of sedimentary rocks (e.g. stratification of lavas, tuff and breccias, sometimes with alluvial intercalations, or with fossil soils on top of weathered parts of older volcanic deposits), and of fractured rock (the best known example probably being polygonal cooling joints in basalt).

The analysis of geological data in combination with data on the hydrological behaviour of the volcanic sequence must result in the selection of a model for the groundwater flow and hence in the selection of the model that must be used to analyse pumping tests or to provide other quantitative data on the groundwater flow.

8.8 Determination of hydraulic characteristics of limestone

In general, limestones have a low primary porosity and hence a low primary permeability. Fracturing will provide a secondary permeability that puts the limestones in the same league as other fractured rocks (Section 8.6). On the other, hand the fracturing is sometimes so intense that the whole rock is criss-crossed by tiny fractures, resulting in a pseudo-isotropic aquifer.

In thoroughly karstified limestone (mature karst), the saturated groundwater flow becomes restricted to the flow through large solution channels. This karst channel flow has more of the characteristics of surface water flow than of groundwater flow.

8.9 Diffusion, dispersion and macrodispersion

Diffusion, dispersion and macrodispersion are transport phenomena additional to advective groundwater flow. Diffusion and dispersion are mathematically combined into the disperse transport term in traditional groundwater transport models. The dispersive flux can be attributed to two processes: molecular diffusion and hydrodynamic dispersion. The first has to be attributed to Brownian motion of species and the second to differences in velocities in porous media due to friction along the edges of the pores and to the tortuosity of stream lines along individual grains (Fig. 8.13a,b). It should be stressed that these two processes at a microscopic level are the one and only two true physical processes that cause 'dispersive transport'. All other due to the mathematical treatment of the phenomenon of dispersion and the method of observation in the field or the laboratory (Domenico and Schwartz, 1990).

The two physical processes are combined to yield the dispersion coefficient, D, that is used in the description of groundwater transport:

$$D = D_{mp} + \alpha \, . \, v$$

where D_{mp} is the aqueous diffusion coefficient in a porous medium, a is the dispersivity and v is the average linear pore water velocity. It can be seen from this equation that molecular diffusion becomes increasingly important with decreasing flow rate. The diffusion coefficient in a porous medium is four times lower or more than that in a free liquid. This can be obtained from tables; for aqueous solutes at 25°C the value is about 1.0×10^{-5} cm^2/s.

Values for the dispersivity, α, are more troublesome to obtain. The best procedure is to create a breakthrough curve by a column experiment using a representative sample. However, this can be unachievable. An estimate can be made from the grain size analysis, because it is well known that the value for the dispersivity increases with increasing average grain size and decreasing sorting. The following expression provides a good estimate for geohydrologically homogeneous conditions.

$$\alpha = (d_{60}/d_{10}) \, . \, d_{50}$$

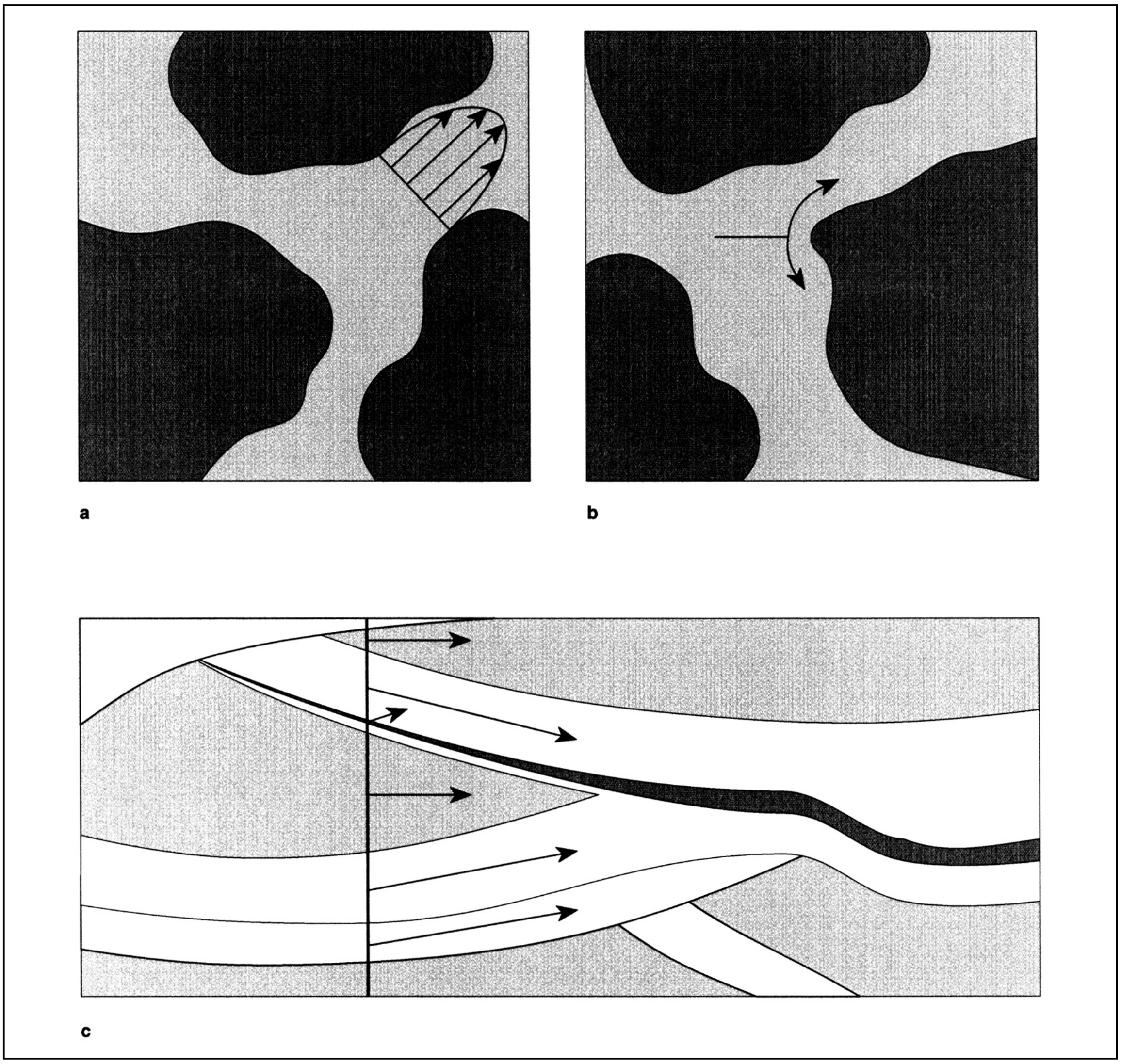
where d_i is the grain diameter below which i% of the grain size distribution falls.

If this approach is unrealistic due to the heterogeneous nature of the porous medium, an upper limit for the value of the dispersivity can be found if one assumes that the value is related to the length of the flow domain studied (L):

$$\alpha = L/10$$

In fact the dispersion is no longer considered as a true physical process but as a bundling of flow 'tubes', each with its own travel time for advective flow at a macroscopic level which is caused by the heterogeneous nature of the subsurface (Fig. 8.13c). This phenomenon is usually referred to as macrodispersion. The principal difference between the two is that dispersion is irreversible in its nature whereas macrodispersion is reversible. The combined act of hydrodynamic dispersion and molecular diffusion perpendicular to the groundwater flow direction together with macrodispersion leads to the partly irreversible nature of macrodispersion as well.

Figure 8.13 The principal processes of hydrodynamic dispersion and macrodispersion



8.10 References and additional reading

ANDERSON, M. P; WOESSNER, W. W., 1991. *Applied Groundwater Modeling*. Academic Press, Inc. San Diego, 381 pp.

BOSCH, H. (1951) Geohydrologisch onderzoek te Bergambacht. Unpublished research report. (In Dutch.)

DOMENICO, P. A.;. SCHWARTZ, F. W, 1990. *Physical and Chemical Hydrogeology; Second Edition*. John Wiley and Sons, New York.

DRISCOLL, F. G. (ed.), 1986. *Groundwater and Wells. Second Edition*. Johnson Division, St Paul, Minnesota, 1 089 pp.

FETTER, C. W. (1993). *Contaminant Hydrogeology.* Prentice Hall, Upper Saddle River, NJ.

GENETIER, B., 1984. *La pratique des pompages d'essai en hydrogéologie*. BRGM, manuels et methodes No. 9, Orléans, 132 pp.

GROUNDWATER MANUAL, 1981. *A Water Resources Technical Publication*. U.S. Department of the Interior, Water and Power Resources Service, U.S. Government Printing Office, Denver, Colorado, 480 pp.

KRUSEMAN, G. P.; DE RIDDER, N. A., 1970. *Analysis and Evaluation of Pumping Test Data*. ILRI Bull. 11, Wageningen, 200 pp.

—— (traduit par A. Meilhac), 1974. *Interprétation et discussion des pompages d'essai*. ILRI Bull. 11-F, Wageningen, 213 pp.

—— (übersetzt von A.W. Uehlendahl), 1973. *Untersuchung und Anwendung von Pumpversuchsdaten*. Verlagsgesellschaft Rudolf Müller, Köln-Braunsfeld, 191 S.

——, (traducción M. Donezar), 1975. *Análisis y evaluación de los datos de ensayos por bombeo*. ILRI Bol. 11-S, Wageningen, 212 pp.

KRUSEMAN, G. P., DE RIDDER, N. A.; VERWEIJ, J. M., 1991. *Analysis and Evaluation of Pumping Test Data (Completely Revised Second Edition)*. ILRI Publ. 47, Wageningen, 377 pp.

THEIS, C.V., 1935. The Relation between the Lowering of the Piezometric Surface and the Rate and Duration of Discharge of a Well using Groundwater Storage. *Trans. Amer. Geophys. Union,* Vol. 16, pp. 519–24.

VAN EYDEN, W. A. A.; KUPER, W. W.; SANTEMA, P., 1963. *Some Methods Used in the Geo-Hydrological Survey of the South-Western Deltaic Area in the Netherlands.* IAHS Publ. 63, pp. 528–57.

9 Isotopes techniques in groundwater investigations

9.1 Introduction

A comprehensive understanding of a groundwater system is necessary if thid resource is to be developed sustainably without adverse effects on the environment. In general the most crucial of the many questions which hydrogeologists may concern the safe yield and the source of water in a groundwater system. In order to provide a satisfactory answer, many associated extremely important, and often difficult, questions need to be answered. Isotope techniques are effective tools for obtaining a variety of hydrological information such as: the origin of groundwater, determination of its age, velocity and direction of flow, interrelations between surface waters and groundwaters, possible interconnections between different aquifers, aquifer characteristics such as porosity, transmissivity, dispersivity, etc. The cost of such investigations is often relatively small in comparison to the cost of classical hydrological techniques, and in addition isotopes provide information which sometimes cannot be obtained by other techniques.

Applications of isotopes in hydrology are based on the general concept of 'tracing', in which either intentionally introduced isotopes or naturally occurring (environmental) isotopes are employed. Environmental isotopes (either radioactive or stable) have the distinct advantage over injected (artificial) tracers in that they facilitate the study of various hydrogeological processes on a much larger temporal and spatial scale through their natural distribution in a hydrological system. Thus, environmental isotope methodologies are unique in regional studies of water resources to obtain time and space integrated characteristics of groundwater systems. Artificial tracers are generally effective for site-specific, local applications.

In this chapter, the basic principles of isotope techniques applied to groundwater studies are discussed. References provided include more details of different applications and examples of field studies. More specific advice on isotope applications in hydrological and environmental studies is available from the Isotope Hydrology Section, International Atomic Energy Agency (IAEA), Wagramerstrasse 5, P.O. Box 100, A-1400, Vienna, Austria.

9.2 Environmental isotopes

Environmental isotopes, both stable and radioactive, occur in the atmosphere and the hydrosphere in varying concentrations. So far, the most frequently used environmental isotopes include those of the water molecule, hydrogen: ^{2}H or D, deuterium, and ^{3}H, tritium and oxygen: ^{18}O, as well as of carbon: ^{13}C and ^{14}C, radiocarbon or carbon-14 occurring in water as constituents of dissolved inorganic and organic carbon compounds. ^{2}H, ^{13}C and ^{18}O are stable isotopes of the respective elements whereas ^{3}H and ^{14}C are radioactive isotopes. The stable isotopes are usually measured, with an isotope ratio mass spectrometer, to find the isotope ratios of the less abundant to more abundant isotope, e.g., ^{2}H:^{1}H and ^{18}O:^{16}O (^{1}H and ^{16}O being the number of atoms of the most abundant isotopes of the respective elements). The radioactive isotopes are measured either by counting their radioactive decays (low-level counting), e.g., by Liquid Scintillation Spectrometry) or, using Accelerator Mass Spectrometry, by counting the number of atoms in a given sample.

9.2.1 Stable isotopes

(i) Stable isotope ratios

Variations in stable isotope ratios of natural compounds are governed by chemical reactions and phase changes due to the energy difference between chemical bonds involving different isotopes of an element. Such energy differences are caused by the relative mass difference between isotopes. The stable isotopes of the light elements show the widest variations because they have the largest relative mass difference. For example, variations in ^{2}H:^{1}H ratio are greater than ^{18}O:^{16}O because the ratio of mass difference between ^{2}H and ^{1}H is 2:1 while the same for ^{18}O and ^{16}O is 1.1:1. In addition to the ^{2}H:^{1}H and ^{18}O:^{16}O ratios in water, various other stable isotopes are used in groundwater studies (Table 9.1).

Table 9.1 Environmental stable isotopes used in groundwater studies

Isotope	*Ratio*	*Natural abundance (atoms/atoms)*	*Isotope*	*Ratio*	*Natural abundance (atoms/atoms)*
^{2}H	$^{2}H/^{1}H$	$1.55 \cdot 10^{-4}$	^{18}O	$^{18}O/^{16}O$	$2.04 \cdot 10^{-3}$
^{3}He	$^{3}He/^{4}He$	$1.38 \cdot 10^{-6}$	^{34}S	$^{34}S/^{32}S$	$4.22 \cdot 10^{-2}$
^{6}Li	$^{6}Li/^{7}Li$	$7.50 \cdot 10^{-2}$	^{37}Cl	$^{37}Cl/^{35}Cl$	0.242
^{11}B	$^{11}B/^{10}B$	0.80	^{81}Br	$^{81}Br/^{79}Br$	0.493
^{13}C	$^{13}C/^{12}C$	$1.11 \cdot 10^{-2}$	^{87}Sr	$^{87}Sr/^{86}Sr$	0.709939
^{15}N	$^{15}N/^{14}N$	$3.66 \cdot 10^{-2}$			

The interest of an isotope hydrologist lies in the relative deviation of the ratio of less abundant heavy isotope to more abundant lighter isotope (exceptions: helium, lithium and boron isotope ratios) with respect to a reference rather than in the 'absolute' isotope ratio of a given sample. Therefore, a material is selected as a primary standard, and its stable isotope ratio defines the zero point of a relative conventional scale. For convenience the measurements are not reported as isotope ratios, but given as relative deviation from the isotope ratio of a standard expressed as δ (delta) in permil (‰), defined as

$$(‰) = \frac{R_{sample} - R_{standard}}{R_{standard}} \times 1{,}000 \qquad (9.1)$$

where, δ (e.g., $\delta^{2}H$, $\delta^{13}C$, $\delta^{15}N$, $\delta^{18}O$, $\delta^{34}S$) is the normalised difference of the isotope concentration ratios R (^{2}H:^{1}H, ^{13}C:^{12}C, ^{15}N:^{14}N, ^{18}O:^{16}O, ^{34}S:^{32}S) of a sample and a standard. For example, hydrogen and oxygen isotopes are presented with reference to the hypothetical Standard Mean Ocean Water (SMOW) (Craig, 1961a), which has been superseded by the Vienna Standard Mean Ocean Water (VSMOW) standard which for all practical purposes is close to SMOW.

(ii) Stable isotopes of hydrogen and oxygen

Most of the applications of stable isotopes of hydrogen and oxygen in groundwater studies employ regularities in their variations in atmospheric precipitation, i.e., in the input to a hydrogeological system under study.

Many natural processes cause variations in isotope compositions of natural waters. The most important are evaporation and condensation. During evaporation, the light molecule of water, $H_2{}^{16}O$, is more volatile than the heavy molecules, i.e., $^{1}H^{2}H^{16}O$ or $H_2{}^{18}O$. Therefore, vapour which evaporates from an ocean is depleted with respect to the ocean water by about 12–15‰ in ^{18}O and 80–120‰ in deuterium. When this atmospheric water vapour undergoes successive cooling and condensation with production of clouds and precipitation, the less volatile (heavy) water molecules condense preferentially, leaving a residual vapour more and more depleted of ^{2}H and ^{18}O. As a result, successive precipitations derived from the same initial vapour mass will be more and more depleted of heavy isotopes. Because the degree of condensation of a vapour mass depends on temperature, a relation between isotope composition of precipitation and its temperature of formation should be expected: as the formation temperature decreases, the δ-values of precipitation decrease (Dansgaard, 1954). This has been observed directly in Antarctic precipitation (Picciotto et al., 1960) and a world-wide relation between ^{18}O of precipitation and mean annual air temperature has been reported (Fig. 9.1).

This dependency on temperature produces seasonal isotope variations of precipitation (there is more heavy isotope depletion in winter precipitation than in summer precipitation), latitude effect (high latitude precipitation has been depleted more than low latitude precipitation), and altitude effect (heavy isotope content of precipitation decreases with increasing altitude) (Friedman et al., 1964; Moser and Stichler, 1970; Rosanski et al., 1993). These effects constrain the use of these isotopes to delineate various hydrogeological processes, to indicate past and present climate changes and to identify palaeowaters.

Through long-term observations made within the framework of the IAEA/WMO Global Network for Isotopes in Precipitation (GNIP), a linear correlation between $\delta^{2}H$ and $\delta^{18}O$ in precipitation samples collected from a world-wide network of stations has been established (Rozanski et al., 1993), which is close to the so-called Global Meteoric Water Line (GMWL) defined by Craig (1961b) and represented by the relationship:

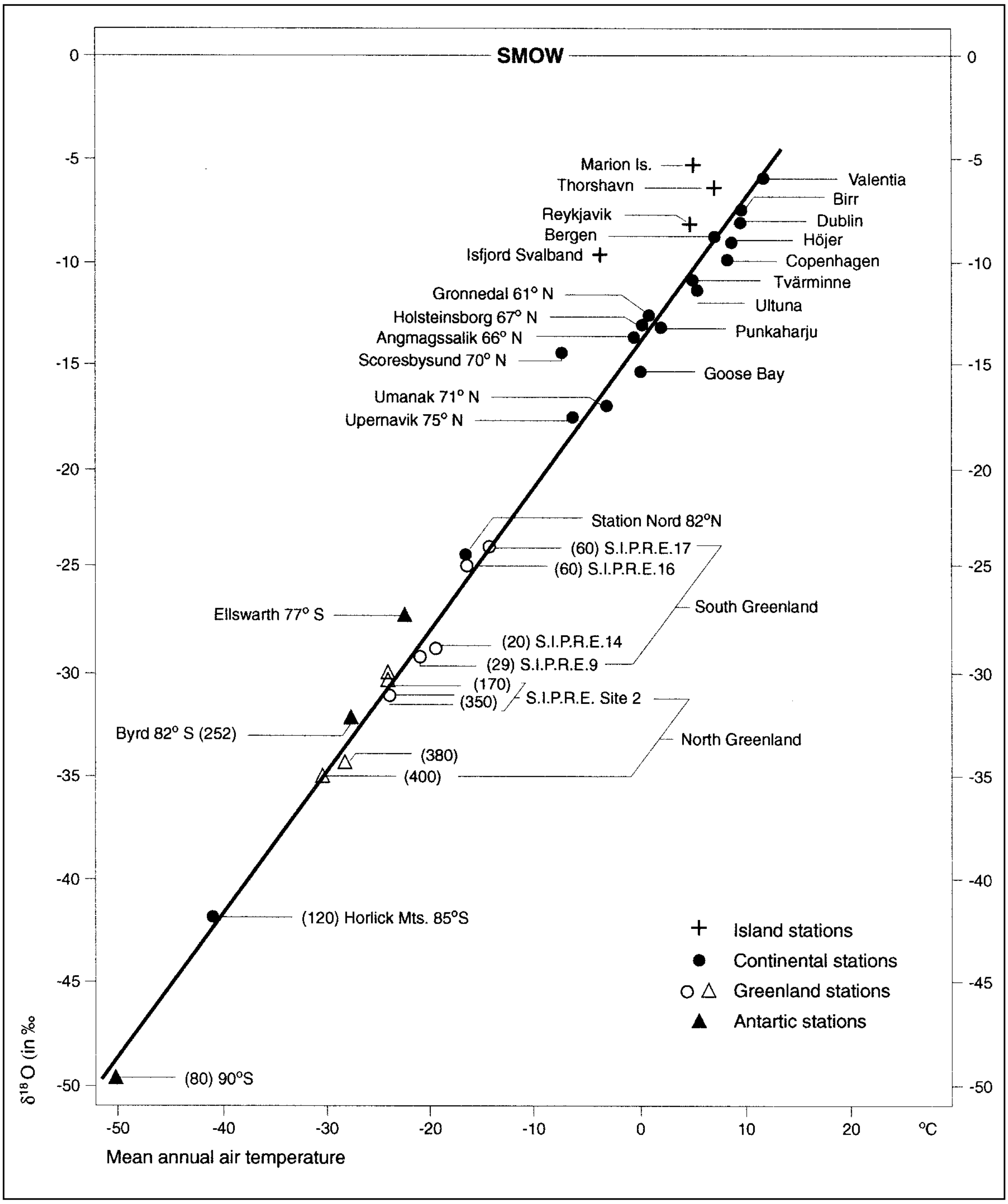
$$\delta^{2}H = 8 \cdot \delta^{18}O + 10 \qquad (9.2)$$

Precipitation which has undergone significant evaporation during its fall does not obey Equation (9.2). Evaporation tends to enrich both of the heavy isotopes in water, but not in the same relative proportion indicated by the above relationship (Craig, 1961b; Ehhalt et al., 1963; Woodcock and Friedman, 1963).

When precipitation infiltrates to recharge groundwater, mixing in the unsaturated zone and selective infiltration of precipitation result in attenuation of seasonal isotopic variations in precipitation. In most aquifers, isotopic composition of water does not change further unless exchange with the oxygen in rocks occurs. This process of exchange is significant and important for high temperature geothermal systems only. The isotopic composition of groundwater is thus related to that of precipitation in the recharge area of an aquifer at the time of recharge. Groundwater may be very old, and the climatic conditions of the region at the time of recharge may have been different from those of today. Due to the correlation between δ-values and temperature (Gat, 1971) this implies that the isotopic composition of precipitation could have been different from the present one.

Groundwater may also be recharged by seepage from surface waters, such as rivers and

Figure 9.1 The annual mean ^{18}O of precipitation as a function of the mean annual air temperature at surface. The figures in parentheses indicate the total thickness (in cm) of the investigated snow layers (Modified after Dansgaard, 1964)



lakes. If most of the recharge is from seepage, the groundwater should reflect the mean isotopic composition of the river or the lake instead of that of local precipitation which could be rather different. The river may collect water which originates from precipitation in a completely different area, for instance in a high mountain region; in this case its heavy isotope content would be depleted compared to that of precipitation in the plain, due to altitude effect.

In the case of lakes or ponds, the water may be considerably enriched in heavy isotopes through evaporation. The enrichment is limited by direct isotopic exchange with atmospheric moisture. Thus, this enrichment is higher where evaporation is more intense with respect to the total volume of water, e.g., in closed lakes and ponds or lakes and rivers in arid areas.

Groundwaters which have undergone evaporation before, during or after recharge are easily recognised by their isotopic composition. Their heavy isotope content is higher than that of non-evaporated waters in the region and they do not obey the relationship in the Equation (9.2) (Craig and Gordon, 1965; Fontes and Gonfiantini, 1967; Dinçer, 1968).

9.2.2 Radioactive isotopes

Among the environmental radioisotopes, tritium and carbon-14 have found the widest application in groundwater studies. Radioactive isotopes (also called radioisotopes) occurring in groundwater originate from natural and/or artificial nuclear processes (Table 9.2).

Table 9.2 Cosmogenic and anthropogenic radioisotopes used in hydrology

		Atmospheric production		In situ *production*	
Radioactive isotope	*Half-life (a)*	*Reaction*	*Initial content in 1 litre of groundwater*	*Reaction*	*Equilibrium content in 1 litre of groundwater*
^{85}Kr	10.76	Anthropogenic	$2.5*10^{-6}$ Bq	None	0
^{3}H	12.43	^{14}N (n,t)	1 Bq	^{6}Li (n,α)	$2*10^{-5}$ Bq
^{39}Ar	269	^{40}Ar (n,2n)	$8.5*10^{3}$ atoms	^{39}K (n,p)	$0.2-2*10^{3}$ atoms
^{14}C	5,730	^{14}N (n,p)	$3{\cdot}10^{-3}$ Bq	^{17}O (n,α)	$3*10^{-5}$ Bq
^{81}Kr	$2.1*10^{5}$	^{80}Kr (n,γ)	1,000 atoms	None	0
^{36}Cl	$3.01*10^{5}$	^{35}Cl (n,γ)	10^{5}–10^{8} atoms	^{35}Cl (n,γ)	$5*10^{7}$ atoms
^{129}I	$16*10^{6}$	Spontaneous fission	$5*10^{4}$ atoms	Spontaneous fission	$3*10^{5}$ atoms

Note:
'Bq' in columns 4 and 6 implies measurement by decay counting using liquid scintillation spectrometry whereas 'atoms' implies measurement of atoms using accelerator mass spectrometry. ^{14}C could also be measured by AMS techniques.

Cosmogenic radioisotopes are produced in nuclear reactions between the nucleonic component of cosmic radiation and the atmosphere. Anthropogenic radioisotopes are produced in nuclear bomb tests and nuclear reactors. The concentrations of all these radioisotopes in groundwater are very low and usually measured by counting their decay rate A in a given sample. The number of atoms N in a sample can be derived from A by the relationship:

$$A = \lambda \, . \, N \tag{9.3}$$

where, λ, the decay constant, is related to half-life $T_{1/2}$ by the equation $\lambda = \ln 2/T_{1/2}$. For long-lived radioisotopes such as ^{36}Cl and ^{129}I, the decay rate becomes unmeasurably small. In these cases the number of atoms has to be measured directly, which is possible using accelerator mass spectrometry (Elmore and Phillips, 1987). This AMS technique is superior to the conventional decay counting for ^{14}C, since AMS requires a very small sample size (up to 1,000 times less than conventional requirement) for analysis.

(i) Tritium

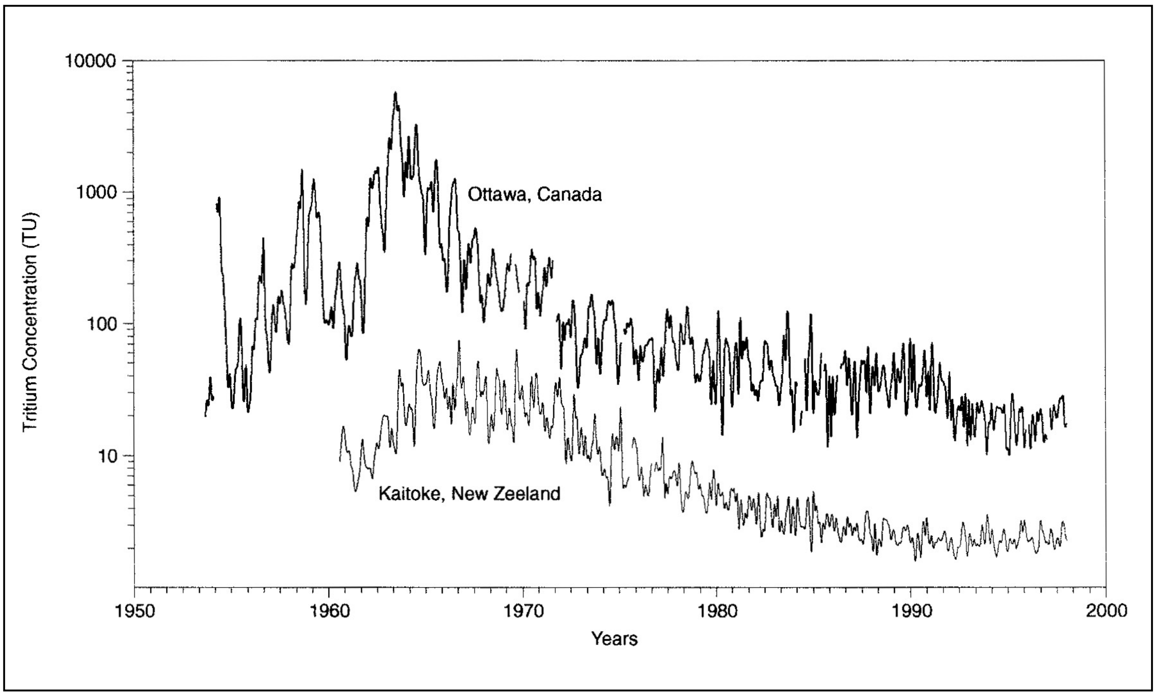
Tritium (^3H), the radioisotope of hydrogen, emits low-energy β radiation ($E_{max.}$ = 18 keV). Tritium content in water is expressed in Tritium Units (TU). 1 TU is equal to 1 atom of ^3H per 10^{18} atoms of ^1H, which is equivalent to 0.118 Bq or 3.193 pCi per litre of water. The half-life of tritium is fixed as 12.43 a (Taylor and Roether, 1982). The concentration of tritium in natural waters is generally very low. In hydrological studies, therefore, electrolytic enrichment of tritium is often carried out prior to decay counting with liquid scintillation or proportional counters.

Environmental tritium occurs in precipitation from both natural and anthropogenic sources. The natural production results from cosmic-ray produced neutrons interacting in the upper atmosphere with nitrogen atoms,

$$^{14}_{7}N + ^{1}_{0}n \rightarrow ^{3}_{1}H + ^{12}_{6}C \qquad (9.4)$$

Tritium oxidises rapidly to HTO and enters the global hydrological cycle. The natural content of tritium in precipitation is estimated to be about 2–5 TU (Craig and Lal, 1961; Lal and Peters, 1962). The second source of tritium is from atmospheric detonation of thermonuclear devices from 1952 to 1962, and minor releases from industrial nuclear facilities. The atmospheric testing injected periodic pulses of tritium into the stratosphere and in the Northern Hemisphere in 1963 its concentration in precipitation had increased by more than three orders of magnitude above that arising from cosmic-ray source (~5 TU). An increase in concentration was also noted in the southern hemisphere but by only two orders of magnitude because of the lower ratio of land to ocean in that hemisphere and the poor trans-equatorial mixing of the air masses. The history of tritium content in precipitation is well known, owing to the global network of stations jointly established by the IAEA and the WMO (Fig. 9.2).

Figure 9.2 Long-term tritium concentration in precipitation at Ottawa (Canada) and Kaitoke (New Zealand)



The International Atomic Energy Agency, Vienna, publishes data on the concentration of stable isotopes and tritium in precipitation samples collected at a large number of stations around the globe, from which it is possible to estimate the tritium deposition at most places of interest. The detailed information, references and data can be found on the Internet site: http://www.iaea.org/programs/ri/gnip/gnipmain.htm.

Over the past decades, groundwater studies have considerably benefited from the transient behaviour of bomb-tritium. Shallow groundwater was naturally labelled by the high, impulse-generated tritium concentration of atmospheric precipitation. The detection of bomb-tritium in shallow groundwater is a fingerprint for a component of recent recharge in the groundwater. Indication of recent replenishment of groundwater is very important for the management of these resources, especially in semi-arid regions where groundwater is often a non-renewable resource. A more quantitative treatment of tritium data in shallow, unconfined aquifers allows of the residence time distribution in groundwater to be determined, from which relevant parameters of the groundwater system can be estimated, especially the recharge rate. Presently, tritium concentration in precipitation is approaching the natural level, which makes such evaluations of tritium data more difficult. It is now often impossible to use tritium for evaluating groundwater recharge in the southern hemisphere, where the bomb-tritium in precipitation is much lower than in the northern hemisphere.

However, the combined measurement of tritium and its decay product, helium (^3He), still provides a powerful tool for estimating groundwater residence time and recharge rate, and for characterising flow and dispersion regimes in shallow aquifers. Assuming that a water mass travels like a parcel (without mixing and dispersion) along the flow path and that there is neither a ^3He loss nor a contribution from sources other than tritium decay within the parcel, the radioactive decay law can be employed to derive the residence time (transit time from the point of recharge to the sampling site):

$$t - t_0 = \frac{1}{\lambda} \ln \left[\frac{1 + {}^3He(t)}{{}^3H(t)} \right] \qquad (9.5)$$

where, ^3H (t) is the tritium concentration at the time of sampling t, ^3He (t) is the tritiogenic ^3He concentration (given in the same measuring unit), and t_0 is the time at which a water parcel reached the water table. The tritiogenic ^3He concentration is obtained from the measured total dissolved ^3He in a sample and subtracting the atmospheric and subsurface contributions. This requires additional measurements of neon and helium-4 in the sample (Schlosser et al., 1989). Note that in equation (9.5) initial tritium concentration ^3H (t_0) is not explicitly included. This quantity can, however, be reconstructed through the combined measurement of ^3H and tritiogenic ^3He according to the equation:

$$^3H(t_n) = {}^3H(t) + {}^3He(t) \qquad (9.6)$$

Thus, with this approach, the transient behaviour of bomb-tritium can be exploited to advantage even if most of this tritium has already decayed.

(ii) Radiocarbon (^{14}C)

Radiocarbon is produced in the transitional region between stratosphere and troposphere by cosmic ray neutrons reacting with nitrogen atoms (Libby, 1965),

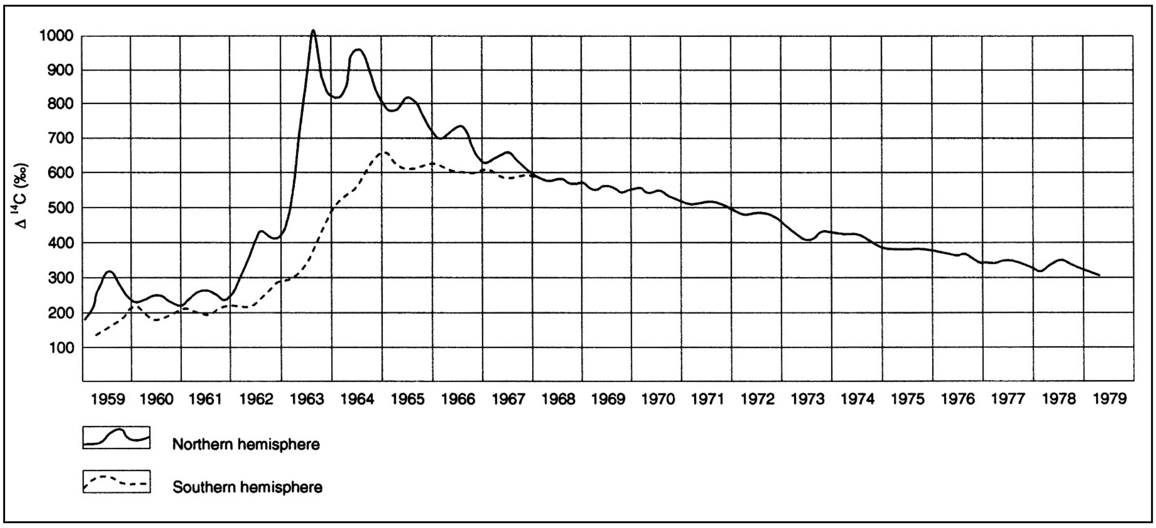
$${}^{14}_{7}N + {}^{1}_{0}n \rightarrow {}^{14}_{6}C + {}^{1}_{1}H \qquad (9.7)$$

^{14}C oxidises to carbon dioxide and becomes a part of the atmospheric carbon dioxide reservoir and subsequently enters the biosphere and the hydrosphere. ^{14}C has also been added to the atmosphere as a result of the testing of thermonuclear devices since 1952 (Fontes, 1983; IAEA, 1983).

^{14}C emits β radiation ($E_{max.}$ = 156 keV) and has a half-life of 5,730 a (Godwin, 1962). For hydrogeological applications, the ^{14}C concentration is expressed as a percentage of modern (pre-bomb) carbon activity of biosphere material used (percent of Modern Carbon or pMC). This refers to the ratio of the activity of a sample to 95% of the activity (in 1950) of the accepted oxalic acid standard from the US National Institute of Standards and Technology (NIST). The 100% corresponds to specific ^{14}C activity of 13.56 ± 0.07 disintegrations per minute per gram of carbon (Olsson, 1968). The activity of ^{14}C is generally measured by decay counting using liquid scintillation spectrometry or, more recently, by measurement of atoms using AMS.

The addition of man-made radiocarbon to the atmosphere has increased the natural levels; the maximum increase was about 100% in the Northern Hemisphere (achieved in 1963) and 70% in the Southern Hemisphere (achieved in 1965) (Fig. 9.3). This effect is important when dealing with very young groundwaters near the recharge areas.

Figure 9.3 Carbon-14 distribution in atmospheric CO_2 (Levin, 1978)



^{14}C determinations are generally carried out on the total dissolved inorganic carbon (TDIC), i.e., dissolved carbonate species CO_2 (aq.), HCO_3^{1-} and CO_3^{2-}, the bicarbonate ions being the species by far prevailing at the pH values normally encountered in groundwater. The major part of carbon enters the groundwater through soil CO_2, during the infiltration processes. However, chemical and biochemical reactions along a flow path can modify the initial ^{14}C content of groundwater. Therefore, in application of radiocarbon dating of groundwater, appropriate adjustments for effects of these reactions should be made in order to determine the initial activity of ^{14}C in the groundwater. Several models have been developed that consider the ^{13}C and other physico-chemical parameters for the required corrections.

The principles for dating groundwater are based on the decay law and the assumptions made with regard to the origin of the TDIC in groundwater. However, the behaviour of carbon species in the solution is not conservative and several correction models have been developed to account for the chemical interaction between the rock matrix, and other possible gaseous phases in the system. The detailed information on various corrections or models can be found in the monographs or textbooks (e.g., Fritz and Fontes, 1980; Clark and Fritz, 1997).

The other radioisotopes compiled in Table 9.2 have potential to date groundwater in an age range of the order of their half-lives, i.e., many tens to some hundreds of thousands of years.

9.2.3 Applications in groundwater studies

(i) Groundwater recharge

A qualitative and quantitative characterisation of groundwater recharge is essential for sustainable development and management of groundwater resources. When withdrawal rate exceeds recharge rate (over-exploitation), considerable lowering of the water table may be observed. Subsequently, the manifestation of the phenomenon could be sea water encroachment in coastal aquifers, induced leakage from adjacent saline aquifers into fresh aquifers or contamination of aquifers from polluted surface water bodies

Aquifers which receive little recharge will exhibit only small fluctuations in groundwater levels, so that a reliable estimate of recharge rate cannot easily be obtained on the basis of classical approaches such as water level monitoring alone. In many recharge rate estimations isotope techniques have provided a cost-effective tool. These techniques constitute virtually the only approach for identification and evaluation of groundwater recharge particularly under arid and semi-arid conditions.

a. Recharge from precipitation

It is possible to identify modern recharge, which occurred during the past forty to fifty years by using bomb-tritium data measured in soil water in an unsaturated zone and in groundwater from shallow, unconfined aquifers and springs. An example of the use of tritium from the unsaturated zone is from Louga, Senegal, where the tritium content of samples taken in 1990 from an unsaturated zone profile showed the 1963 tritium peak at 12 m below the land surface. The depth to the water table was 34 m. By applying a simulation model, the rate of modern recharge was estimated to be 28 mm per year, corresponding to about 10% of the annual rainfall (Tandia et al., 1993). Note that Louga is located in the Sahel region where precipitation events, although sporadic, have a regular recurrence.

Similar investigations have been carried out using the ^{36}Cl bomb-peak in precipitation (e.g. Liu et al., 1995). The maximum ^{36}Cl concentration occurred in precipitation around 1955, almost 10 years prior to the tritium peak in precipitation. Presently, the tritium and ^{36}Cl peaks can only be found in the unsaturated zone if the recharge rate is low and the unsaturated zone sufficiently thick (e.g., sand dunes in arid and semi-arid zones). Under more humid conditions, these peaks have already reached the water table.

For a shallow unconfined aquifer recharged by vertical infiltration of precipitation, the vertical penetration depth H of the thermonuclear tritium peak below the water table can be determined and the average recharge rate R estimated using the relationship:

$$R = \frac{n\,H}{t - t_0} \tag{9.8}$$

where, n is the effective porosity, t the sampling time and, in the case of bomb-tritium, $t_0 = 1963$. The applicability of the method is limited to the upper region of shallow aquifers, ideally for the top layer of a water table where vertical movement of water is dominant (Andres and Egger, 1985). Equation (9.8) can also be used when the water samples are collected from supply wells tapping the whole thickness of an unconfined aquifer. In that case, H stands for the aquifer thickness (assumed to be constant along flow direction) and the $(t - t_0)$ term is replaced by mean residence time derived from the measured tritium content of a sample (Zuber, 1986). In a more comprehensive demonstration of this approach, tritium concentrations of water samples collected from an unconfined aquifer near Sturgeon Falls, Ontario, Canada were evaluated using a one-dimensional advection-dispersion model (Robertson and Cherry, 1989). A recharge rate of 150 mm/a (equivalent to about 16% of the annual precipitation) was obtained.

The ^{3}H–^{3}He method represented by equations (9.5) and (9.6) has enhanced the potential of bomb-tritium to estimate groundwater recharge rates through residence time determination. In a shallow, sandy aquifer system in the southern New Jersey coastal plain, the apparent $^{3}H/^{3}He$ ages were found to be in good agreement with results obtained by other approaches like chlorofluorocarbon (CFC) dating and model-simulated flow path analysis (Szabo et al., 1996). In this case, the vertical flow velocity was found to be of the order of 1 m/year.

Nevertheless, even in cases of low vertical flow velocities, the identification of the $^{3}H-^{3}He$ peak can be used for dating and thus, for recharge rate estimation (recharge rate = porosity x vertical flow velocity). In addition, the mere presence of tritiogenic ^{3}He in groundwater with no measurable tritium provides evidence of modern recharge. For example, an ^{3}He content of about 3 TU in groundwater with no tritium can result in an aquifer where there is modern recharge with water that has a tritium content of about 3 TU near the water table. Although the rate of recharge in such an aquifer would be very low, it must be present in order to provide a continuous source of ^{3}He and to overcome diffusive ^{3}He losses. The tritium and ^{3}He data can be modelled to estimate recharge and transport parameters in unsaturated and saturated zones (Dillon and Aggarwal, 1999).

Under certain circumstances, the residence time and thus recharge rate of young groundwater can also be estimated from measurements of the seasonal $^{2}H/^{1}H$ and $^{18}O/^{16}O$ variations. It has been shown by Maloszewski et al. (1983) that the mean residence time of groundwater is a function of attenuation of the amplitude of its seasonal variation relative to the seasonal variation in local precipitation replenishing the groundwater. The applicability of this method is limited to those areas where precipitation shows a pronounced seasonal variation, e.g., in mountainous areas. The method was applied by Ramspacher et al. (1992) to estimate the recharge rate of an Alpine groundwater system in Austria.

Many karst springs along the northeast Adriatic coast of Italy show seasonal changes in the stable isotope composition of groundwater. The temporal pattern showed the most depleted values during the summer months, suggesting a six-month time lag between the signal observed in precipitation and the values measured in the springs. These results were unexpected, considering that karst aquifers usually show a rapid transit time from recharge area to the sampling point. On the basis of the location of the springs it was possible to prove that during the winter months, the dominant recharge area is located closer to the coastal area, while during the summer period, the precipitation at higher elevation (and therefore, isotopically more depleted) was preferentially contributing to the recharge of the karst system (Flora and Longinelli, 1989).

b. Recharge from surface water

Groundwater often consists of a mixture of recharge from surface water (lakes or rivers) and local precipitation. It is important to know the proportions of these recharge components in order to (1) increase the sustainable supply of drinking water through bank infiltration ('groundwater enrichment', induced recharge) and/or (2) prevent drinking water pollution by infiltration of water from a contaminated surface water source. Different recharge components can be identified from the stable isotope compositions of groundwater because evaporation of water in surface water bodies, in particular under semi-arid and arid conditions, leads to an enrichment of the heavy isotopes ^{2}H and ^{18}O. A simple isotopic balance equation can then be used to estimate the relative proportions of surface water and precipitation in recharge. The accuracy of this determination generally depends on the magnitude of the difference in isotopic compositions of the two components and under ideal conditions is in the order of a few percent. Measuring the isotopic composition of water samples from the High Dam Lake, Egypt and adjacent groundwater, Aly et al. (1993) found that infiltration of lake water was limited to a distance of about 10 km from the lake.

Isotopes can also be used to identify the lack of recharge to an aquifer from surface water bodies. Kulkarni et al. (1999), by employing environmental isotopes and hydrochemistry, concluded that a limestone aquifer in the semi-arid part of Rajasthan State, India, did not receive recharge from the local lakes.

Rivers carrying water derived from precipitation at much higher altitude are isotopically depleted compared to local precipitation. Moreover, river water can show a seasonal variation, which usually is observed in wells near the river with reduced amplitude and after a time lag. The time lag as well as the change in the mean isotopic composition gives the minimum time (transit time) required for river water and possibly its dissolved pollutants to reach a groundwater supply well and the fraction of river water (possibly polluted) relative to other recharge sources. By measuring ^{18}O in a bank infiltration along a section about 100 m from the River Neckar near Heilbronn, Germany, estimated.a mean transit time of 7 to 42 days and fractions of river water from about 100% to 30%.

c. Palaeorecharge

Groundwaters in shallow aquifers typically have residence times of decades to hundreds of years. In contrast, deeper and less permeable aquifers that extend for many kilometres can have through-flow times of thousands of years. If the flow regime is simple and mixing is minimal, such aquifers can serve as archives of information about environmental conditions at the time of recharge. The stable isotopes of hydrogen and oxygen of the palaeowater reflect the air surface temperature and the air mass circulation at the time when it was formed by infiltrating precipitation (Rozanski et al., 1993). Palaeotemperatures derived from noble gas analyses complement, and are sometimes more meaningful, than those derived from oxygen-deuterium analyses.

In arid climates, river water may be enriched in ^{2}H and ^{18}O relative to groundwater, if it was replenished under former more humid conditions. For example, groundwater adjacent to the river Nile along its course in Sudan and Egypt is such palaeowater significantly depleted in ^{2}H and ^{18}O relative to the present Nile water. Making use of these differences in the isotopic compositions of the mixing components, Vrbka et al. (1993) have estimated the fraction of Nile water in groundwater collected along the river Nile between Khartoum and Atbara, Sudan. Assuming a mixture between palaeowater and Nile water, they found Nile water fractions between 40% and 80% at the different sampling sites. Palaeorecharge in the aquifers of arid regions of Jalore and Barmer districts (a part of the Thar Desert) in western Rajasthan, India was identified using environmental isotopes (Rao and Kulkarni, 1997). It was observed that deeper palaeowaters were depleted of stable isotope compositions compared to present-day precipitation. The shallow aquifers, on the other hand, were replenished by infiltration of water through river channels during episodic floods caused by sporadic rain events.

Stute et al. (1992) studied isotopic compositions of groundwater in the Carizzo aquifer in south Texas and found a 5°C to 6°C reduction in temperature during the glacial period, consistent with the snow-line estimates. Palaeohydrology studies based on isotope methods also indicated that long-term climate change has strongly influenced modern groundwater flow systems through changes of their boundary conditions, e.g. influence of the continental ice shield on groundwater recharge (Siegel, 1990 and 1991). Love et al. (1994) have demonstrated the effects of glacial/interglacial climate change on recharge in South Australia and Fontes et al. (1991) have used environmental tracers to show how shifts in the position of the Niger River since the end of the last glaciation have affected the distribution of groundwater recharge. Stute and Talma (1998) found that during the last glacial maximum southern Africa was characterised by 5°C to 6°C lower temperatures and by an expansion of the winter rainfall region that received its moisture from the Atlantic Ocean.

(ii) Groundwater transit time

The radioactive decay of environmental radioisotopes (Table 9.2) and the transient nature of some of them (bomb-^{3}H, anthropogenic ^{85}Kr, bomb-^{14}C and bomb-^{36}Cl) make these isotopes a unique tool for determining the groundwater residence time ('age'), i.e. the length of time the water has been isolated from the atmosphere (Davis and Bentley, 1982). In unconfined aquifers there is preferentially a vertical gradient of the isochronal surfaces (surface of constant residence time of the water), while in confined aquifers the dominating feature is a lateral gradient. It can be shown (Jordan and Froehlich, 1990) that in the former case this gradient is approximately proportional to the inverse of the recharge rate (cf. Equation (9.8)), while in the latter case the gradient is inversely related to the flow velocity. Therefore, the hydrogeologically relevant parameters primarily addressed by groundwater dating with radioactive isotopes are the recharge rate and flow rate of groundwater in unconfined and confined aquifers, respectively.

One of the approaches to determine groundwater flow rate by a suitable radioisotope is based on measuring the decrease in radioisotope concentration along the flow path. If radioactive decay alone is governing the decrease in radioisotope concentration, then the radioactive decay law yields for groundwater flow time $t_2 - t_1$ between two wells (well 1 and 2):

$$t_2 - t_1 = \frac{T_{1/2}}{\ln 2} \cdot \ln\left(\frac{C_1}{C_2}\right) \tag{9.9}$$

where C_1 and C_2 are the radioisotope concentration at wells 1 and 2, and $T_{1/2}$ is the half-life of a radioisotope used. Consequently, the groundwater flow velocity (v) is given by:

$$v = \frac{x_2 - x_1}{t_2 - t_1} \tag{9.10}$$

where, $x_2 - x_1$ is the distance between the two wells. With porosity (n) of an aquifer, the groundwater flow rate (V_f) (volume of water flowing per unit time across the unit cross section of an aquifer) can be estimated by:

$$V_f = n \cdot v \tag{9.11}$$

This simple approach requires access to at least two wells along the flow path in an aquifer. Often, however, the flow direction is not well known or only one well is available for sampling. In these cases the 'absolute' age (t) needs to be determined rather than the difference of ages at two wells (flow time). The respective relationship for 'absolute' groundwater age is obtained from Equation (9.9) by replacing the C_1 by C_0, setting $C_2 = C$, $t_2 = t$ and $t_1 = t_0 = 0$. Therefore, this approach requires knowledge of the initial concentration C_0 of the respective radioisotope. The reconstruction of C_0 is one of the major problems in groundwater dating with radioisotopes.

Under natural conditions, groundwater movement is generally very slow, often a few metres per year. Therefore, after a distance of a few kilometres along the flow path, the groundwater is already old and its age is beyond the dating range of ^{3}H, $^{3}H/^{3}He$ and CFCs. The most common radiometric approach to determining groundwater residence times in large aquifers has been ^{14}C. Its half-life of 5,730 a makes it a suitable tool for the dating of groundwater in an age range from about 1 to 40 ka (Geyh, 2000).

An integrated mass balance approach may be used to understand geochemical and isotopic evolution of groundwater and factors affecting the use of ^{14}C for groundwater dating. Plummer et al. (1994) developed an interactive geochemical model NETPATH (available through Internet on the USGS site) for ^{14}C dating, based upon the geochemistry along a flow path and boundary conditions provided by the user. However, such geochemical models are based on the assumption that CO_2 produced in the root zone alone is the source of ^{14}C in infiltrating water. The ^{14}C content of this CO_2 is assumed to be nearly in equilibrium with the atmosphere. Recent studies (Aggarwal and Dillon, 1998; Affeck et al., 1998; Bacon and Keller, 1998) indicate that carbon dioxide may also be produced in deeper parts of the unsaturated zone and near the capillary fringe. The ^{14}C content of this CO_2 would be much different, and generally lower than that in the root zone. The initial ^{14}C content of groundwater, therefore, can be much lower than that at atmospheric equilibrium without radioactive decay or water-rock interaction. Under such conditions, the usual corrections for dating purposes are invalid.

Radiocarbon dating of groundwater using ^{14}C of dissolved organic carbon (DOC) has also been developed (Murphy et al., 1989). The DOC of groundwater is derived in the soil zone, but, unlike the TDIC, the DOC remains unaffected by dilution through carbonate reactions, and so provides an independent method of ^{14}C groundwater dating. Wassenaar and co-workers (1992) examined the advantages and disadvantages of the DOC against the TDIC method. Murphy et al. (1989) presented a detailed analysis of the interrelationship between carbon isotopes, carbon species chemistry, and microbial activity on the dissolved organic and inorganic components in the Middendorf aquifer. More recently, Artinger et al. (1995) discussed further methodological aspects, emphasising that DOC dating should be focused on the fulvic acid component of DOC. In groundwater with a high DOC concentration (>100 mg/L), they also found that ^{14}C-free fulvic acid could be released from sedimentary organic carbon (e.g. lignite) to dilute soil-derived fulvic acid. It is expected that improved organic geochemical methods may in future be available to correct ^{14}C for subsurface contributions.

Very slow moving groundwater in deep confined aquifers extending over tens and, in some cases, several hundreds of kilometres, can reach ages of tens and even hundreds of thousands of years. These ages are beyond the dating range of ^{14}C and require the use of very long-lived radioisotopes. Of the three long-lived radioisotopes ^{81}Kr, ^{36}Cl and ^{129}I (Table 9.2), only ^{36}Cl has found wider practical use so far. Interpretation of ^{36}Cl data in terms of groundwater age is often hampered by insufficient knowledge of *in situ* production (Andrews and Fontes, 1992). ^{36}Cl combined with other dating tools has been used to study the Great Artesian Basin aquifer system in Australia (Torgersen et al., 1991; Love et al., 2000). Application of the other two radioisotopes is still in an exploratory stage. Recently, Collon et al. (1999) reported the first applications of ^{81}Kr to groundwater dating in the Great Artesian Basin of Australia. They obtained mean groundwater residence times of 225,000 to 400,000 year. The interpretation of these values in comparison with available ^{36}Cl dates is still being worked on. The authors concluded that the results confirm ^{81}Kr as a reliable tool to date old groundwaters.

Among the various radionuclides of the natural decays series, combined use of the uranium isotopes ^{238}U and ^{234}U has shown some potential to date old groundwater (Ivanovich and Harmon, 1992). Froehlich and Gellermann (1987) reviewed various attempts made by several authors in using these isotopes, especially the $^{234}U/^{238}U$ activity ratio, as a dating tool. They concluded that the major constraint of this approach comes from sorption-desorption reactions, which can reduce the 'lifetime' (the reciprocal value of the radioactive decay rate λ) of uranium in the liquid phase more than one order of magnitude below the radioactive lifetime of ^{234}U ($2.45 \cdot 10^5$ a). Therefore, the applicability of the uranium isotope dating depends on the geochemical conditions in a groundwater system. One of the practical applications of the ^{234}U:^{238}U disequilibrium method is found in the Milk River aquifer study (Ivanovich et al., 1991). The ages derived from the uranium isotopes were in reasonable agreement with the ^{36}Cl dates and the hydraulically calculated values (Froehlich et al., 1991). Musgrove and Banner

(1993) have used ^{87}Sr, $^{234}U:^{238}U$ and other heavy element isotope analyses, more commonly associated with rock than water, to trace the sources and rates of a deep groundwater flow system in the United States of America.

Among the noble gases directly or indirectly produced in groundwater and in the aquifer material by radioactive decay, ^4He has proved to be the most useful for groundwater studies. This is mainly due to the relatively rapid rate of ^4He production compared to other noble gases. ^4He is formed by neutralisation of α particles produced in α-decay processes in natural decay series (^{238}U, ^{232}Th and ^{235}U). A fundamental advantage of using decay products such as ^4He for dating is the fact that as time progresses, more ^4He will be present in groundwater and its measurement will become easier. This is in contrast with the atmospheric radionuclides, which will become more difficult to detect as the age of groundwater increases. The half-life of most α-producing radionuclides is more than two orders of magnitude than the age of the oldest groundwater which might be dated. Thus, groundwater in homogeneous, confined aquifers often experiences a nearly linear increase in ^4He with time. However, the measured ^4He concentrations usually exceed the values expected from this *in situ* production P (typical values in sandstones are between 10^{-12} and 10^{-11} cm^3 STP $^4He/g \cdot a$). This is mainly due to diffusion of crustal helium produced in strata above and below an aquifer in question. Assuming a confined aquifer of the thickness H, effective porosity n, and a crustal ^4He flux F (often between about 1 and $6 \cdot 10^{-6}$ cm^3 STP $^4He/cm^3 \cdot a$), the groundwater age t follows from the equation:

$$[^4He] = (P + \frac{F}{nH} t) \tag{9.12}$$

where, $[^4He]$ is the concentration of ^4He in the groundwater. A typical field study using this tool, was performed in the Great Artesian Basin, Australia (Torgersen et al., 1991).

It can be seen that there are various radioisotopes available to date groundwater and thus study its flow dynamics. Given the uncertainties inherent in each of the methods for groundwater dating, in groundwater studies as many methods should be used, as possible.

(iii) Interconnections between aquifers

Both the groundwater dynamics within a system of aquifers and the risk of groundwater contamination can be influenced by hydraulic interconnections between aquifers. Environmental isotopes, especially the stable isotopes, can be used to investigate such interconnections, provided the isotopic composition of the groundwater in the aquifers is different.

A typical example of this approach is the identification of a hydraulic connection between the 'Continental intercalaire', a large Lower Cretaceous aquifer in northwestern Africa, with the overlying 'Complexe terminal' aquifer near the Gulf of Gabes in Algeria (Gonfiantini et al., 1974a). The $\delta^{18}O$ and δ^2H values of the groundwater in the confined part of the 'Continental intercalaire' were found to be relatively uniform; they were respectively about 3 and 20‰ lower than in the 'Complexe terminal'. In two areas, however, these stable isotopes provided a clear fingerprint that the 'Continental intercalaire' discharged into the overlying aquifer, in spite of the fact that the two aquifers are separated by a clayey formation several hundreds of metres thick. It was found that in these areas the groundwater of the 'Complexe terminal' had the same isotopic composition as the 'Continental intercalaire'. This finding confirmed earlier expectations that the El Hamma fault system provides a hydraulic connection through the rather thick aquiclude.

Intense exploitation of aquifers can cause leakage from overlying and underlaying aquifers. Payne et al. (1980) reported a typical case study related to a vertically stratified aquifer system at Hermosillo, northwestern Mexico. Here, an alluvial aquifer is underlain by a low-permeability clay formation of about 100 m thickness. Below this clay layer is a confined aquifer with a piezometric head about 30 m above the water table of the overlying alluvial aquifer. The groundwaters of these two aquifers had different stable isotope compositions. Heavy exploitation of the alluvial aquifer caused a vast cone of depression and induced leakage from the deep aquifer into the shallow one. The stable isotope data indicated that contribution of deep groundwater into the shallow aquifer was up to 20%.

Isotopes can also be used to prove a lack of hydraulic interconnections between aquifers. Ho et al. (1991) applied the stable isotope approach in a study of the Mekong Delta aquifer system, Vietnam. They found that the upper Pleistocene aquifer was hydraulically isolated from the lower Pliocene aquifer.

(iv) Groundwater salinisation

In many situations it is necessary to identify the mechanism of salinisation in order to prevent or alleviate the cause. Salinity may be used as a practical tool in water resource investigation to estimate recharge and discharge, to investigate palaeohydrology, and in the understanding and management of groundwater in coastal regions (Edmunds and Droubi, 1998). Isotope techniques can be used to distinguish the importance of the following processes which may lead to salinisation of groundwater:

- leaching of salts by percolating water; the origin of salts may be evaporite deposits, aeolian salts, or the products of weathering,
- intrusion, present or past, of salt water bodies such as sea water, brackish surface water, brines and connate water,
- concentration of dissolved salts by evaporation.

For example, the stable isotope composition of salt springs of the Sarvistan and Caspian Gates in Iran was found to be the same as that of local groundwater. It was concluded that the salinity originated through dissolution of rock salt beds by infiltrating recharge waters (Zak and Gat, 1975). Isotope geochemical methods have been used extensively to demonstrate the evaporative origins of saline groundwaters in inland basins, for example in chotts of Algeria (Gonfiantini et al., 1974b), in the alluvial aquifers of the Indo-Gangetic Plain in Haryana, India (Kulkarni et al., 1996). Combinations of geochemical techniques and various ratios may be used to determine salinity of different ages and origins. The isotope geochemical approach was applied to study the origin of salinity in the coastal, alluvial, multi-aquifer system in the Mahanadi Delta, India (Kulkarni et al., 1998). The findings indicated that the groundwater salinity in aquifers originated due to transgressions of sea during the Late Pleistocene and the Holocene.

The origin of chloride in groundwater can be traced by the stable chlorine isotope ratio, which also can be applied to investigate the evolution of deep formation waters (Eggenkamp, 1994). A multi-component approach using concentrations of major, minor, and trace elements along with isotope compositions (^{2}H, ^{3}H, ^{11}B, ^{13}C, ^{14}C, ^{18}O, ^{34}S, ^{37}Cl, ^{87}Sr, etc.) would be best suited for studying the origin of salinity. However, some of these techniques are still in an exploratory stage.

(v) Groundwater pollution

Pollution of aquifers by anthropogenic contaminants is one of the problems of high concern in the management of water resources. Environmental isotopes can be used to trace the pathways and predict spatial distribution and temporal change of pollutants in aquifers. This information is critical in order to be able to understand the source of contaminants, assess their scale and migration, and plan for remediation.

Measurements of concentration and stable isotope composition of sulphate and nitrate in groundwaters have been widely used to identify the sources of sulphate/nitrate pollution as well as the microbial sulphate reduction and denitrification processes, respectively. For example, Kendall (1998) studied the sources of nitrate in catchment areas during the snow melt season in spring. In many of the catchment areas an increase of nitrate concentration was observed in the early spring. The nitrogen isotope data provided evidence that the soil nitrate is a major source of the observed increase rather than atmospheric deposition, as previously assumed. Boettcher et al. (1990) quantitatively assessed the microbial denitrification in a sandy aquifer in northern Germany. They found a linear relationship between the $\delta^{18}O$ and $\delta^{15}N$ values of residual nitrate of groundwater where denitrification was identified. Various applications of sulphur isotopes have been discussed in detail by Pearson and Rightmire (980) and by Krouse (1980).

Concentration and stable isotope composition of hydrocarbons and their degradation products can together provide a powerful tool for pollution assessment and remediation. For example, Aggarwal and Hinchee (1991) used the $\delta^{13}C$ of soil CO_2 to substantiate aerobic biodegradation of hydrocarbons at three sites contaminated with jet fuel in the United States of America. Kelley et al. (1997) measured the concentrations and $\delta^{13}C$ values of benzene, toluene, ethylbenzene and xylene under bioremediation conditions at a gasoline-contaminated site in southern California. Their data suggest the presence of at least two sources of gasoline contamination (leaded and unleaded gasoline) at the investigation site. Aggarwal et al. (1997) were the first to demonstrate in a laboratory study the combined use of oxygen and carbon isotope analyses of carbon dioxide and molecular oxygen, respectively, for monitoring the biodegradation of fuel hydrocarbons. They concluded that the combined use of stable carbon isotope composition of carbon dioxide and oxygen isotope composition of the electron acceptors (molecular oxygen, nitrate, or sulphate) provides a robust tracer for the verification and quantification of microbiological processes associated with hydrocarbon-contaminated groundwater.

9.2.4 General remarks on environmental isotopes

Environmental isotopes are a supplementary tool for hydrogeological investigations and should be employed as an integral part of studies. The amount and quality of information that they are able to provide is strictly dependent on the degree of hydrogeological knowledge about the area under investigation and also on the knowledge of the investigator. Generally, an integrated approach using isotope, hydrogeological, and hydrochemical data will lead to the optimum use of these techniques and to a logical interpretation.

In studying the isotope content of groundwater, a major requirement is to obtain a representative and uncontaminated sample. For stable isotopes and tritium, relatively small samples are required: 20 mL are sufficient for stable isotopes, 500 mL for tritium. Containers must be airtight to avoid evaporation and exchange leading to alteration of isotope composition. Polyethylene or polypropylene bottles with inner cones in their screw caps are generally used. Bottles should be filled completely and care taken to minimise exposure of samples to the atmosphere. ^{14}C sampling is carried out by separation of dissolved carbonate species from a large volume of water (usually 50–100 L) to yield a few grams of carbon, preferably at the sample site. The details of the sampling procedures, storage and precautions are discussed by Clark and Fritz (1997).

Furthermore, isotopes can trace dispersion and infiltration of pollutants in landfills as well as quantify degradation and migration of pollutants. Isotopes are also applied extensively to study: leakages from dams and reservoirs, stream flow measurements, effluent dispersion, suspended sediment and bedload movement in ports and harbours, lake dynamics, sedimentation in lakes and reservoirs, geothermal systems, glaciology, etc. Various monographs and textbooks describe these applications.

9.3 Artificial isotopes

Artificial radioisotopes can be measured in extremely low concentrations and often *in situ*, making possible the design of convenient and efficient field experiments. However, safety precautions, special equipment and trained personnel are required for handling radioisotopes. Therefore, artificial radioisotope should be considered when other tracers such as fluorescent dyes would not meet the needs of a problem. Fluorescent dyes became the first option for many surface water and karst system studies. The main limitation for their use is that water samples need to be measured outside the system, either with portable equipment or in the laboratory. Some recent developments may overcome this difficulty. Artificial radioisotopes have added advantage in that they can be used in a variety of conditions and environments and the quantity required is very small.

Radioactive tracers can be used to determine the following parameters:

1. aquifer characteristics: (a) porosity, (b) transmissivity, (c) dispersivity;
2. direction and velocity of groundwater flow;
3. stratification of aquifers.

9.3.1 Radioactive tracers

When choosing radioactive tracers the following points should be taken into consideration:

- The isotope should have a half-life compatible to the presumed duration of the observations. The unnecessary use of long-lived isotopes will leave these isotopes in the water for a long time, creating a persistent health hazard and potential interference if the experiment needs to be repeated.
- The isotope should not be adsorbed by the inorganic or organic solids in the aquifer.
- It is advantageous to measure the concentration of radioactive tracers in the field. For this reason γ-emitters are generally preferred.
- The isotope must be available at a reasonable cost when and where required.

Frequently used radioactive tracers are listed in Table 9.3.

The chemical form of the radionuclide used in water tracing experiments plays an important role. Anions are very useful as water tracers because usually they are not adsorbed onto a matrix. In general, cations are not good tracers. In the form in which they are commonly delivered, they may be subject to strong adsorption, interfering with the validity of their tracing of the water mass. To minimise this effect within an aquifer, cationic tracers need to be transformed into neutral molecules or anionic compounds. For this, they are combined with complexing agents. The most commonly used in hydrological practice is a chelated metallic compound formed by means of ethylene diamine tetracetic acid (EDTA).

Although not fully meeting the half-life and radiation requirements mentioned above, tritium is an ideal marker for water since it is incorporated in the water molecule. Hence it is used in spite of the fact that *in situ* detection is not possible and samples must be analysed in the laboratory subsequent to an experiment. In certain cases, tritiated water (HTO) may suffer some delay in comparison with water bulk velocity as a result of ion exchange with clays (Kaufman and Orlob, 1956; Knutsson and Forsberg, 1967). Minimal amounts of artificial tritium should be used to minimise interference with groundwater resource studies based on environmental tritium.

Table 9.3 Radioisotope tracers commonly used in groundwater investigations

RADIONUCLIDE	^{3}H (T)	^{51}Cr	^{82}Br	^{99m}Tc	^{131}I	^{198}Au
CHEMICAL FORM	HTO	Cr-EDTA	Br^-	TcO_4^-	I^-	$HAuCl_4$
HALF-LIFE	12.43 y	27.8 d	35.7 h	6.0 h	8.05 d	64.8 h
GAMMA RAY ENERGY AND YIELD (%)	No gamma	0.324 (9%)	0.55 (71%)	0.140 (99%)	0.08 (2%)	0.41 (99%)
			0.62 (43%)	0.142 (1%)	0.28 (5%)	0.68 (1%)
			0.70 (28%)		0.36 (82%)	
			0.78 (83%)		0.64 (9%)	
			0.83 (24%)			
			1.04 (27%)			
			1.32 (27%)			
			1.47 (17%)			
AIR KERMA-RATE ϒ constant, *µGy .m² .h -1.GBq -1*		4.86	432,4	20..3	66.8	67.6
DERIVED MAXIMUM CONCENTRATION *(kBq/m³ (1)*	125000	62000	5000	125000	100	1240
MINIMUM DETECTABLE CONCENTRATION *(kBq/m³)*	18.5 (2) 1.1 (3)	148 (4)	7.4 (4) 0.18 (5)	7.0 (4)	14.8 (4)	12.9 (4)

(1) Maximum concentration in drinking water for members of the public
(2) Direct measurement using a liquid scintillation counter
(3) Liquid scintillator counting after electrolitic enrichment
(4) Direct measurement using a NaI(T1) detector (2˝x2˝ Ø) submerged in the water with saturation geometry

9.3.2 Techniques

The techniques most commonly used to determine aquifers parameters are (Margrita and Gaillard, 1991):

- Single-well methods in natural flow conditions;
- Multi-well methods in natural flow conditions;
- Multi-well methods in pumping conditions.

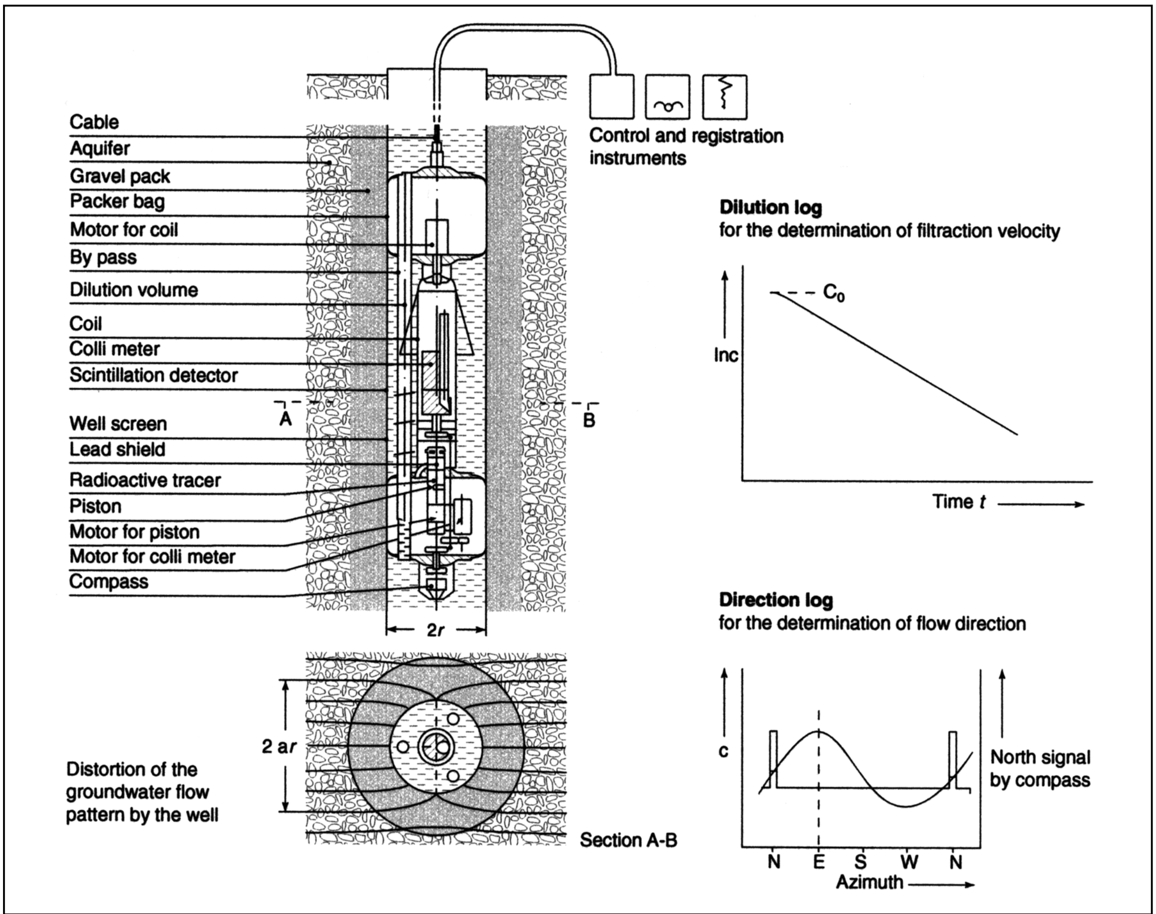
In the first case, the parameters that can be quantified are Darcy velocity, its vertical profile and vertical velocity.

The introduction of tracer into a borehole may be done by pouring through a thin pipe, by crushing an ampoule at the depth of interest, or by using a special injection device. Injection can be performed at one or several depths to facilitate mixing of the tracer solution with the standing water of the borehole. After the release and mixing, the radioactivity in the borehole is measured by a probe, usually a scintillation counter but sometimes a robust Geiger counter, inserted at the chosen depth. Specialised probes have been constructed which seal off a defined volume of the borehole by inflatable packers, release radiotracer into this volume, and measure the radioactivity by detectors in or immediately above the release volume. Figure 9.4 shows a probe that can measure both tracer dilution with time and direction of flow (Drost, 1989).

In the Single-well Dilution Technique (Halevy et al., 1967), the dilution rate of the tracer by the natural flow of water through the borehole is observed. In the Single-well Pulse Technique (Borowczyk et al., 1965), tracer is forced into the aquifer by pumping water into the well and then retrieved by pumping water out of the well.These single-well techniques have been widely used in the past decades to obtain local parameters of aquifers. (Drost and Neumaier, 1974; Tazioli, 1977; Plata Bedmar, 1981, 1983; Mandel et al., 1985; Drost, 1989).

When movement of water between boreholes is to be observed (this is the Multiple-well Technique) tracer is injected in a borehole and detected in another one. Under natural flow conditions, direction of groundwater flow needs to be known. For this, several holes are drilled around the injection point to intercept tracer movement. The parameters to be obtained in this

Figure 9.4 Tracer probe for dilution and direction logging (Modified after Drost, 1989)



test are velocity, longitudinal and transverse dispersion coefficients, and kinematic porosity. The time scale of such an experiment is usually rather long when groundwater velocity is low. The pumping method is usually preferred in this case. The observation well is pumped and the radioactivity of the pumped water monitored at the surface (Halevy and Nir, 1962). It is important that such pumped water be disposed of in such a way that it cannot return inadvertently to the system under investigation. In some experiments, water pumped from the observation well is returned to the injection well intentionally to establish a closed circuit. The pumping flow must be chosen so that the velocity induced by pumping is greater than the natural velocity of the aquifer. It is assumed that the isochrone line around the pumping piezometer is circular (Margrita and Gaillard, 1991).

9.3.3 Applications

(i) Infiltration studies in the unsaturated zone

Information on the movement of water in the unsaturated zone is needed whenever one is trying to set up a water balance, either for the unsaturated zone itself or for the groundwater in the saturated zone beneath it.

Radioactive tracers are applied to study the movement of infiltrating water, and sometimes of pollutants, in the unsaturated zone. The technique includes the injection of one or more tracers at a certain depth, either as a point source or in a horizontal plane, and the measurement of the displacement of the labelled horizon after a certain time period. To determine the infiltration rate and other transport parameters, it is required to determine the soil moisture content and the tracer concentration in the samples collected along a depth profile.

Tritiated water and other radiotracers have been widely used to determine recharge rates under different climatic conditions (Zimmermann et al., 1966, 1967; Datta et al, 1978; Nair et al., 1979). Athavale et al. (1980) studied the variability of recharge in several regions in India under different climatic conditions and soil types. If the main soil types are properly represented, this approach can be used to determine the regional water balance at local or regional scales.

Multi-tracer experiments can be used to study the movement of pollutants, by determining the retardation factor of different solutes compared to infiltration rate of water. In some cases, organic compounds (pesticides) or fertilisers are labelled with other artificial isotopes like ^{15}N, ^{14}C, ^{32}P, ^{22}Na, etc. (Nicholls et al., 1982).

(ii) Effective porosity of an aquifer

The principle of the method of porosity determination in the saturated zone is based on the approximate equality of porosity ρ (Void volume/Total volume) and partial volume of water s equals Volume of water/Total volume.

Tracer is introduced into a well and a second well, at a distance r, is pumped. Ignoring the dispersion of tracer on its path between wells, its arrival at the second well signifies that the volume of water pumped, V, is

$$V = \pi r^2 b s \qquad (9.13)$$

where b = aquifer thickness, r = distance between wells and s = effective porosity. The prime requirements are that the distance between the wells be large compared to the aquifer thickness, ($r > b$), radial pumping velocities be larger than the natural velocities, and the cone of depression at the pumping well be small compared to volume of water pumped. A typical experiment of porosity determination has been fully described by Halevy and Nir (1962).

(iii) Transmissivity

The transmissivity T, first introduced by Theis in his non-equilibrium equation, characterises the ability of an aquifer to transmit water (Hantush, 1963). Along with the storage coefficient, hydraulic conductivity, specific storage and specific yield, transmissivity is one of the formation parameters characterising basic hydraulic properties of an aquifer. It is related to the hydraulic conductivity K (mean filtration coefficient) and aquifer thickness, b, as,

$$T = bK \qquad (9.14)$$

Mercado and Halevy (1966) have found that the volume of water pumped during the experiment referred to in section 9.3.3.1, measured until the tracer peak arrives at the next well, is inversely proportional to the value of transmissivity. They have determined transmissivity of two layers of the same aquifer by means of two injections and one observation well. Transmissivity then was found from the relationships:

$$\begin{aligned} V_1 &= \pi r_1^2 b_1 s_1 T/T_1 \\ V_2 &= \pi r_2^2 b_2 s_2 T/T_2 \qquad (9.15) \\ T_1 + T_2 &= T \end{aligned}$$

where T is total transmissivity; $T_{1,2}$ is partial transmissivity of the layers 1 and 2; $r_{1,2}$ are distance between injection and observation wells; $b_{1,2}$ are thickness of the layers 1 and 2; $s_{1,2}$ are effective porosity of the layers 1 and 2.

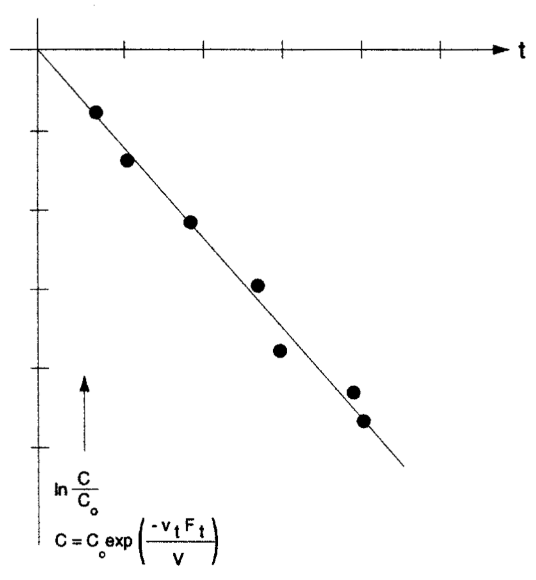
(iv) Dispersivity

In problems of artificial recharge of aquifers by water of a lower quality, or the disposal of wastes into the groundwater, a subject of interest may be the coefficient of dispersivity, D, occurring in the tracer transport equation (Halevy et al., 1967). This value characterises the mixing property of an aquifer. It is commonly estimated by finding the theoretical curve which best fits the experimental tracer breakthrough curve in an observation well. Parameters on such theoretical curves are dispersion coefficients derived from the mathematical model of the tracer transport (Lenda and Zuber, 1970).

(v) Groundwater flow velocity

Groundwater flow velocity under the natural hydraulic gradient is generally evaluated from Darcy's formula. Using a tracer, it is possible to measure directly the groundwater velocity, and from this to evaluate aquifer permeability. The method consists of injecting tracer into the well and then following its concentration decrease with time (Fig. 9.5).

Figure 9.5 Schematic diagram showing the tracer concentration changes with time in a borehole dilution experiment (Moser et al., 1963)



The principle applies for any tracer, but again radioactive isotopes are easier to detect *in situ* with a high degree of accuracy, at very low concentrations (Moser et al., 1957, 1963; Mairhofer, 1963; Guizerix et al., 1963). The disadvantages of the method are that the results are valid only in a restricted area neighbouring the well. Several measurements in different wells and at different depths are often necessary to establish a good picture of the groundwater flow in a given aquifer.

The horizontal filtration velocity V_f of groundwater (also called Darcy velocity) is given by:

$$V_f = -\frac{V}{\alpha\, Ft} \ln \frac{[C]}{[C_0]} \tag{9.16}$$

where, V is the measuring volume (the borehole volume in which dilution takes place), F is the cross section of the measuring volume perpendicular to the direction of the undisturbed groundwater flow, and t is the time interval between measurement of concentrations $[C_0]$ and [C]; α is the correction factor accounting for the distortion of flow lines due to the presence of a borehole. The calculation of the value of α on the basis of well construction data is discussed by Halevy et al. (1967). The above expression is a particular solution of the differential equation describing dilution rate of the tracer.

In practice, several readings are required to derive results not affected by incomplete mixing processes and residual currents. The lower limit of velocity, mainly dictated by the diffusion rate, is 10^{-7} m/s.

If effective porosity, s, is known one can estimate the bulk velocity, V_{bulk}, as:

$$V_{bulk} = \frac{V_f}{s} \tag{9.17}$$

Potential (hydraulic) gradients, dh/dx, can be determined from the vertical distribution of radioactivity in the observation wells if these boreholes are entirely cased (Weinberger, et al., 1967). Knowing the potential gradient, one can also deduce the hydraulic conductivity, K, applying the Darcy law:

$$K = \frac{V_f}{dh/dx} \tag{9.18}$$

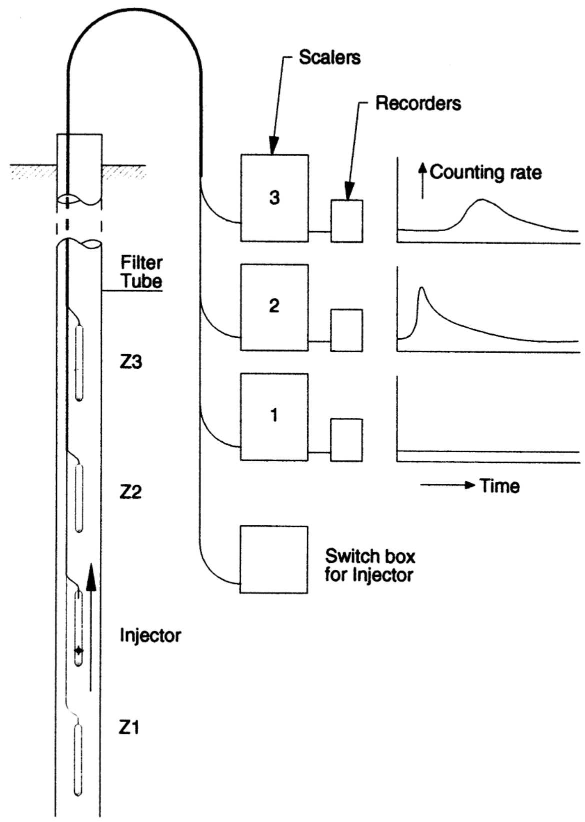
Knowledge of vertical velocity component in a borehole is important in such problems as determination of the amount of water entering or leaving the well from different permeable strata, or amount of interchange between zones. A single-well dilution technique may be applied. A series of detectors placed at different depths is commonly used for such measurements (Fig. 9.6).

The method is specially recommended when expected vertical flow velocities are less than 2.10^{-2} m/s; in this region, due to the influence of friction, tracer methods are more efficient than the use of mechanical current meters. The lower velocity limit is estimated at 10^{-7} m/s. In turbulent flow, accuracy of measurement is reported to be better than 10% but less is known about the accuracy in laminar flow (Halevy et al., 1967).

(vi) Direction of groundwater flow

The multi-well technique can be used to determine the direction of groundwater flow, especially when there is radial symmetry in the well location, or if there are observation wells in the approximate direction of flow. The direction of flow, however, may also be determined by means of a single-well technique. Tracer is added to a segment of a borehole and is carried away in the direction of the flow. Activity is then detected by a special directionally oriented probe rotated by means of a stiff metallic rod for rather shallow wells, or in more sophisticated devices, by means of a motor included in the probe. A diagram like that of Fig. 9.7 is then obtained, which shows the direction of groundwater flow. In this case, contrary to the flow velocity measurements, radioisotopes that are relatively rapidly absorbed by the formation are the most appropriate ones to use.

Figure 9.6 Arrangement of a set of detectors for determining vertical flow in a borehole (Moser et al., 1963)



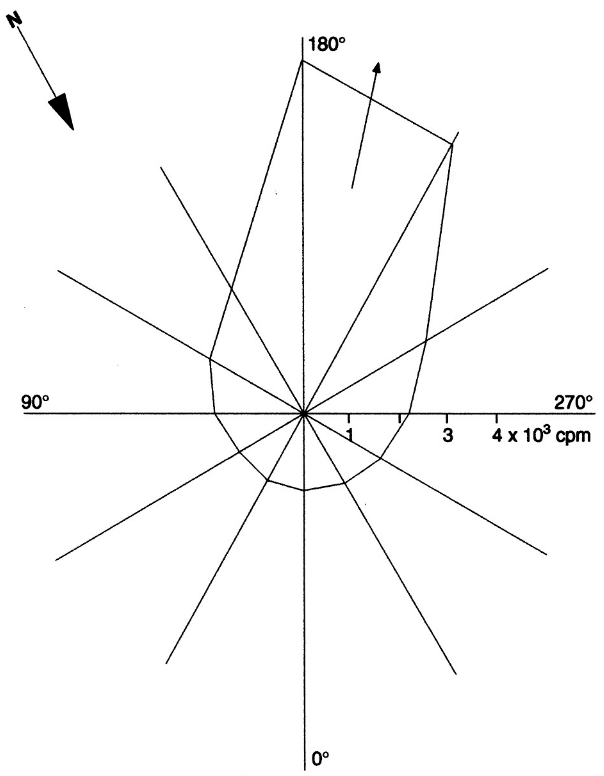
^{198}Au in the form of chloride ($^{198}AuHCl_4$) and ^{51}Cr in the form of chloride ($^{51}CrCl_3$) are commonly used. In most laboratory tests the difference between measured and true flow direction was less than 3°. The reproducibility of the method in the field is better than 10%. A simplified version of the method has been used by Wurzel and Ward (1965). A cylindrical metal gauze placed in the well at the injection point adsorbs radioactive tracer ($^{51}CrCl_3$). The gauze is removed, cut in many sections parallel to the axis and the activity of each section is measured in the laboratory. The section giving the highest activity indicates the direction of water flow.

9.3.4 Practical considerations

In the past decades, the use of isotope tracer methods for the determination of parameters of groundwater flow has gradually increased. The most fruitful field of application has been the solution of local problems in civil engineering, in feasibility studies, site investigations, groundwater pollution studies, and in the design, maintenance and control of hydraulic works (Drost, 1989).

The results of isotope tracers compare well with other conventional techniques and they can give more detailed information. Values obtained from pumping tests represent the mean hydraulic properties over the whole aquifer thickness. Dilution tests may arrive at the variations of hydraulic conductivity with depth (Drost, 1989). As a result of dilution logs and pumping tests in more than 20 unconfined porous aquifers, the hydraulic conductivities derived from dilution logs, K (D), and those from pumping tests, K (P), fit a regression line (Fig. 9.8):

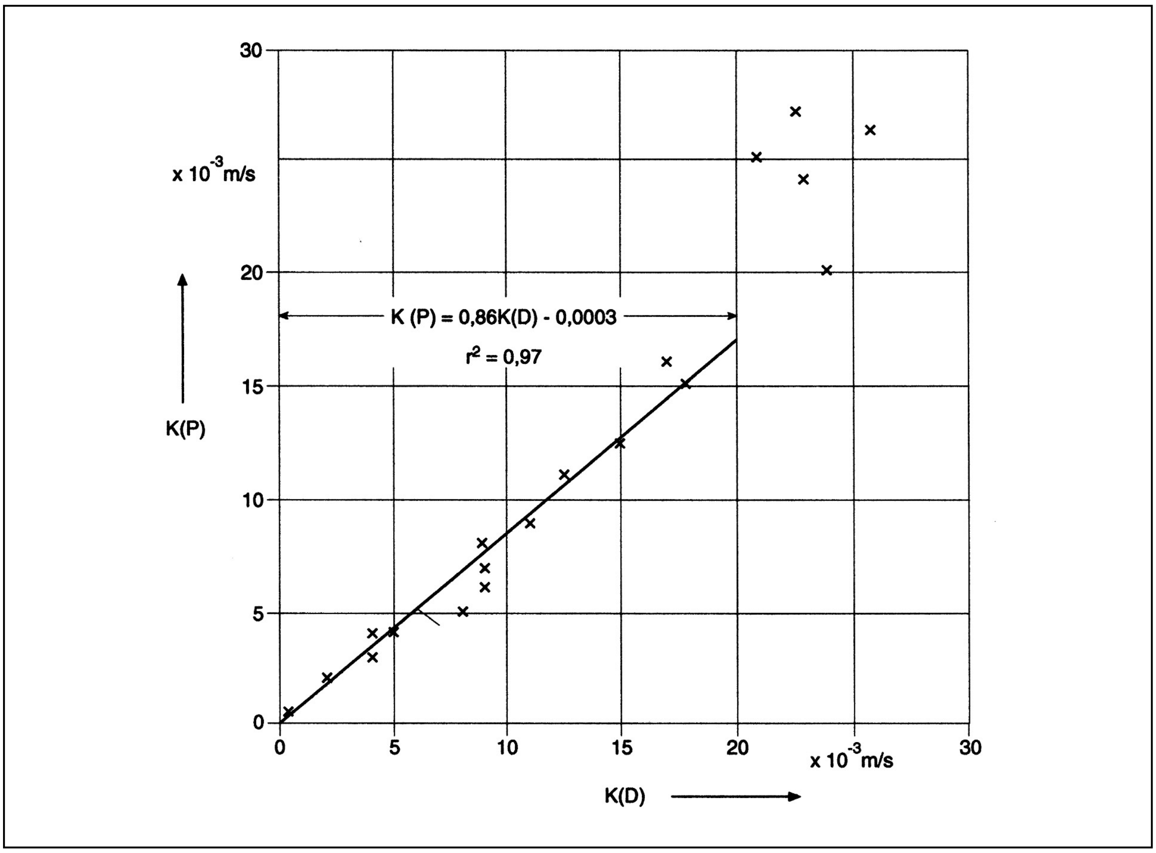
Figure 9.7 Diagram of water flow direction in a borehole. Activity injected, 0.6 mCi (Halevy et al., 1967)



$$K(D) = 1.15 \cdot K(P) + 0.0003, (r^2 = 0.97) \quad (9.19)$$

Since a commercial service in the application of these techniques is seldom available, it is advisable to seek the co-operation of universities or research institutes in the initial use of methods. Some difficulties in performing field measurements may be expected until the investigator becomes experienced in the new techniques involved. The experimenter must acquaint himself with all safety and legal requirements connected with the handling of radioactive materials.

Figure 9.8 Unconfined porous aquifers. Relation of hydraulic conductivities determined by tracer dilution techniques (K(D)) and by pumping tests (K(P)) (Drost, 1989)



9.4 References and additional reading

AFFEK, H. P.; RONEN, D.; YAKIR, D., 1998. About Production of CO_2 in the Capillary Fringe of a Deep Phreatic Aquifer. *Water Resour. Res.*, Vol. 34, pp. 989–96.

AGGARWAL, P. K.; HINCHEE, R. E., 1991. Monitoring *in Situ* Biodegradation of Hydrocarbons by Using Stable Carbon Isotopes. *Environ. Sci. Technol.*, Vol. 25, pp. 1178–80.

AGGARWAL, P. K.; FULLER, M. E.; GURGAS, M. M.; MANNING, J. F.; DILLON, M. A., 1997. Use of Stable Oxygen and Carbon Iisotopes Analyses for Monitoring the Pathways and Rates of Intrinsic and Enhanced in Situ Biodegradation. *Environ. Sci. Technol.*, Vol. 31, pp. 590–6.

AGGARWAL, P. K.; DILLON, M. A., 1998. Stable Isotope Composition of Molecular Oxygen in Soil Gas and Groundwater: A Potentially Robust Tracer for Diffusion and Oxygen Consumption Processes. *Geochim. Cosmochim. Acta*, Vol. 61, pp. 577–85.

ALY, A. I. M.; FROEHLICH, K.; NADA, A.; AWAD, M.; HAMZA, M.; SALEM, W. M., 1993. Study of Environmental Isotope Distribution in the Aswan High Dam Lake (Egypt) for Estimation of Evaporation of Lake Water and its Recharge to Adjacent Groundwater. *Environ. Geochem. and Health*, Vol. 15, pp. 37–49.

ANDRES, G.; EGGER, R. ,1985. A New Tritium Interface Method for Determining the Recharge Rate of Deep Groundwater in the Bavarian Molasse Basin. *J. Hydrol.*, Vol. 82, pp. 27–38.

ANDREWS, J. A.; FONTES, J. CH., 1992. Importance of the in Situ Production of ^{36}Cl, ^{39}Ar, and ^{14}C in Hydrology and Hydrogeochemistry. In: *Isotope Techniques in Water Resources Development*, International Atomic Energy Agency, Vienna, pp. 245–69.

ARTINGER, R.; BUCKAU, G.; KIM, J. I.; GEYER, S.; WOLF, M., 1995. Influence of Sedimentary Organic Matter on Dissolved Fulvic Acids in Groundwater: Significance for Groundwater Dating with ^{14}C in Dissolved Organic Matter. In: *Isotopes in Water Resources Management*, Vol. 1, International Atomic Energy Agency, Vienna, pp. 57–72.

ATHAVALE, R. N.; MURTI, C. S.; CHAND, R., 1980. Estimation of Recharge to the Phreatic Aquifers of the Lower Maner Basin, India, by Using the Tritium Injection Method. *J. Hydrol.*, Vol. 45, pp. 185–202.

BACON, D. H.; KELLER, C. K., 1998. Carbon Dioxide Respiration in the Vadose Zone: Implications for Groundwater Age Dating. *Water Resour. Res.*, Vol. 34, pp. 3069–77.

BOETTCHER, J.; STREBEL, O.; VOERKELIUS, S.; SCHMIDT, H. L., 1990. Using Isotope Fractionation of Nitrate-Nitrogen and Nitrate-Oxygen for Evaluation of Microbial Denitrification in a Sandy Aquifer. *J. Hydrol.*, Vol. 114, pp. 413–24.

BOROWCZYK, M.; GRAUCZAK, J.; ZUBER, A., 1965. Radioisotope Measurements of the Groundwater Flow Direction by Single Well Method. *Nukleonika*, Vol. 10, pp. 19–28.

CLARK, I.; FRITZ, P., 1997. *Environmental Isotopes in Hydrogeology*. Lewis Publishers, Boca Raton, New York, 328 pp.

COLLON, P.; KUTSCHERA, W.; LEHMANN, B. E.; LOOSLI, H. H.; PURTSCHERT, R.; LOVE, A.; SAMPSON, L.; ANTHONY, D.; COLE, D.; DAVIDS, B.; FAUERBACH, M.; HARKEWICZ, R.; MORRISSEY, D. J.; SHERRILL, B. M.; STEINER, M.; PARDO, R. C.; PAUL, M., 1999. Development of Accelerator Mass Spectrometry (AMS) for the Detection of ^{81}Kr and First Application to Groundwater Dating. In: *Isotope Techniques in Water Resources Development and Management*, International Atomic Energy Agency, Vienna, CD-ROM.

COPLEN, T. B., 1993. Uses of Environmental Isotopes. In: Alley, W. D. (ed.), *Regional Groundwater Quality*. Van Nostrand Reinhold, New York, pp. 227–54.

CRAIG, H., 1961a. Standard for Reporting Concentrations of Deuterium and Oxygen-18 in Natural Waters. *Science*, Vol. 133, pp. 1833–4.

——, 1961b. Isotopic Variations in Meteoric Waters. *Science*, Vol. 133, p. 1702.

CRAIG, H.; GORDON, L. I., 1965. Deuterium and Oxygen-18 Variations in the Ocean and the Marine Atmosphere. In: *Stable Isotopes in Oceanographic Studies and Paleotemperatures*, CNR-Laboratorio di Geologica Nucleare, Pisa, pp. 9–130.

CRAIG, H.; LAL, D., 1961. The Production of Natural Tritium. *Tellus*, Vol. 13, pp. 85–105.

DANSGAARD, W., 1954. The ^{18}O Abundance in Fresh Water. *Geochim. Cosmochim. Acta*, Vol. 6, pp. 241–60.

DANSGAARD, W., 1964. Stable Isotopes in Precipitation. *Tellus,* Vol. 16, pp. 436–68.

DAVIS, S. N.; BENTLEY, H. W., 1982. Dating Groundwater; A Short Review. In: L. A. Currie (ed.), *Nuclear and Chemical Dating Techniques – Interpreting the Environment Record*, American Chemical Society, Washington, D.C., pp. 187–222.

DATTA, P. S.; GUPTA, S. K. ; SHARMA, S. C., 1978. Simulation of Water Movement through Vadose Zone. In: S. K. Gupta and P. Sharma (eds.), *Current Trends in Arid Zone Hydrology*. Today and Tomorrow Printer and Publishers, New Delhi, p. 93.

DILLON, M. A.; AGGARWAL, P. K., 1999. A Combined Vadose Zone – Saturated Zone Model for Tritium and Helium Transport in Groundwater. In: *Isotope Techniques in Water Resources Development and Management*, International Atomic Energy Agency, Vienna, Paper SM-361/32, CD-ROM.

DINÇER, T., 1968. The Use of Oxygen-18 and Deuterium Concentrations in the Water Balance of Lakes. *Water Resour. Res.*, Vol. 4, pp. 1289–306.

DROST, J. W.; NEUMAIER, F., 1974. Application of Single Borehole Methods in Groundwater Research. In: *Isotope Techniques in Groundwater Hydrology,* International Atomic Energy Agency, Vienna, pp. 241–53.

DROST, J. W., 1989. *Single-Well and Multi-Well Nuclear Tracer Techniques – A Critical Review.*

Technical Documents in Hydrology, International Hydrological Programme, UNESCO, Paris, 96 pp.

EDMONDS, W. M.; DROUBI, A., 1998. Groundwater Salinity and Environmental Change. In: *Isotope Techniques in The Study of Environmental Change*, International Atomic Energy Agency, Vienna, pp. 503–18.

EGGENKAMP, H. G. M., 1994. D^{37}Cl – The Geochemistry of Chlorine Isotopes. *Geologica Ultraiectina* No. 116, Thesis, Univ. Utrecht, 151 pp.

EHHALT, D.; KNOTT, K.; NAGEL, J. F.; VOGEL, J. C., 1963. Deuterium and Oxygen-18 in Rain Water. *J. Geophys. Res.*, Vol. 68, pp. 3775–80.

ELMORE, D.; PHILLIPS, F. M., 1987. Accelerator Mass Spectrometry for Measurement of Long-Lived Radioisotopes. *Science*, Vol. 236, pp. 543–50.

FLORA, O.; LONGINELLI, A., 1989. Stable Isotope Hydrology of a Classical Karst Area, Trieste, Italy. In: *Isotope Techniques in The Study of The Hydrology of Fractured and Fissured Rocks*. International Atomic Energy Agency, Vienna, pp. 203–14.

FONTES, J. CH.; GONFIANTINI, R., 1967. Comportment isotopique au cours de l'évaporation de deux bassins Sahariens. *Earth Planet. Sci. Lett.*, Vol. 3, pp. 258–66, 386.

FONTES, J. CH., 1983. Dating of Groundwater. In: *Guidebook on Nuclear Techniques in Hydrology* (1983 Edition), International Atomic Energy Agency, Vienna, pp. 284–317.

FONTES, J. CH.; ANDREWS, J. N.; EDMUNDS, W. M.; GUERRE, A.; TRAVI, Y., 1991. Palaeorecharge by the Niger River (Mali) Deduced from Groundwater Geochemistry. *Water Resour. Res.*, Vol. 27, pp. 199–214.

FRIEDMAN, I.; REDFIELD, A. C.; SCHOEN, B.; HARRIS, J., 1964. The Variation of the Deuterium Content of Natural Waters in the Hydrologic Cycle. *Rev. Geophysics*, Vol. 2, pp. 177–224.

FRITZ, P.; FONTES, J. CH. (eds.), 1980. *Handbook of Environmental Isotope Geochemistry*, *The Terrestrial Environment–A*, Vol. 1. Elsevier, Amsterdam, 545 pp.

FROEHLICH, K.; GELLERMANN, R., 1987. On the Potential Use of Uranium Isotopes for Groundwater Dating. *Chem. Geol. (Isotope Geoscience Section* and*)*, Vol. 65, pp. 67–77.

FROEHLICH, K.; IVANOVICH, M.; HENDRY, M. J.; ANDREWS, J. N.; DAVIS, S. N.; DRIMMIE, R. J.; FABRYKA-MARTIN, J.; FLORKOWSKI, T.; FRITZ, P.; LEHMANN, B. E.; LOOSLI, H. H.; NOLTE, E., 1991. Application of Isotope Methods to Dating of Very Old Groundwaters : Milk River Aquifer, Alberta, Canada. *Appl. Geochemistry*, Vol. 6, pp. 465–72.

GAT, J. R., 1971. Comments on the Stable Isotope Method in Regional Groundwater Investigations. *Water Resourc. Res.*, Vol. 7, pp. 980–3.

GEYH, M. A., 2000. An Overview of ^{14}C Analysis in the Study of Groundwater. *Radiocarbon*, Vol. 42, pp. 99–114.

GODWIN, H., 1962. Half-Life of Radiocarbon. *Nature*, Vol. 195, pp. 984–5.

GONFIANTINI, R.; CONRAD, G.; FONTES, J. CH.; SAUZAY, G.; PAYNE, B. R., 1974a. Etude isotopique de la nappe continentale intercalaire et de ses relations avec les autres nappes du Sahara septentrional. In: *Isotope Techniques in Groundwater Hydrology*, Vol. I, International Atomic Energy Agency, Vienna, pp. 227–41.

GONFIANTINI, R.; DINÇER, T.; DEREKOY, A. M., 1974b. Environmental Isotope Hydrology in the Hodna Region, Algeria. In: *Isotope Techniques in Groundwater Hydrology*, Vol. I, International Atomic Energy Agency, Vienna, pp. 293–16.

GUIZERIX, J.; GRANDCLEMENT, G.; GAILLARD, B.; RUBY, P., 1963. Appareil pour la mesure des vitesses relatives des eaux souterraines par la méthode de dilution ponctuelle. In: *Radioisotopes in Hydrology*, International Atomic Energy Agency, Vienna, pp. 25–35.

HALEVY, E.; NIR, A., 1962. The Determination of Aquifer Parameters with the Aid of Radioactive Tracers. *J. Geophys. Res.*, Vol. 61, pp. 2403–9.

HALEVY, E.; MOSER, H.; ZELLHOFER, O.; ZUBER, A., 1967. Borehole Dilution Techniques: A Critical Review. In: *Isotopes in Hydrology*, International Atomic Energy Agency, Vienna, pp. 531–64.

HANTUSH, M. S., 1963. Hydraulic of Wells. In: V. T Chow (rd.), *Advances in Hydroscience*, Vol. 1, pp. 281–432.

HO, D. H.; ARANYOSSY, J. F.; LOUVAT, D.; HUA, M. Q.; NGUYEN, T. V.; NGUYEN, K. C., 1991. Environmental Isotope Study Related to the Origin, Salinisation and Movement of Groundwater in the Mekong Delta (Vietnam). In: *Isotope Techniques in Water Resources Development 1991*, International Atomic Energy Agency, Vienna, pp. 415–28.

INTERNATIONAL ATOMIC ENERGY AGENCY, 1983. *Isotope Techniques in The Hydrogeological Assessment of Potential Sites for The Disposal of High-Level Radioactive Wastes*. Technical Report Series No. 228, International Atomic Energy Agency, Vienna, 151 pp.

INTERNATIONAL ATOMIC ENERGY AGENCY, 1989. *Guide to Safe Handling of Radioisotopes in Hydrology*. IAEA Safety Series.

IVANOVICH, M.; FROEHLICH, K.; HENDRY, M. J., 1991. Uranium-Series Radionuclides in Fluids and Solids, Milk River Aquifer, Alberta, Canada. *Appl. Geochemistry*, Vol. 6, pp. 405–18.

IVANOVICH, M.; HARMON, R. S. (eds.), 1992. *Uranium-Series Disequilibrium: Applications to Earth, Marine, and Environmental Sciences*. 2nd ed., Clarendon Press, Oxford, 907 pp.

JORDAN, H.; FROEHLICH, K., 1990. Groundwater Dating in Monitoring Groundwater Resources. In: *Groundwater Monitoring and Management*, IAHS Publ. No. 173, pp. 139–51.

KAUFMAN, W. J.; ORLOB, G. T., 1956. Measuring Groundwater Movement with Radioactive and Chemical Tracers. *J. Am Water Works Ass.*, Vol. 48, pp. 559–72.

KELLEY, C. A.; HAMMER, B. T.; COFFIN, R. B., 1997. Concentrations and Stable Isotope Values of BTEX in Gasoline-Contaminated Groundwater. *Environ. Sci. Technol.*, Vol. 31, pp. 2469–72.

KENDALL, C., 1998. Tracing Nitrogen Sources and Cycles in Catchments. In: *Isotope Tracers in Catchment Hydrology*. Elsevier, Amsterdam, pp. 519–76.

KNUTSSON, G.; FORSBERG, H. E., 1967. Laboratory Evaluation of ^{51}Cr-EDTA as a Tracer for Groundwater Flow. In: *Isotopes in Hydrology*, International Atomic Energy Agency, Vienna, pp. 629–52.

KROUSE, H. R., 1980. Sulphur Isotopes in Our Environment. In: P. Fritz and J. Ch. Fontes (eds.), *Handbook of Environmental Isotope Geochemistry*, *The Terrestrial Environment–A*, Vol. 1. Elsevier, Amsterdam, pp. 435–71.

KULKARNI, K. M.; KULKARNI, U. P.; SHARMA, S., 1999. Application of Environmental Stable Isotopes in Mine Hydrology: A Field Example. In: *Proc. 8th ISMAS Symposium on Mass Spectrometry*, Vol. II, Indian Society for Mass Spectrometry, Mumbai, pp. 652–5.

KULKARNI, K. M.; NAVADA, S. V.; NAIR, A. R.; RAO, S. M.; SHIVANNA, K.; SINHA, U. K.; SHARMA, S., 1998. Drinking Water Salinity Problem in Coastal Orissa, India – Identification of Past Transgressions of Sea by Isotope Investigation. In: *Isotope Techniques in The Study of Environmental Change,* International Atomic Energy Agency, Vienna, pp. 293–306.

KULKARNI, K. M.; NAVADA, S. V.; RAO, S. M.; NAIR, A. R.; KULKARNI, U. P.; SHARMA, S., 1996. Effect of the Holocene Climate on the Composition of Groundwater in Parts of Haryana, India – Isotopic Evidence. In: *Isotopes in Water Resources Management,* Vol. II, International Atomic Energy Agency, Vienna, pp. 439–54.

LAL, D.; PATERS, B., 1962. Cosmic Ray Produced Isotopes and Their Applications in Problems of Geophysics. *Progr. Elementary Particle Cosmic Ray Phys.*, Vol. 6, pp. 1–74.

LENDA, A.; ZUBER, A., 1970. Tracer Dispersion in Groundwater Experiments. In: *Isotope Hydrology 1970*, International Atomic Energy Agency, Vienna, pp. 619–41.

LEVIN, I., 1978. Regionale modellierung des atmosphärischen CO_2 auf grund von C-13 and C-14 messungen, Diplomarbeit, Institut für Umweltphysik, Univ. Heidelberg.

LIBBY, W. F., 1965. *Radiocarbon Dating*. University of Chicago Press, Chicago, Illinois, 175 pp.

LIU, B.; PHILLIPS, F. M.; HEINES, S.; CAMPBELL, A. R.; SHARMA, P., 1995. Water Movement in Desert Soil Traced by Hydrogen and oxygen Isotopes, Chloride and Chlorine-36, Southern Arizona. *J. Hydrol.*, Vol. 168, pp. 91–110.

LOVE, A. J.; HERCZEG,A. L.; LEANEY, F. W.; STADTER, M. F.; DIGHTON, J. C.; ARMSTRONG, D., 1994. Groundwater Residence Time and Palaeohydrology in the Otway Basin, South Australia: ^{2}H, ^{18}O, and ^{14}C Data. *J. Hydrol.*, Vol. 153, pp. 411–42.

LOVE, A. J.; HERCZEG, A. L.; SAMPSON, R. G.; CRESSWELL, R. G.; FIFIELD, L. K., 2000. Sources of Chloride and Implications for ^{36}Cl Dating of Old Groundwaters, Southwestern Great Artesian Basin, Australia. *Water Resour. Res.*, Vol. 36, pp. 1561–74.

MAIRHOFER, H., 1963. Bestimmung der Strömungsrichtung des Grundwassers in einem einzigen Bohrloch mit Hilfe radioaktiver Elemente. *Atompraxis*, Vol. 9, pp. 2–4.

MALOSZEWSKI, P.; RAUERT, W.; STICHLER, W.; HERMANN, A., 1983. Application of Flow Models in an Alpine Catchment Area Using Tritium and Deuterium Data. *J. Hydrol.*, Vol. 66, pp. 319–30.

MANDEL, S.; BUGAYEVSKY, M.; BACHMAT, Y.; DROST, W., 1985. Estimation of Dispersivity and Effective Porosity by a Single Well Tracer Technique. In: *Scientific Basis for Water Resources Management*, IAHS, Wallingford, pp. 219–29.

MARGRITA, R; GAILLARD, B., 1991. Use of Artificial Tracers for the Determination of Aquifer Parameters. In: *Use of Artificial Tracers in Hydrology*, IAEA-TECDOC-601, International Atomic Energy Agency, Vienna, pp. 131–42.

MERCADO, A.; HALEVY, E., 1966. Determining the Average Porosity and Permeability of a Stratified Aquifer with the Aid of Radioactive Tracers. *Water Resour. Res.*, Vol. 2, pp. 525–31.

MOSER H.; NEUMAIER, F.; RAUERT, W., 1957. Die Anwendung radioaktiver isotope in der Hydrologie. *Atomkernenergie*, Vol. 2, pp. 225–31.

MOSER H.; NEUMAIER, F.; RAUERT, W., 1963. New Experience with the Use of Radioactive Isotopes in Hydrology. In: *Radioisotopes in Hydrology*, International Atomic Energy Agency, Vienna, pp. 283–95.

MOSER, H.; STICHLER, W., 1970. Deuterium Measurements on Snow Samples from the Alps. In: *Isotope Hydrology 1970*, International Atomic Energy Agency, Vienna, pp. 43–57.

MURPHY, E. M.; DAVIS, S. N.; LONG, A.; DONAHUE, D.; JULL, A. J. T., 1989. C-14 in Fractions of Dissolved Organic Carbon in Groundwater. *Nature*, Vol. 337, pp. 153–5.

MUSGROVE, M. L.; BANNER, J. L., 1993. Regional Groundwater Mixing and the Origin of Saline Fluids: Midcontinent, United States. *Science*, Vol. 259, pp. 1877–82.

NAIR, A.; PENDHARKAR, R.; NAVADA, S. V.; RAO, S. M., 1979. Groundwater Recharge Studies in Maharashtra – Development of Isotope Techniques and Field Experience. In: *Isotope Hydrology 1978*, Vol. II, International Atomic Energy Agency, Vienna, pp. 803–26.

NICHOLLS, P. H.; BROMILOW, R. H.; ADDISCOTT, T. M., 1982. Measured and Simulated Behaviour of Fluometuron, Aldoxycarb and Chloride ion in a Fallow Structured Soil. *Pesti. Sci.*, Vol. 13, pp. 475–83.

OLSSON, I. U., 1968. Modern Aspects of Radiocarbon Dating. *Earth Sci. Rev.*, Vol. 4, pp. 203–18.

PAYNE, B. R.; QUIJANO, L.; LATTORE, C., 1980. A Case Study of Leakage between Aquifers in Hermosillo, Mexico, Using Environmental Isotopes. In: *Arid Zone Hydrology: Investigations with Isotope Techniques*, International Atomic Energy Agency, Vienna, pp. 113–30.

PEARSON, F. J., JR; RIGHTMIRE, C. T., 1980. Sulphur and Oxygen Isotopes in Aqueous Sulphur Compounds. In: P. Fritz and J. Ch. Fontes, (Eds.), *Handbook of Environmental Isotope Geochemistry*, *The Terrestrial Environment–A*, Vol. 1, , Elsevier, Amsterdam, pp. 227–58.

PICCIOTTO, E., DE MAERE, X. ; FRIEDMAN, I., 1960. Isotopic Composition and Temperature of Formation of Antarctic Snows. *Nature*, Vol. 187, pp. 857–9.

PLATA BEDMAR, A., 1981. Empleo de técnicas isotópicas en el estudio de la problemática de la futura presa de San Marcos, Proyecto Tabacundo (Ecuador). Informe GANOP I-1A2/81, CEDEX, Ministerio de Obras Públicas, Madrid.

——, 1983. Single Well Techniques Using Radioactive Tracers. In: *Tracer Methods in Isotope Hydrology*, International Atomic Energy Agency, Vienna, pp. 17–46.

PLUMMER, N. L.; MICHEL, R. L.; THURMAN, E. M.; GLYNN, P. E., 1993. Environmental Tracers for

Age Dating Young Groundwater. In: W. D. Alley (ed.) *Regional Groundwater Quality*, Van Nostrand Reinhold, New York, pp. 255–94.

PLUMMER, N. L.; PRESTEMON, E. C.; PARKHURST, D. L., 1994. An Interactive Code (NETPATH) for Modelling Net Geochemical Reactions Along a Flow Path, Version 2.0. *U.S. Geological Survey Water Resources Investigations Report 94 – 4169*, U.S.G.S., Reston, VA.

RAMSPACHER, P.; ZOJER, P. H.; FROEHLICH, K.; STICHLER, W., 1992. The Recharge of Large Springs from a Carbonate Aquifer Near Innsbruck Applying Environmental Tracers. In: H. Hoetzl and A. Werner (eds.), *Tracer Hydrology*. A. A. Balkema, Rotterdam, pp. 251–7.

ROBERTSON, W. D.; CHERRY, J. A., 1989. Tritium as an Indicator of Recharge and Dispersion in a Groundwater System in Central Ontario. *Water Resour. Res.*, Vol. 25, pp. 1097–109.

RAO, S. M.; KULKARNI, K. M., 1997. Isotope Hydrology Studies on Water Resources in Western Rajasthan. *Current Science*, Vol. 72, pp. 55–61.

ROZANSKI, K.; ARAGUÁS, L.; GONFIANTINI, R., 1993. Isotopic Patterns in Modern Global Precipitation. In: *Climate Change in Continental Isotopic Records*, American Geophysical Union, Washington DC, pp. 1–36.

SCHLOSSER, P.; STUTE, M. C.; SONNTAG, C.; MÜNNICH, K. O., 1989. Tritiogenic ^{3}He in Shallow Groundwater. *Earth Planet Sci. Lett.*, Vol. 94, pp. 245–56.

SIEGEL, D. I., 1990. Sulphur Isotope Evidence for Regional Recharge of Saline Water During Continental Glaciation, North-Central United States. *Geology*, Vol. 18, pp. 1054–6.

——, 1991. Evidence for Dilution of Deep, Confined Groundwater by Vertical Recharge of Isotopically Heavy Pleistocene Water. *Geology*, Vol. 19, pp. 433–6.

STUTE, M.; SCHLOSSER, P.; CLARK, J. F.; BROECKER, W. S., 1992. Paleotemperatures in the Southwestern United States Derived from Noble Gas Measurements in Groundwater. *Science*, Vol. 256, pp. 1000–3.

STUTE, M.; TALMA, A. S., 1998. Glacial Temperatures and Moisture Transport Regimes Reconstructed from Noble Gases and d^{18}O, Stampriet Aquifer, Namibia. In: *Isotope Techniques in the Study of Past and Current Environmental Changes in the Hydrosphere and Atmosphere*, International Atomic Energy Agency, Vienna, pp. 307–18.

SZABO, Z. D.; RICE, E.; PLUMMER, L. N.; BUSENBERG, E.; DRENKARD, S.; SCHLOSSER, P., 1996. Age Dating of Shallow Groundwater with Chlorofluorocarbons, Tritium/Helium-3, and Flow Path Analysis, Southern New Jersey Coastal Plain. *Water Resour. Res.*, Vol. 32, pp. 1023–38.

TANDIA, A. A.; ARANYOSSY, J.-F.; GAYE, C. B., 1993. Evaluation de la recharge des nappes phréatiques sous climat semi-aride. Exemple du Nord-Sénégal. In: *Les ressources en eau au Sahel*, IAEA-TECDOC-721, International Atomic Energy Agency, Vienna, pp. 39–51.

TAYLOR, C. B.; ROETHER, W., 1982. A Uniform Scale for Reporting Low-Level Tritium Measurements in Water. *Int. J. Radiat. Isotop.*, Vol. 33, pp. 377–82.

TAZIOLI, G. S., 1977. Groundwater Circulation in the Coastal Carbonate Aquifers of Apulia. In: *Hydrodynamic Diffusion and Dispersion in Porous Media*, IAHR, Pavia, pp. 187–99.

VRBKA, P.; JACOB, H.; FROEHLICH, K.; SALIH, M. K., 1993. Identification of Groundwater Recharge Sources in Northern Sudan by Environmental Isotopes, *J. Environment. Hydrol.*, Vol. 1, pp. 8–16.

TORGERSEN, T.; HABERMEHL, M. A.; PHILLIPS, F. M.; ELMORE, D.; KUBIK, P.; JONES, B. G.; HEMMICK, T.; GROVE, H. E., 1991. Chlorine-36 Dating of Very Old Groundwater, III. Further Studies in the Great Artesian Basin, Australia. *Water Resour. Res.*, Vol. 27, pp. 3201–14.

WASSENAAR, L. I.; ARAVENA, R.; FRITZ, P., 1992. Radiocarbon Contents of Dissolved Organic and Inorganic Carbon in Shallow Groundwater Systems: Implications for Groundwater Dating. In: *Isotope Techniques in Water Resources Development 1991*, International Atomic Energy Agency, Vienna, pp. 143–51.

WEINBERGER, W. A.; ELLIS, W. R.; SEATONBERG, B. W.; ANDREW, J. T. G., 1967. Radioisotopes as Groundwater Tracers. *J. Geophys. Res.*, Vol. 72, pp. 4081–91.

WOODCOCK, A. H.; FRIEDMAN, I., 1963. The Deuterium Content of Raindrops. *J. Geophys. Res.*, Vol. 68, pp. 4477–83.

WURZEL, P.; WARD, P. R. B., 1965. A Simplified Method of Groundwater Direction Measurement in a Single Borehole. *J. Hydrol.*, Vol. 3, pp. 97–105.

ZAK I.; GAT, J. R., 1975. Saline Waters and Residual Brines in the Shiraz-Sarvistan Basin, Iran. *Chem. Geol.*, Vol. 16, p. 179.

ZIMMERMANN, U.; MUNNICH, K. O.; ROETHER, W.; KREUTZ, W.; SCHAUBACH, K.; SIEGEL, O., 1966. Tracers Determine Movement of Soil Moisture and Evapotranspiration. *Science*, Vol. 152, pp. 346–7.

ZIMMERMANN, U.; MUNNICH, K. O.; ROETHER, W., 1967. Downward Movement of Soil Moisture Traced by Means of Hydrogen Isotopes. *Am. Geophys. Union Monogr.*, No. 11, pp. 28–36.

ZUBER, A., 1986. Mathematical Models for the Interpretation of Environmental Radioisotopes in Groundwater Systems. In: P. Fritz and J. Ch. Fontes (Eds.), *Handbook of Environmental Isotope Geochemistry, The Terrestrial Environment - B*, Vol. 2. Elsevier, Amsterdam, pp. 1–59.

10 Hydrogeological mapping

10.1 Introduction

The hydrogeological map has three main functions:

1. It is a tool for visualising and understanding the regional hydrogeological conditions observed or deduced from various investigations and analyses. Insight into these conditions is not obtained as instantly as insight into the conditions controlling surface water. It requires the combination of a great amount of widely dispersed data of various kinds, whose analysis is largely a matter of interpretation. Therefore mapping is an important tool for coherently arranging coherently the available spatial information; furthermore, it is an essential means of increasing our insight.
2. It is a means of communication between hydrogeologists, in the way the geological map is for geologists. It is an indispensable supplement to monographs and, long before the era of computerised 'information systems', it was already a veritable pictorial 'database' that was immediately and easily accessible. It also became a valued teaching aid.
3. It is a means of communication between the hydrogeologist and various kinds of non-specialists who require information about groundwater and its users.

Consequently, hydrogeological mapping serves two purposes: scientific and utilitarian.

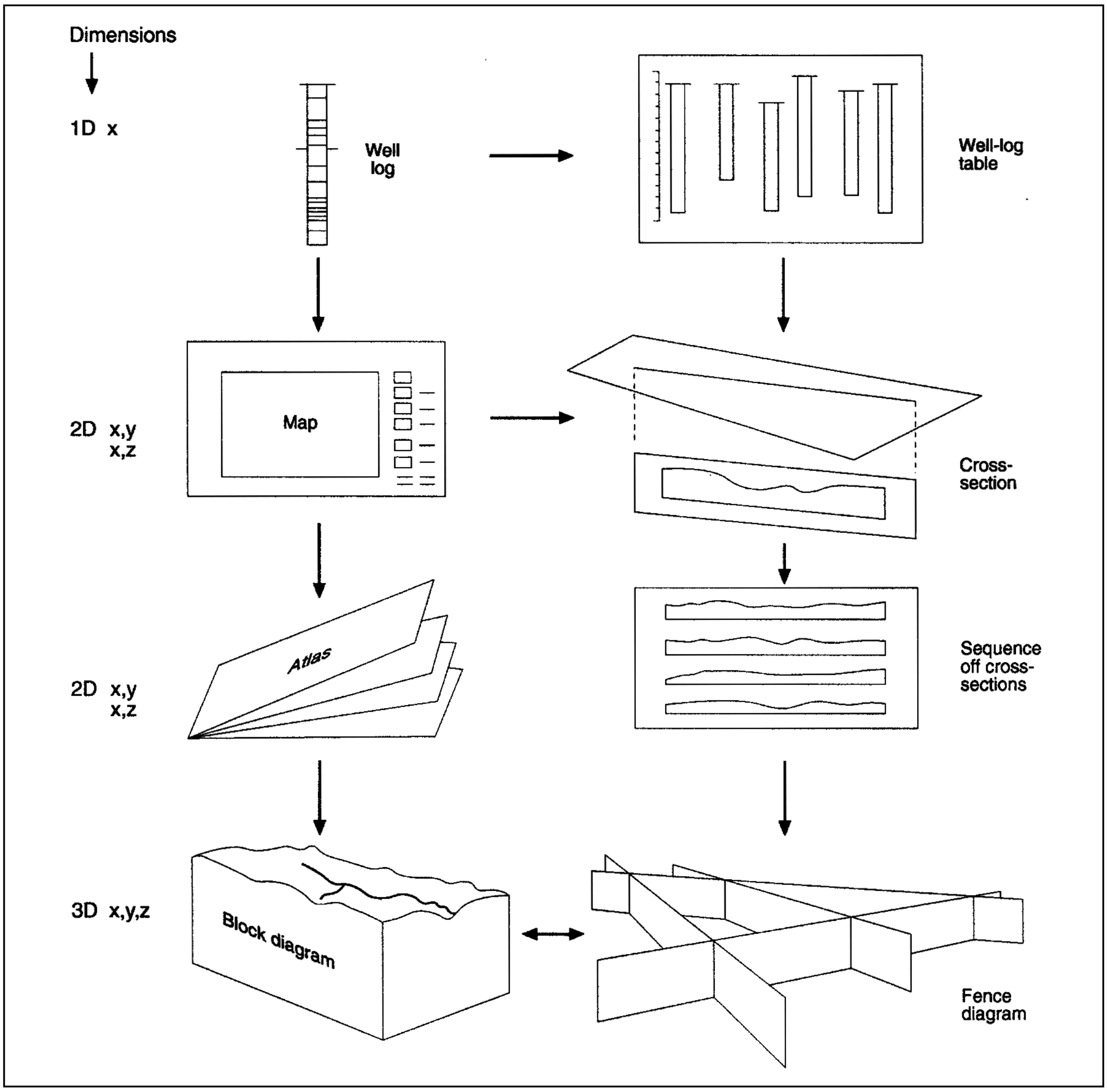
10.2 The role and place of hydrogeological mapping

10.2.1 Maps among the methods of graphic representation

Hydrogeological maps have a central position among the methods of graphic presentation. They depict only what can be presented in two horizontal dimensions in a given territory (Fig. 10.1).

By definition a map depicts space not time. However, it is not devoid of temporal references: it represents the state of knowledge at a given moment, or data on variables over time. The map may depict a situation at a given date, in principle the same for the whole map if the variable is spatio-temporal (for example, piezometric levels within a single body of groundwater). Alternatively it may depict several situations at different dates, so they can be compared; or it may depict a mean situation for a specified reference period. Sometimes a cartogram can be produced by superimposing on the map the graphs depicting a given variable at various observation points.

Figure 10.1 Spatial graphics



10.2.2 Maps among other methods of storing and representing information

In presenting data or representing more complex and synthetic information, maps are closely linked to other tools. They supplement data banks or databases, arrays of information and simulation models (Fig. 10.2).

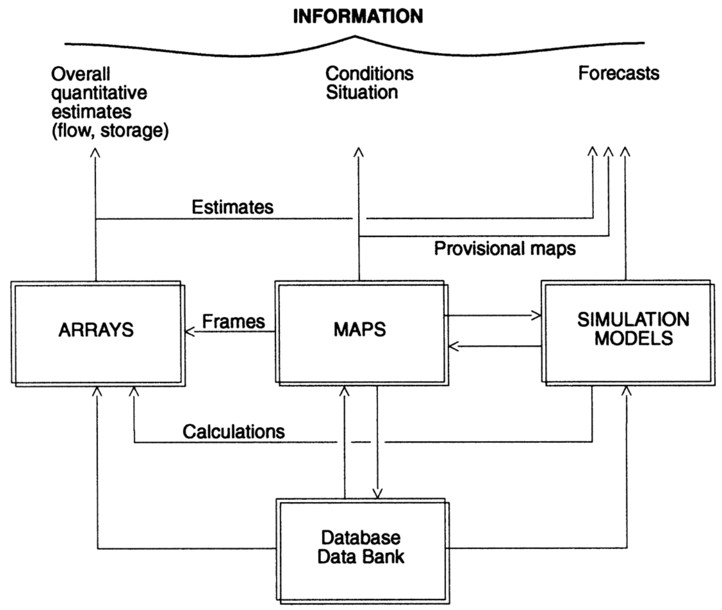
(a) *Relation to data banks*

The analytical map that shows point or line data precedes the data bank by delivering entries to it, e.g. coordinates, digitisation, etc.; and follows it by presenting data bank output, e.g. maps of selected points, and isolines of regional variables drawn by appropriate interpolation software.

(b) *Relation to arrays of information*

A map is especially useful for the analytical presentation of the water budget of an aquifer system: notably to present the reference area, or its subdivisions, and the flow conditions at the various boundaries.

Figure 10.2 Relation between the tools and the modes of expression



(c) *Relation to hydrodynamic simulation model*
The input data for a model are mainly prepared as parametric maps that are derived from data banks, and maps showing the delimitation of areas satisfying specific conditions that may be explicit transformations of classic hydrogeological maps. The number of maps increases according to the structural complexity of the studied aquifer system and the complexity of the simulated phenomena. Ultimately, the output of the model will also be expressed as maps.

10.2.3 Role of mapping in the development of hydrogeological insight into an area

Hydrogeological mapping plays different roles at two distinct stages in the accumulation of knowledge:

- At the stage of preliminary reconnaissance the collection of the data for the hydrogeological map is part of what is usually described as the 'inventory' work. The requirements for the preparation of each map sheet dictates the framework and the work programme of the fieldwork including field mapping and underground investigations. The results are presented mainly (but not exclusively) as either reconnaissance maps or as detailed 'regular' maps that systematically cover the area.
- At a more advanced stage, the hydrogeological maps are drawn using also the results of local and regional studies, wellfield data and results from underground engineering work, rather than data from systematic prospecting. It is a matter of synthesis and integration of scattered data. Therefore the hydrogeological maps are more a synthesis of the recorded data than of a systematic coverage. They may be a part of national or international syntheses that will serve scientific or didactic purposes, or they will present a particular macro-economic orientation.

10.3 What can and must hydrogeological maps depict?

Hydrogeological maps show the data and information that can be rendered on a map. The subjects of hydrogeological mapping are many and varied, and a list of them reads like the 'menu' of a complex theoretical cartographic legend.

10.3.1 Inventory

(i) First bifurcation, reflecting the subject and scope of hydrogeology, distinguishes two subjects, both of which can be described and depicted cartographically:

(a) Rocks:
- the nature, geometry and structure of geological formations, seen from the viewpoint of their role *vis-à-vis* the presence and circulation of water in relation to its properties, e.g. groundwater reservoirs or aquicludes,
- formations that to some extent transmit or store water,
- formations in contact with or isolated from the surface conditions.

The hydrogeological characteristics can be presented in different ways, e.g. on a lithological scale portraying the nature of the rock, or on a geological scale with emphasis on the regional structures. Moreover they can be presented qualitatively (types) or quantitatively (parameters).

(b) Groundwater:
- its presence, extent, and abundance,
- its hydrodynamic character (hydraulic head or potential; flow direction; flow velocity), its chemical and thermal characteristics, and
- its hydraulic connection with surface water.

These characteristics can be described quantitatively or on various scales.

(ii) Artefacts

It is usually desirable to include on the map structures and sites that have a relationship to naturally occurring groundwater or water in general. This may include:
- ad hoc investigation sites (e.g. test holes),
- permanent observation structures (piezometers and hydrometric stations),
- other data-gathering stations,
- exploitation structures (e.g., wells, boreholes, springs, injection sites, underground engineering works) and
- any sites of surface activity likely to influence the flow regime or the quality of the groundwater (irrigation, slurry-spreading, drainage, etc.).

(iii) Typology and classification

Many of these items are characterised by qualitative, semi-quantitative (ordinal) or quantitative typologies or classifications:
- water-bearing/non-water-bearing rocks, or categories of permeability and transmissivity;
- porous or discontinuous formations (fissured, fractured, karstic);
- different conditions at the boundaries of groundwater systems, or involving the relationships between groundwater and surface water;

- types of springs or categories of average rate of flow; hydrochemical or water-quality categories;
- types of construction.

(iv) Data and information

A whole range of information may be presented on maps: from directly observed relatively simple field data to highly complex, processed and synthesised information geared to foreseeable questions. It is not always possible to show all these on the same map, however (see Table 10.1).

Table 10.1 Mappable subjects, classified by nature and degree of conceptualisation

Subjects. Degree of conceptualisation	*Hydrogeological formations and structures*	*Groundwater*	*Artefacts*
Basic factual data (measurements, observations)	Kind of outcropping formations. Hydrogeological contour. Local depth of reservoir. Upper or lower confining bed. Local thickness of a formation. Type of overlying formation	Location of observation stations (x, y, z). Depth of water table	Location of well, borehole, drain. Depth of structure
Results of simple data processing (simple calculations and interpolation)	Masked hydrogeological contours. Local elevation of a reservoir's upper or lower confining bed. Isohypse of a formation's upper or lower boundary. Lines of equal thickness (isopachs)	Local hydraulic potential (dated, average). Variation of hydraulic head. Rate of flow (average). Lines of equal potential (piezometric surface)	
Results of complex data processing	Hydrodynamic parameters of the formation: permeability, porosity, transmissivity (zoning; surface conditions controlling infiltration (type and thickness of superficial unsaturated formations).Structural class (according to the conditions at depth)	Boundary of groundwater. Isobaths. Equal ionic or total salinity lines (isohalines). Boundary of unconfined/confined groundwater body. Site of potential artesian flow. Flow paths and mean local flow velocities. Groundwater flow divides. Degree of connection between groundwater body and river. Local fluxes or fluxes per inflow and outflow sector for a defined groundwater body	Specific. Discharge
Information	Accessibility. Probability of successful boring. Zones of maximum productivity (in a multi-layered system)	Water quality (normative), distribution. Vulnerability to pollution. Sensitivity to changes in surface conditions. Degree of transformation of present state as compared to a natural state (e.g. depression cones)	Probable productivity or injectivity. Withdrawal rate

10.3.2 Data of field surveys and subsurface investigations

Only part of the data needed for map-making can be collected by field observations, hydrogeological surveys (e.g. well inventory, groundwater levels, EC values, pumping data, spring inventory, surface water discharge) and mapping of contours, in combination with the interpretation of air photographs and remote sensing images. The additional information comes from the interpretation of underground reconnaissance data, e.g. structural and hydrodynamic data from drilling logs, data on the analysis of water and rock samples. The latter is usually mainly entails compiling data from reports and documents

10.3.3 Shapes and dimensions

Geometrical, mappable objects can be clustered into three formal groups distinguished essentially by the possibility of depicting them together on the same map:

- points that can be represented by symbols,
- linear (on the map) objects, and
- zonal, two-dimensional (at least on the map) objects (see Table 10.2).

Table 10.2 Clustering of mappable objects

Subjects *Dimensions*	*Formations and* *hydrogeological structures*	*Groundwater*	*Artefacts*
Points	Reference point. Location	Spring. Swallow hole. Reference point	Well. Borehole. Piezometer. Hydrometric station. Groundwater dam
Lines			
Actual	Hydrogeological contour. Line of tectonic disturbance, e.g. fracture, fault	Underground channel (karst). Watercourse; Littoral zone (as specified limit)	Drain. Gallery. Canal. Buried pipelines
Virtual	Contour lines of structural variables, e.g. isohypses, isobaths, isopachs. Boundaries, e.g. of a zone of the same class of transmissivity	Isopleths of hydrodynamic variables, e.g. equipotentials, lines of equal fluctuation, concentration and depth. Divide. Streamlines (vectors). Boundary of groundwater body/confinement	
Zones or regions	Zones belonging to a specific class, e.g. type of formation, transmissivity, deep structure, accessibility	Extent of defined ground-water body. Area of artesian flow. Area of outflow or seepage. Zones belonging to a specific class, e.g. average recharge, flow regime, hydrochemical characteristics, water quality	Irrigated or drained area. Intensive farmed area

While points and lines or families of (differentiated) lines can to a large extent coexist on the same map by virtue of the variety of possible modes of representation (shape, colour, etc.), it is much more difficult to superimpose zonal representations, hence the need in group 3 to choose a priority subject.

10.3.4 Mapping programme

The hydrogeological data and information that can be expressed on a map are not equally significant whatever the scale. The scale is therefore in itself a technical selection factor, but it may also be dictated by the choice of the principal subject to be mapped. Nevertheless, even at a given scale, the abundance of subjects may be such that juxtaposing them would result in a map so overloaded as to be unintelligible. A selection must therefore be made and the possibility and advantage explored of distributing the subjects over several maps; thus programme selectivity and a plurality of maps with distinct and complementary programs go together.

A mapping programme must try to reconcile:

- the essential data and information that must be presented,
- the groups of data or information that are complementary and mutually enlightening, in particular 'couplings' of data relating the terrain to the groundwater,
- consistency of the degree of abstraction: priority being given to objective descriptions with minimum interpretation or to the presentation of more complex and elaborate scientific or practical information,
- compatibility of presentations (linked to the dimensions of the subjects).

Finally, the need for a clear and understandable presentation limits the amount of information that can be shown.

10.3.5 Hydrogeological map sensu stricto

This should satisfy three principles:

(i) Scientific orientation ('cognitive')

The hydrogeological map must have a scientific content and therefore gives priority to an objective and explicit presentation of the hydrogeological conditions. This does not exclude the presentation of interpolations and structural hypotheses, Of secondary importance the presentation of derived information that the hydrogeologist can infer or that should be expressed on supplementary maps. The map therefore will show:

- 'hydrolithological' classes, associated with average estimates of permeability or transmissivity rather than with productivity figures;
- isohypses of significant surfaces such as the upper or lower confining bed of an aquifer rather than isobaths;
- the equipotential contours of the piezometric surface, rather than isobaths;
- classes of salinity of groundwater rather than of normative quality.

(ii) Balance in the presentation of the hydrogeological structures and the groundwater characteristics

The hydrogeological map must represent the data on the aquifers, including discontinuities and gaps, and the data on the dynamics and the physical or chemical characteristics of the groundwater they contain. They must explicitly render the relationships between 'content' and 'container' where those relationships are not obvious.

(iii) Balance between local characteristics and regional hydrogeological conditions

The hydrogeological map must represent both the spatial distribution of different continuous or discontinuous variables of local significance – such as the nature of surface formations or the potential of a groundwater body – and structural characteristics (contours and classes) giving visual expression to hydrogeological structures that are significant at the given scale. A hydrogeological map must in particular make it possible to identify the configuration and boundary conditions of the principal water-bearing systems of the area it covers.

10.3.6 Presentation of surface water

Three options may dictate the choice of the surface hydrological data that will be presented on a hydrogeological map:

(i) Minimum 'geographical' option

The hydrographic network is shown on the topographic base map for planimetric purposes only; it is presented in the traditional blue color. The hydrometric stations are shown as well, including relevant data about surface runoff (average rate of flow, areas of upstream basins, number of years of measurement, etc.).

(ii) Hydrological option

Watercourses are classified according to their average flow, deduced by interpolating the data from the hydrometric stations, or by the base flow that shows mainly shows the magnitude of the groundwater runoff collected by the watercourses.

(iii) Hydrogeological option

Presentation of watercourses according to the degree of hydraulic continuity with the underlying aquifers and the kind of interaction with the groundwater (whether draining, infiltrating, none at all, etc.), if desired, combined with quantitative data on the interactive flows. This option is suitable for the explicit mapping of aquifer systems and their boundary conditions. It requires more complex, less classical modes of representation that differentiate shapes and sometimes colours.

10.3.7 Specific purpose maps

Specific purpose groundwater maps are derived from the hydrogeological map and give priority, or are exclusively devoted to the information required to respond to the identified needs and questions of specific users (Table 10.3).

10.4 Classification of hydrogeological maps

The variety and abundance of mappable data means that they must be sorted and distributed among several maps. The difference between the purposes for which the maps are made offer criteria for the choice of a mapping programme. Moreover, the diversity of scales is at once a cause and an effect of their multiplicity. Programme, purpose and scale are thus the essential base of any classification of hydrogeological maps in the broad sense (Table 10.4).

Table 10.3 Users and information requirements

Users	*Useful mappable information*
Planners of groundwater-capture structures (wells and boreholes), drainage or underground engineering work. Users of groundwater	Presence of aquifer: depth and thickness of the aquifer, depth to water. Productivity: probable specific rate of flow, pumping level. Estimate of the zone of influence (calculation parameters). Probability of successful drilling. Water quality
Expert on protection of capture infrastructure	Conditions of recharge; extent of recharge area. Direction of surface runoff and groundwater flow. Proximity of active works upstream. Protective capacity of the soil
Land use impact expert	Water table elevation and depth below surface. direction of runoff. groundwater flow trajectory. Protective capacity of the soil cover (factors affecting vulnerability to pollution)
Resource assessors; Public administrators; Planners	Extent of groundwater reservoir and boundary conditions. Average inflow per unit of area; zones. Aquifer/stream connections. Distribution of parameters determining accessibility and probable exploitation costs. Groundwater quality zones

10.4.1 Programme and purpose

Programme and purpose obviously go together. Similarly, the communication purpose is inseparable from the users and their information needs. Those needs are explicitly expressed by the users or they are attributed to them; that is, the supply of information may anticipate unformulated needs or it may proceed from a desire to inform. The simplest grouping of the users (see also Table 10.3) is:

(a) specialists and professionals, e.g. hydrogeologists, hydraulic engineers and hydrologists, whose background is identical to that of those who designed and drew the map. They require a mapping programme that gives priority to descriptive and objective data, explicitly separating fact and interpretation;

(b) the users of groundwater and other underground resources and underground space. They include:
- the users of groundwater for domestic and industrial purposes,
- the users that may affect the groundwater by activities in the underground (borings, underground infrastructural works, mines, quarries, hydraulic installations, etc.),
- land owners undertaking construction work,
- users of land and water resources for agricultural purposes (effects on the depth of water, pollution risks),
- environmental managers and planners and various intermediaries between specialists and 'the public',

each with a need for a specific mapping programme that focuses on selected information as material for replies to specific questions.

Table 10.4 Classification of hydrogeological maps

Purpose	*Scale* — *Large*	*Scale* — *Medium and small*
Scientific	Hydrogeological map, detailed or 'regular'	Composite hydrogeological maps
	Hydrogeological reconnaissance map	Provisional composite hydrogeological maps
	Specialised hydrogeological maps: parametric maps (structural, hydrodynamic, piezometric, transmissivity, etc.)	Hydrogeological maps of basins and distinct groundwater formations. 'Geohydrodynamic' maps of groundwater systems. Hydrological parameter maps; maps of underground flow; maps of aquifer inflows. Maps of groundwater storage
Practical	Hydrogeochemical maps, general or specific	Composite hydrogeochemical maps. Maps of thermal or mineral waters
	Works productivity map. Maps of depth of groundwater and of upper confining bed. Groundwater production costs map	Classification by various criteria of groundwater resources. Map of groundwater thermal potential
	Specific quality map. Map of groundwater vulnerability to specific pollution	Map of general groundwater quality. Map of groundwater vulnerability to pollution in general

10.4.2 Scale

Although above all a matter of convention, the usual definitions are:
- large scale: 1:20,000 – 1:100,000;
- medium scale: 1:200,000 – 1:500,000;
- small scale: 1:1,000,000 and upwards.

10.4.3 Scientific maps and practical maps

The difference in purpose – and therefore of the programmes – has led to the design and production of two families of maps:

(1) scientific hydrogeological maps in the strict sense of the term, whose primary purpose is to represent natural conditions, and human activity and its effects in a given situation. They are preceded by reconnaissance maps, supplemented by more specific maps (parametric, in particular) and even reduced for atlases.

(2) practical maps, whose principal purpose is to inform and whose programme must consist of the information required to meet the assigned object. The utilitarian maps are the product of selective interpretations of the scientific maps. The hydrogeological map is the necessary basis of all practical maps, even though they may also use supplementary data.

The usual classification, presented in Table 10.4, shows these two families in relation to the usual scales. The differences in scale correspond to differences in 'levels of information', detail of interpolations and classifications, and reliability and precision of local information available from the map.

Table 10.5 reproduces Struckmeier´s (1989) classification of maps that is more explicit regarding levels of information and more 'technical' in its distinction of uses.

Table 10.5 System for classifying hydrogeological maps (After Struckmeier, 1989)

Level of information / *Possible use*	***Low*** *(scarce and heterogeneous data from various sources)*	***Advanced*** *(+ systematic investigation programmes, more reliable data)*	***High*** *(+ hydrogeological systems analysis and groundwater models)*
Reconnaissance and exploration	General hydrogeological map (aquifer map)	Hydrogeological parameter maps (map sets, atlases)	Regional groundwater systems maps (conceptual model representations)
Planning and development	Map of groundwater resource potential	**Specialised hydrogeological maps** (planning maps)	Graphic representation derived from Geographic Information Systems (maps, sections, block diagrams, scenarios)
Management and protection	Map of groundwater vulnerability		
Possible use / *Parameters of representation*	static ------------------------------------- time-dependence --------------------------→ dynamic low --- reliability ------------------------------------→ high low --------------------------------------- cost per unit area -------------------------------→ high large ←------------------------------------ area represented --------------------------------- small small --- scale ---------------------------------------→ large		

10.4.4 Terminology

The classification of maps according to different criteria is less an end in itself than a means of highlighting their differences and specialisations. It is less important to classify maps according to one system or another, than to give each type a clear name consistent with its content. Consequently, it seems desirable to retain the name *hydrogeological map (in a broad sense)* as a generic name, but to restrict the use of the name

- *hydrogeological map (sensu stricto)* to maps that express scientific knowledge relating to the geological structures that determine the occurrence and circulation of groundwater and to the groundwater itself, in other words maps that describe and explain static and dynamic hydrogeological conditions, and
- *groundwater resources maps* to utilitarian maps with a more specific and practical purpose.

10.5 Hydrogeological map making

10.5.1 The language of maps; properties and constraints

(i) Modes of expression

Six sorts of basic visual features may modify the signs that make up graphic language in general and map language in particular (Bertin 1973):

- size,
- value (from light to dark),
- shape,
- grain (screen, pattern and ornament),
- colour,
- orientation (of axially asymmetric signs).

These features may be combined in a single sign: one shape may have several sizes or orientations, a colour may have several values, etc. They are divided into two sorts:

- The first are called quantitative or ordered and may express either:
  (a) continuous gradations, of which *size* is the only absolute, direct mode of expression; or
  (b) quantitative or ordinal classifications, *i.e.* in a number of classes that are, however, limited (by convention, *value* and *size* again).
- The others, *shape*, *grain*, *colour* and *orientation*, are called qualitative, unordered or selective and can express differences of type or category; e.g. colour in different shades, and grain, with a large number of possible patterns. They can be continuous or not and orientated or not and can in addition be differentiated by colour. They offer the greatest range of options, particularly for filling in areas (zones).

(ii) Topology; features and dimensions

Map 'writing' uses three types of signs, each with a different dimension:

- *points* (dimensionless), located at x and y;
- solid or broken *lines* (one dimension), which can be indicated in digital shape by sets of points defined in terms of x and y;
- *zones* – expanses or areas – (two dimensions), indicated by lines or contours or fuzzy edges.

Five of the six visual features can modify the signs depicting these three types of features. Zones, however, cannot have size, shape or orientation.

(iii) Construction; assembly and superposition

The production of a thematic map is a work of assembly and involves matching the signs that express the data and information according to the adopted mapping programme. Therefore the compatibility and superimposability of signs are essential selection criteria when the key is prepared.

(iv) Degrees of perception

When the various visual features appear together they are not all perceived to the same degree. The order of perceptibility of the signs must therefore match the order of importance of the subjects, *i.e.* of the information to be presented.

- *Colour* is the most immediately perceptible feature. It must therefore describe the most important variables, especially for zones. In point or linear marking, only a few pure colours are clearly discernible.
- *Value* is, after colour, the most perceptible feature in zone marking, but it is barely perceptible in point or linear marking.
- *Size* is the most perceptible feature for points and lines.
- *Shape* is less perceived, and its perception is subordinated to size. The shape of a point sign is only noticed if the size exceeds 2 mm.
- *Grain* is perceived first by its value and only secondly by its shape.
- *Orientation* is the least immediately perceptible feature.

(v) Recommendations

- The range of a feature must express only one physical variable, qualitative or quantitative. Its gradation must be consistent, particularly if it is an ordered feature size, value, or scale of value of the represented variable (e.g. rate of flow, permeability.
- The qualitative, unordered features (colour, shape and grain) must express qualitative types and not quantitative classes. For example, classes of transmissivity or flow should not be represented by different colours or grain.
- It is recommended that superposition of the various signs and patterns is tested before deciding upon a key that presents them separately.

10.5.2 Key

A key or legend is a table that shows the correspondence between definitions of mappable objects (meanings) and the signs that represent them. Two general principles help to make this correspondence clearer:

- obvious dimensional analogies: point, and line data (contours), and areas of equal value relating to expanses are quite naturally represented by points, lines and zones, respectively;
- similar subdivisions between on the one hand, the material to be *shown* (e.g. areas of distribution of hydrogeological conditions or classes of defined variables; primordial boundaries), and *read* when consulting the map (information on various parameters or qualitative features); and on the other hand, between primarily *visible* (e.g. zonal markings) and primarily *legible* signs (points and lines).

(i) Rules

There are five rules to be observed when choosing a key:

- Consistency of the degree of perceptibility (visibility and legibility) with the order of importance of the presented data and information;
- Consistency of the features of the signs with the represented subjects: 'eloquent' correspondence facilitates immediate understanding. In particular, discontinuities in lines or 'grain' (patterns) best express actual discontinuities in space or time; binary oppositions (black/white, empty/full, positive/negative, etc.) are very appropriate for expressing the many conceptual dualities in hydrogeology (aquifer/non-aquifer, free/confined groundwater, fresh/salt water, etc.);
- Intelligibility: the map must be clear, particularly if it has been designed for non-specialist users. This calls for superimposability of signs and patterns;

- Conformity with recommended standards, particularly those that are international or tested by previous use;
- The standardisation of hydrogeological map keys and their derivatives should, however, allow for some flexibility and the possibility of change and enrichment deriving from increased knowledge and conceptual innovation;
- Printability: practical difficulties that could arise at the drawing and publication stages should be minimised.

(ii) Structure

The key to a hydrogeological map is generally structured in four sections devoted respectively to:

- hydrogeological formations and structures;
- groundwater;
- hydrography;
- artefacts (works, observation stations, and locations of significant human activity).

(iii) General keys

A general key is both a universal map programme, a 'menu' to be chosen from, and a dictionary of corresponding conventional signs. The first aim of a national or international general hydrogeological key is the standardisation of modes of expression; the establishment of a common language between map publishers, rather than standardisation of map programmes. The variety of programmes is still clearly subordinate to the diversity of conditions and the multiplicity of purposes. Nevertheless, a common programme is needed where a regional or national territory is to be mapped by different authors, and still more so where an international map is to be made.

The most universal general key is the international hydrogeological map key developed and adopted by the IAHS and the IAH in 1967 and subsequently approved and recommended by UNESCO (1970); a revised second edition was published by UNESCO in 1983. A third revised edition was included in the Hydrogeological Mapping Guide prepared by the IAH and UNESCO and published in 1995 (Struckmeier and Margat, 1995).

(iv) Keys to individual maps

The key to a particular hydrogeological map is the expression of a given map programme and is therefore selective; it should include only those signs actually used on the map. Preparing the key and drawing the map are mutually interactive activities.

(v) Derived special-purpose maps

The keys to specific purpose maps that give information about groundwater resources with more practical objectives (Section 10.4 and Table 10.4), must observe the same rules. They are not standardised, but they must be particularly simple and intelligible as they are designed for non-specialist users. They should be easy to produce and frequently updated.

10.6 Producing hydrogeological maps

10.6.1 Preliminary choice

The preliminary choices, area mapped, topographic base maps, and scale, are interdependent:

- The *area* may be imposed by the topographical sheet, e.g. if the map will be part of the 'regular' large-scale or medium-scale coverage of a national territory or, on a smaller scale, by the sheet of an international map. It may also correspond to an administrative or political district (a province or region); this applies to any national map. Or it may be adapted to a natural physical domain such as a sedimentary or river basin or an individual aquifer.
- The available *topographic* base maps often have to be supplemented or revised, in particular as regards the drainage pattern, and simplified.
- The choice of *scale*, dictated first and foremost by the map's purpose, which suggests either a large or a small scale, is also limited by the scale of the available topographic base maps.

10.6.2 Programming

The four main phases in the production of a hydrogeological map have been shown in the flow chart of Figure 10.3.

(i) Data collection and drawing the hydrogeological map

From the outset drawing a hydrogeological map may or may not be one of the objects of regional hydrogeological inventory and prospecting operations. In any case, it is at this stage that useful basic data are collected (see section 10.2.3). But the actual 'drawing of a hydrogeological map' may include more complete, specific observations and sometimes investigations of the sub-surface. This always includes research: the use of any existing documents, published or unpublished (literature, maps, files and so on), including of course earlier hydrogeological reports and studies, not forgetting technical archives (connected with prospection for oil or with mining operations).

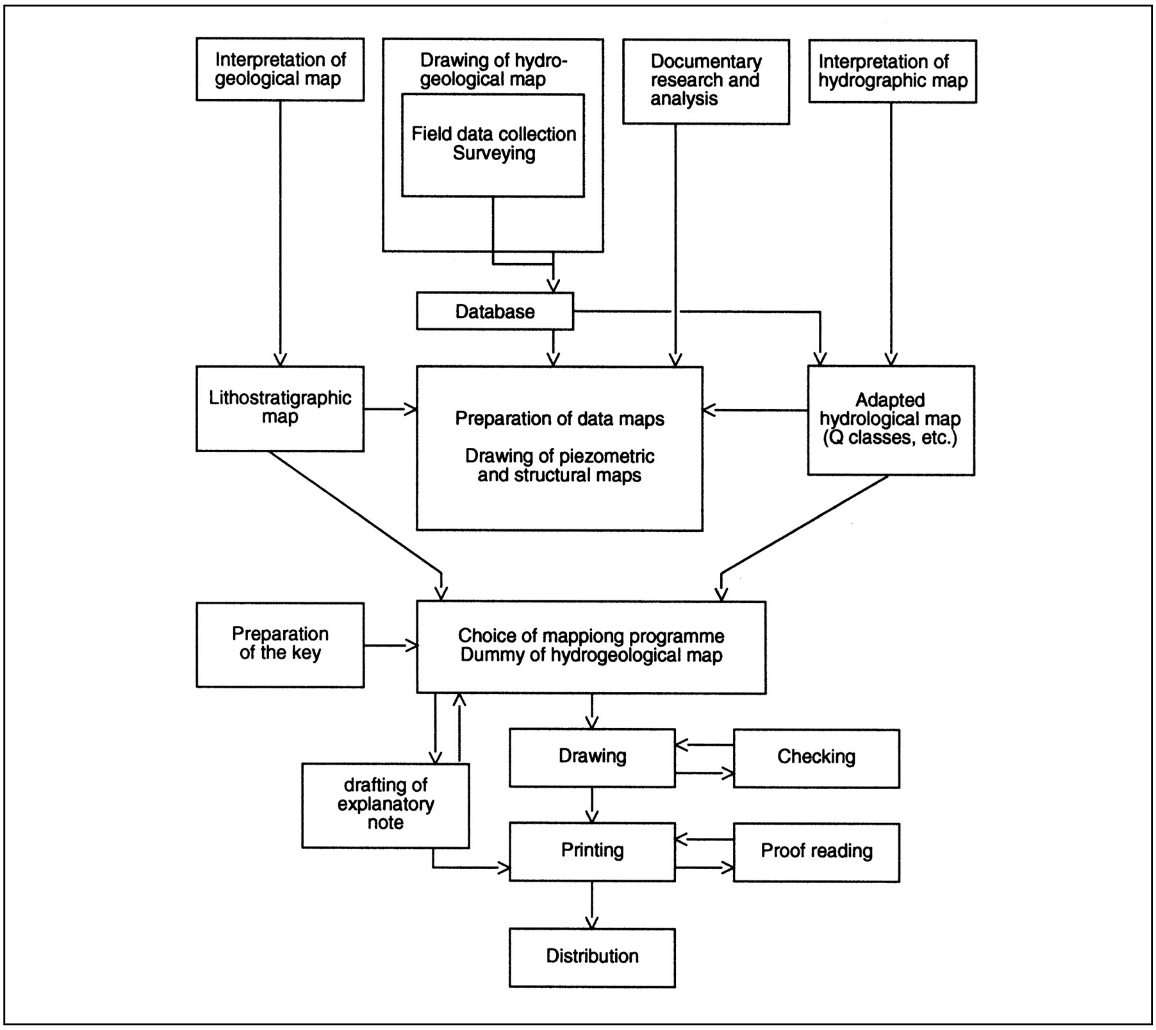
(ii) Data analysis and data preparation

The preparatory phase involves transferring all the data gathered or interpreted from fieldwork (e.g. classified water points, classified and checked hydrographic detail, hydrogeological contours) on to work-maps in each subject at the same scale as the map in preparation. This may include the drawing of parametric maps, e.g. piezometric maps, maps of structural isohypses or transmissivity maps. During this phase gaps and uncertainties may appear, necessitating further fieldwork or reinterpretation of basic data.

(iii) Preparing the dummy

This central phase begins with the choice of the mapping programme and the preparation of the draft *key* on the basis of the prepared information, among which a choice may have to be made according to the purpose of the map. As work proceeds, alterations or improvements to the programme and the key may be desirable. It is advisable to make the dummy as far as possible a '*facsimile*' of the projected map as regards signs and colours, so as to be able to judge the

Figure 10.3 Flow chart of the production of a hydrogeological map



compatibility and perceptibility of the whole. Material shown in the margin (inset maps, sections and diagrams) should also be prepared at this stage.

(iv) Publication

This phase is largely the responsibility of others and neither at the drawing stage nor at the printing stage is it specific to the production of hydrogeological maps. The role of the hydrogeologist and draughtsman is limited to checking and correction at two stages:

- checking the drawing before printing;
- examining and correcting the proofs, particularly with regard to colour.

These two phases can partially be merged by the use of computer-assisted mapping techniques that directly memorise the contours of the original dummy.

10.6.3 Explanatory note

The explanatory note to a hydrogeological map should fulfil three functions:

- Supplementing and explaining the definitions in the key, following the same order. In particular, indicating the degrees of approximation – variables for the same datum according to zone – and the sources, *e.g.* the basic data.

- Giving instructions as to how to use the map; explaining the use of certain data or information.
- Providing additional data or information: this may be mappable but not represented on the main map (to avoid overloading), but appearing on maps appended in the note, or it may be unmappable, e.g. historical data, cross sections, or explanatory, e.g. particulars on methodology used.

10.7 References and additional reading

Including a selection of publications on hydrogeological maps or keys to national or international maps and a number of 'hydro-ecological' maps concerning the vulnerability of groundwater to pollution.

ALBINET, M.; MARGAT, J., 1971. *Cartographie de la vulnérabilité à la pollution des nappes souterraines. Présentation de la carte de France à 1:1,000,000.* IAHS, XVth General Assembly, U.G.G.I., Moscow, Russia.

ALLER, L.; BENNETT, T.; LEHR, J. H. ET AL., 1985. *DRASTIC: A Standardized System for Evaluating Groundwater Pollution Potential Using Hydrogeological Settings.* EPA/600/2-85/018, Government of United States of America, Washington D.C.

ANDERSEN, L. J.; GOSK, E., 1989. Applicability of Vulnerability Maps. *Environ. Geol. Water Sciences,* Vol. 13, pp. 39–43.

BARROCU, G.; BIALLO, G., 1993. *Application of GIS for Aquifer Vulnerability Evaluation; Application of Geographic Information Systems in Hydrology and Water Resources Management*. IAHS Publ. N° 211.

BERTIN, J., 1973. *Simiologie Graphique*. Gauthier-Villàrs, 2e éd., Paris et La Haye, 431 pp.

BRGM, 1992. *Cartographie Hydrogéologique de l'Afrique*. Hydrogéologie, N° 1-2, Éditions BRGM, Orléans, 122 pp.

CASTANY, G., 1968. *Prospection et exploitation des eaux souterraines* (Chap. 13: Cartographie des eaux souterraines). Dunod, Paris, France, 717 pp.

CHAMPAGNE, L.; CHAPUIS, R. P., 1993. Évaluation et cartographie de la vulnérabilité à la pollution des formations aquifères de la MRC de Montcalm selon la méthode DRASTIC. *Sc. et Tech. de l'Eau,* Vol. 26, N° 3.

COLLIN, J. J., 1989. An Attempt to Evaluate the Socio-Economic Impact of Hydrogeological Maps. IAH, *Mem. Intern. Symp.: Hydrogeological Maps as Tools for Economic and Social Development, Hannover, Germany,* pp. 319–31.

GALOFRE, A., 1976. Mapas hidrogeológicos. In: E. Custodio y M. R. Llamas, *Hidrología subterránea,* Ediciones Omega, Barcelona, pp. 1545–57.

GILBRICH, W. H.; HEINDL, L. A.; KELLER, R.; NEWSON, W. D.; ROGOVSKAYA, N. Y.; S. BUCHAN, (eds.) 1977. *Hydrological maps*. UNESCO/WMO, Paris/Geneva, 204 pp.

GRIMMELMANN, W. F., 1989. Hydrogeological Mapping in the Federal Republic of Germany. IAH, *Mem. Intern. Symp. on Hydrogeological Maps as Tools for Economic and Social Development, Hannover, Germany,* pp. 75–8.

GRIMMELMANN, W. F.; KRAMPE, K. D.; STRUCKMEIER, W., (eds.), 1986. Hydrogeological Mapping in Asia and the Pacific Region. *IAH Intern. Contr. to Hydrogeol.,* Vol. 7, 410 pp.

IAH, 1989. *Mem. Intern. Symp. on Hydrogeological Maps as Tools for Economic and Social Development, Hannover, Germany.* IAH, Hannover, 598 pp.

KARRENBERG, H.; STRUCKMEIER, W., 1978. The Hydrogeological Map of Europe. Episodes, *IUGS Geol. News, 1978/4,* pp. 16–8.

KHOURI, I.; DROUBI, A.; ZEBIDI, H.; ZERYOUHI, H.; I.; HAWA; A., 1982. *ACSAD Legend for the Water Resources Map of the Arab Countries.* ACSAD, Damascus, W.R./p-31, 29 pp.

KOVALEVSKY, V. S.; MAKSIMOVA, N. G., 1989. Maps for Predicting the Variability of Groundwater Resources in the Future. IAH., *Mem. Intern. Symp. on Hydrogeological Maps as Tools for Economic and Social Development, Hannover, Germany,* pp. 83–91.

LAU, J. E., COMMANDER, D. P.; JACOBSON, G., 1987. Hydrogeology of Australia; Map at Scale 1:5,000,000 and Text. *BMR Bull.,* No. 227, Canberra, 21 pp.

MARGAT, J., 1966. La cartographie hydrogéologique. *Chron. d'Hydrogéol.,* BRGM, Paris, Vol. 9, pp. 7–32.

——, 1978. *Nouvelle légende de carte hydrogéologique*. Doc. BRGM 78 SGN 473 HYD, 45 pp.

——, 1981. La carte hydrogéologique de la France à 1 : 1 500 000. Systèmes aquifères. *Bull. BRGM (2) III*, Orléans, pp. 98–9.

MARGAT, J.; ROGOVSKAYA, N., 1979. Cartographie des ressources en eaux souterraines. *IAH, Intern. Sympos. Vilnius, U.S.S.R, Rapport général, thème VIII, Mém. IAH, Vol. XV-2.* Moscow (Doc. BRGM 79 SGN 783 HYD J.M.)/Orléans, 18 pp.

PALOC, H., 1975. Cartographie des eaux souterraines en terrains calcaires. In: *Hydrogeology of karstic terrains,* IAH publ., IUGS, Série B, No. 3, pp. 137–48.

SARIN, A., 1989. What is a Hydrogeologic Map and for What Does It Serve. IAH, *Mem. Intern. Symp. on Hydrogeological Maps as Tools for Economic and Social Development,* Hannover, pp. 31–40.

STRUCKMEIER, W.; ENGELEN, G. B.; GALITZIN, M. S.; SHAKCHNOVA, R. K., 1986. Methods of Representation of Water Data. In: G. B. Engelen and G. P. Jones (eds.), *Developments in the Analysis of Groundwater Flow Systems,* IAHS Publ. 163, pp. 47–63.

STRUCKMEIER, W. 1989. Types and Uses of Hydrogeological Maps. IAH, *Mem. Intern. Symp. on Hydrogeological Maps as Tools for Economic and Social Development,* Hannover, pp. 17–30.

STRUCKMEIER, W. F.; MARGAT, J., 1995. Hydrogeological Maps; A Guide and Standard Legend. IAH, *Intern. Contr. to Hydrogeology,* Vol. 17. Heise, Hannover, 177 pp.

UNESCO, 1970. *International Legend for Hydrogeological Maps.* UNESCO/IAHS/IAH/Inst. Geol. Sci., London, 101 pp.

UNESCO, 1983. *International Legend for Hydrogeological Maps, Revised Draft*. UNESCO Techn. Document, SC-84/WS/7, Paris, 51 pp.

VAN DUIJVENBOODEN, W.; VAN WAEGENINGH, H. G., 1987. Vulnerability of Soil and Groundwater to Pollutants. TNO Com. Hydrol. Research, The Netherlands, *Proceed and Inform.,* N° 38, pp. 433–6.

VARTANYAN, G. S.; SHPAK, A. A.; GALITSIN, M. S.; OSTROVSKY, V. N.; SHAKHNOVA, R. K.; L. A. OSTROVSKY, 1989. Main Conception and Experience of Hydrogeological Mapping Development in the USSR. IAH, *Mem. Intern. Symp. on Hydrogeological Maps as Tools for Economic and Social Development,* Hannover, pp. 269–76.

VILLUMSEN, A.; JACOBSEN, O. S.; SONDERSKOV, C., 1983. *Mapping the vulnerability of groundwater reservoirs with regard to surface pollution.* Danm. Geol. Unders. Arbog. 1982, Copenhagen, pp. 27–38.

ZAPOROZEC, A.; VRBA, J., (eds.), 1995. Mapping of Groundwater Vulnerability. IAH Groundwater Protection Commission, IAH, *Intern. Contrib. to Hydrogeology,* Vol. 16. Heisse, Hannover.

11 Assessment of groundwater resources and groundwater regime forecasting

11.1 Introduction

An assessment of the groundwater resources of an aquifer system involves developing of a quantitative understanding of the flow processes which operate within the aquifer. Three features must be considered:

- how water enters the aquifer system,
- how water passes through the aquifer system and
- how water leaves the aquifer system.

Groundwater regime forecasting involves a study of modifications to the flow processes due to changes in any of these features; these changes may be natural or man-made.

The assessment of groundwater resources requires a detailed understanding of the aquifer system. Information gained from the techniques described in the preceding chapters is vital for the development of an understanding of the aquifer system. If conditions in the aquifer have been modified due to the withdrawal of water from the aquifer and the effects of this exploitation have been monitored over a number of years, it is easier to identify the groundwater flow processes and to predict further changes. The key to the assessment of groundwater resources is the *formulation*, which involves the description of the essential features of the flow processes. Groundwater problems are complex in three-dimensional space and usually vary in time; it is rarely possible to have complete information about all features such as the variations in the component permeabilities within the aquifer system. The primary aim of the formulation is to identify the *essential* features which strongly influence the flow processes; the following references contain valuable information about the formulation of groundwater problems (Bear, 1979; Bouwer, 1978; Balek, 1989; Driscoll, 1986; Freeze and Cherry, 1979; Marino and Luthin, 1982; de Marsily, 1986; McWhorter and Sunada, 1977; Rushton and Redshaw, 1979; Todd, 1980; Walton, 1991).

To quantify the flow processes a model is introduced. The model may be simple, such as the use of Darcy's Law (Chapter 1) in one-dimension, or it may be a complex three-dimensional time-variant representation; in practice it is advisable to use both simple and complex models. For complex problems a model in the form of a computer package is often used to obtain numerical results. However, it is essential first to carry out a detailed formulation and then choose a suitable package, rather than selecting a package and modifying the physical problem to fit the assumptions and idealisations inherent in the package. This issue is similar to the frequent misuse of methods of pumping test analysis (Chapter 8) when a type curve method is selected which has a similar shape to the field results even though the conditions for which the type curve is devised are very different from those in the field.

When historical responses have been represented adequately by a model, the model can then be used to forecast the likely consequences of changes in inflows, conditions within the aquifer system or outflows. Typical examples of forecasting include a study of changes in the exploitation of the aquifer system and the effect on the water table, stream flows or quality of the abstracted water. If the changes being investigated result in aquifer conditions which are significantly different from the current conditions, predictions must be treated with caution

The stages in an investigation are as follows:

(i) collect together all the hydrological and hydrogeological information and data for the catchment,
(ii) identify the important flow processes and confirm the magnitude of the components by carrying out a preliminary flow balance,
(iii) develop a preliminary model to see whether the model reproduces the general trends of the aquifer response,
(iv) in the light of the initial modelling results, review all the hydrogeological information and field data and develop an improved understanding of the flow processes. This may involve further field investigations and the use of the model to investigate its sensitivity to changes in various parameters,
(v) produce a revised model,
(vi) repeat the stages listed above until the model reproduces hydrographs of groundwater heads and flows to a prescribed accuracy,
(vii) the model can then be used for predictive purposes.

Even if the investigation does not make use of a computer model, most of the stages listed above should be used in an investigation.

11.2 Identifying and quantifying groundwater resources

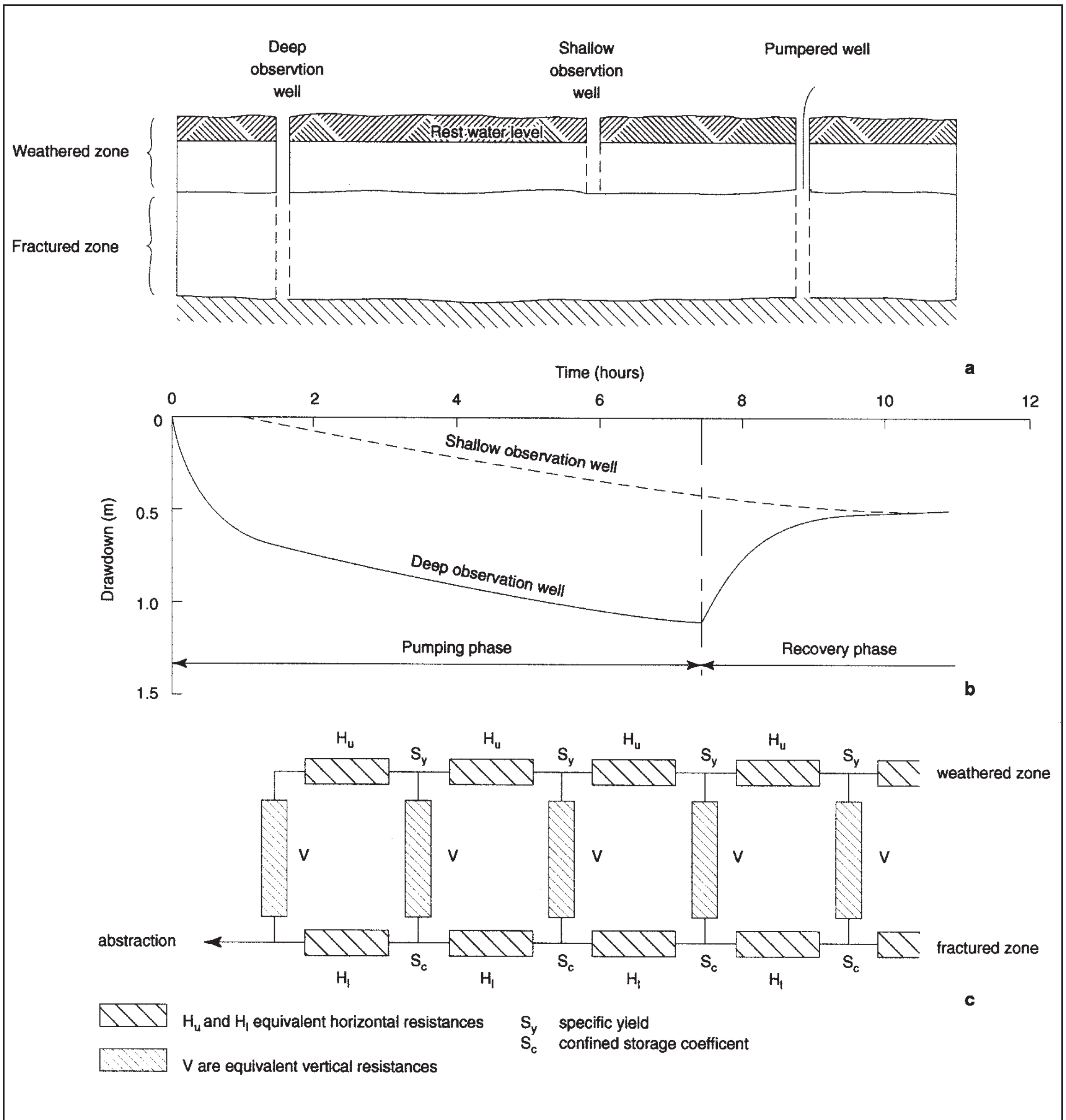
11.2.1 Formulation

The formulation of a groundwater problem involves identifying the critical flow processes including the inflows and outflows. This section considers the formulation of the physical problem; the formulation in mathematical terms is considered in Section 11.3.2. An inadequate understanding of conditions in an aquifer may lead to unwise exploitation, with serious consequences for the reliability of the aquifer model. Three case studies representative of most groundwater problems have been selected, to introduce how a physical formulation is carried out; these examples from India are *outlined* below and are subsequently used to illustrate procedures throughout this chapter.

The groundwater resources of the Dulapally Basin in the hard rock area of Central India have been exploited for many centuries using large-diameter dug wells in the weathered zone of a granitic aquifer. In an attempt to increase the yield, boreholes were drilled into the underlying fractured zone, Fig. 11.1, and good yields were obtained. It was assumed that a new aquifer had been found, primarily because the yield of the fractured zone was good and there was little effect on the weathered zone during test pumping. However, following the drilling of many production boreholes in the fractured aquifers, yields fell off rapidly. Dewatering of the weathered zone also occurred, leading to a severe reduction in yields of the original large-diameter wells. In assessing the groundwater resources, the investigators had failed to identify the interconnection between the weathered and fractured zones. This example shows that a study of a multi-layered aquifer must allow for the possibility of transfer between the layers.

A similar situation occurs in many alluvial aquifers. A typical example is provided by the Mehsana alluvial aquifer, Fig. 11.2. Falling yields of shallow wells in the near surface aquifers

Figure 11.1a, b, and c Example of weathered-fractured aquifer in India: (a) pumped and observation wells, (b) general form of drawdowns for pumping test, (c) numerical model to represent aquifer system



prompted the drilling of deep boreholes through clay zones to tap the deeper aquifers. Often these original trial boreholes tapped a plentiful supply of water and it was possible to identify a distant recharge zone which appeared to supply the deeper aquifers, Fig. 11.2a. However, as extensive exploitation of the aquifer occurred ,with many production boreholes drilled into the deeper aquifers, the yields of boreholes decreased. In addition, in the water table of the upper aquifers dropped.

Figure 11.1d and e Example of weathered-fractured aquifer in India, (d) detailed comparison between field and modelled results, (e) main flow mechanisms

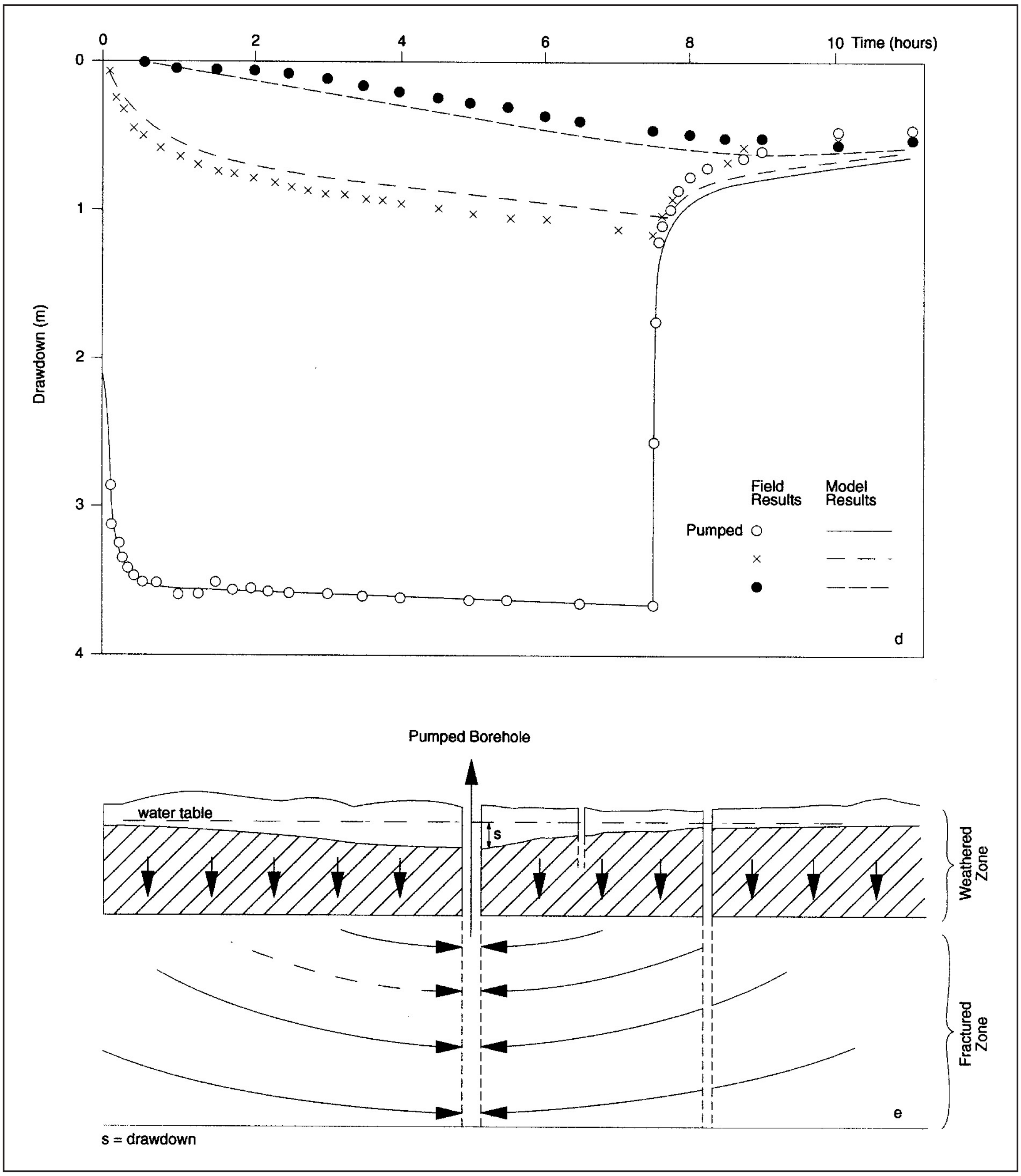


Figure 11.2a and b Example of alluvial aquifer: (a) simplified diagram of aquifer system, (b) history of exploitation of the aquifer system

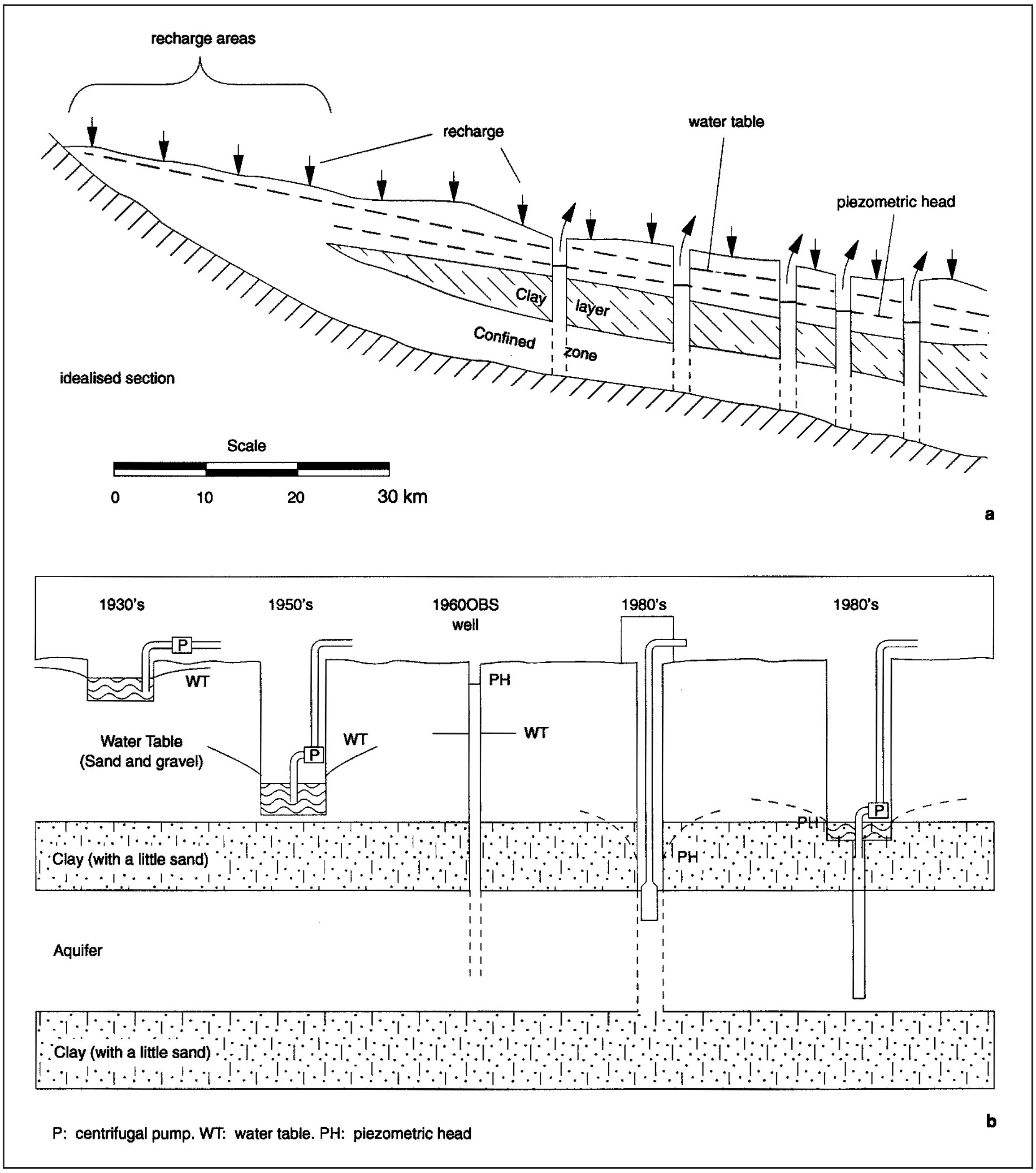


Figure 11.2c and d Example of alluvial aquifer: (c) response of shallow and deep observation wells, (d) change of flow mechanisms from before to after exploitation

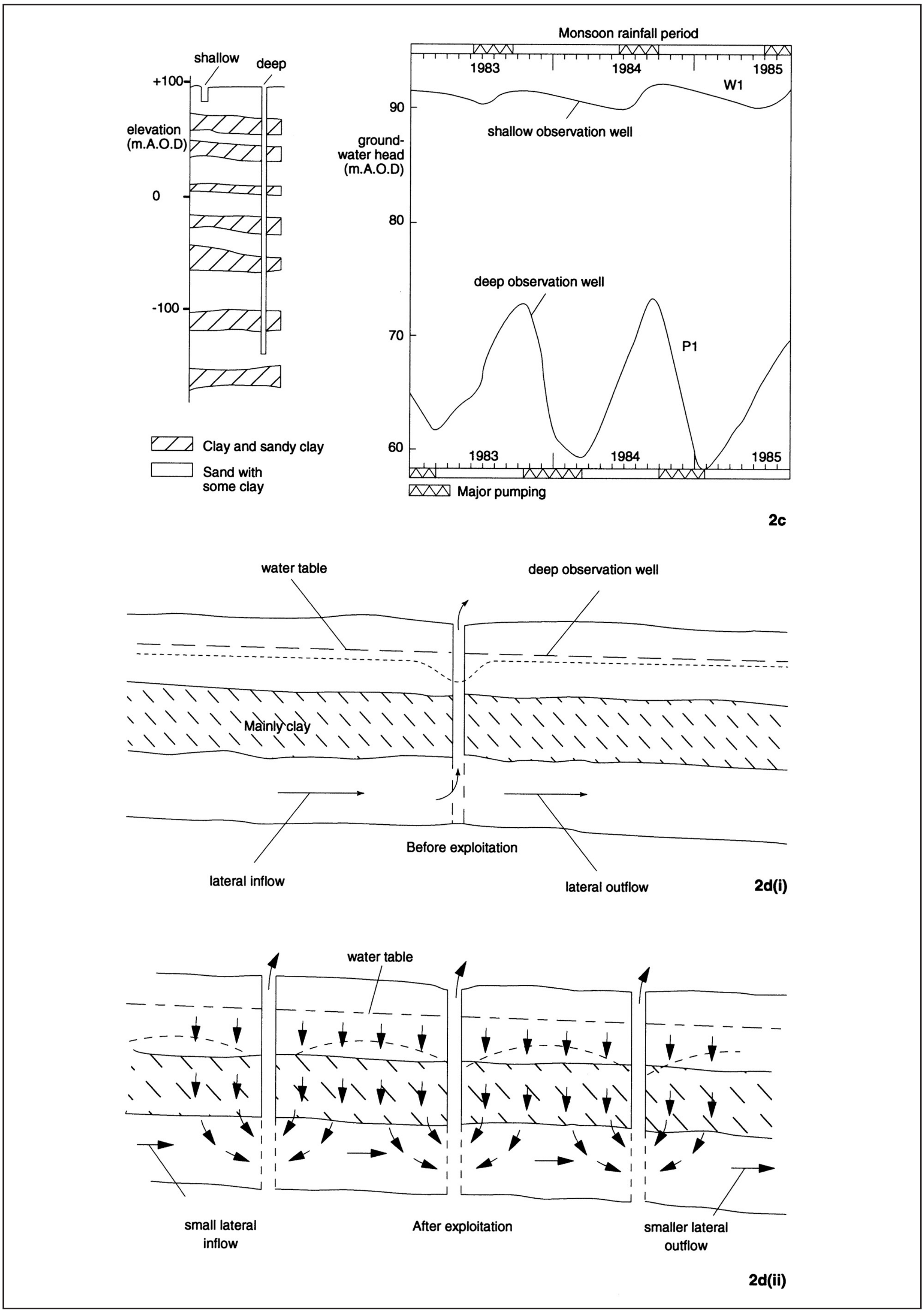


Figure 11.2e Simplified model of the aquifer system

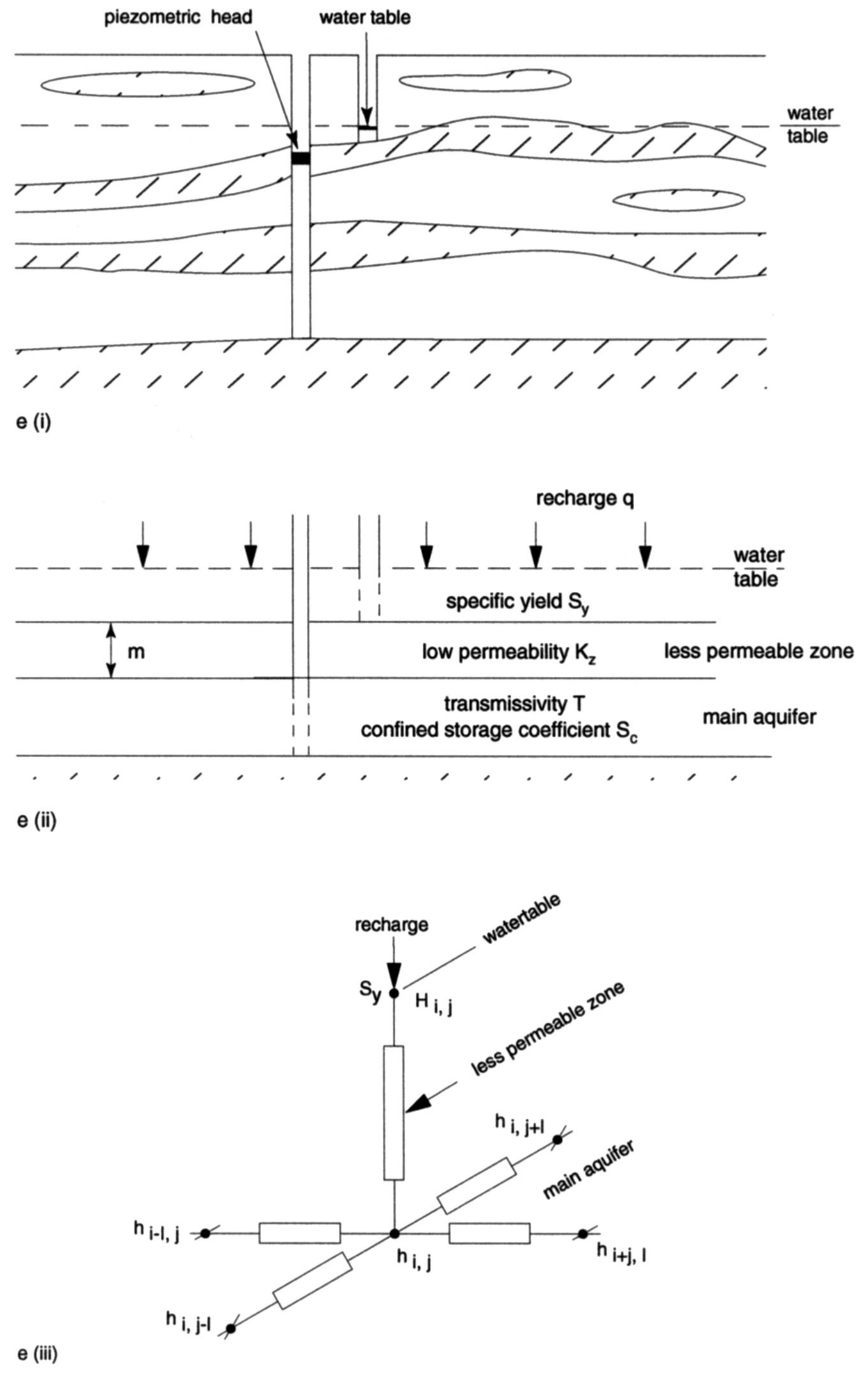
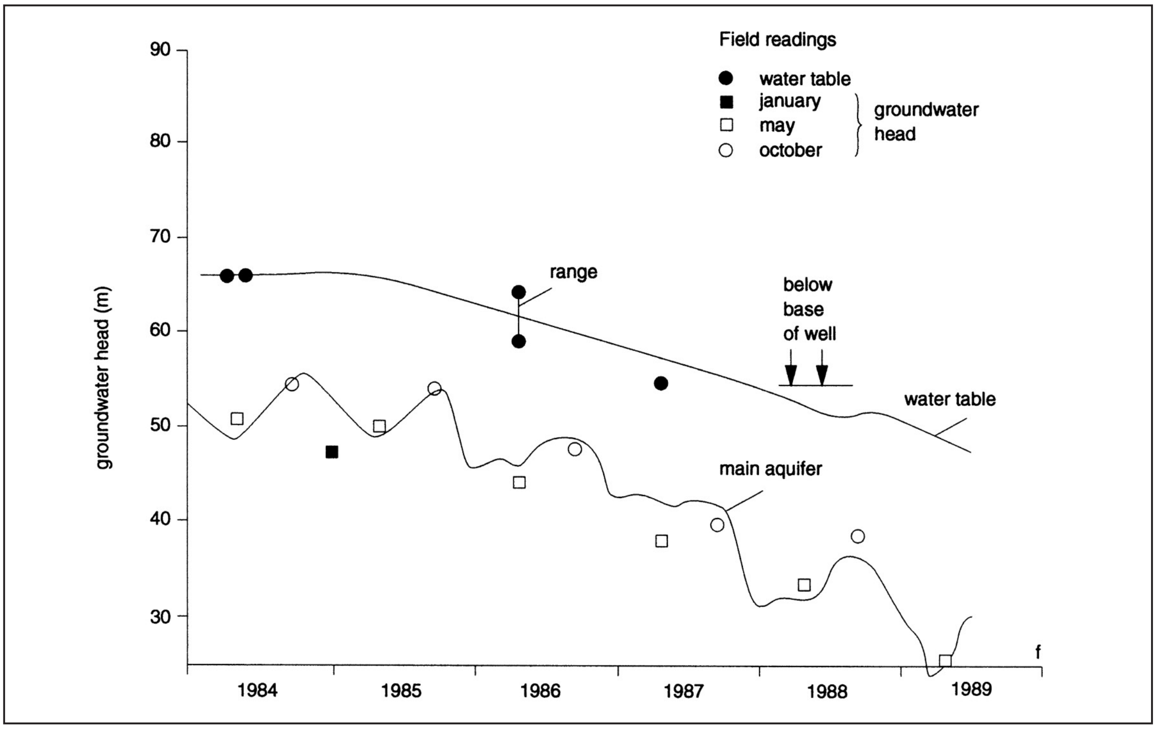


Figure 11.2f Comparison of field and modelled groundwater head fluctuations



The third example refers to a Miliolite Limestone aquifer in coastal Western India in which excellent yields occur under high water table conditions but the yield reduces significantly as the water table falls, Fig. 11.3. Another issue is that encroachment of saline water occurs within two kilometres of the coast, Fig. 11.3a. Artificial recharge was proposed as a means of flushing out the saline water and improving the overall yield of the aquifer system. However, pilot schemes indicated that neither of these goals would be achieved.

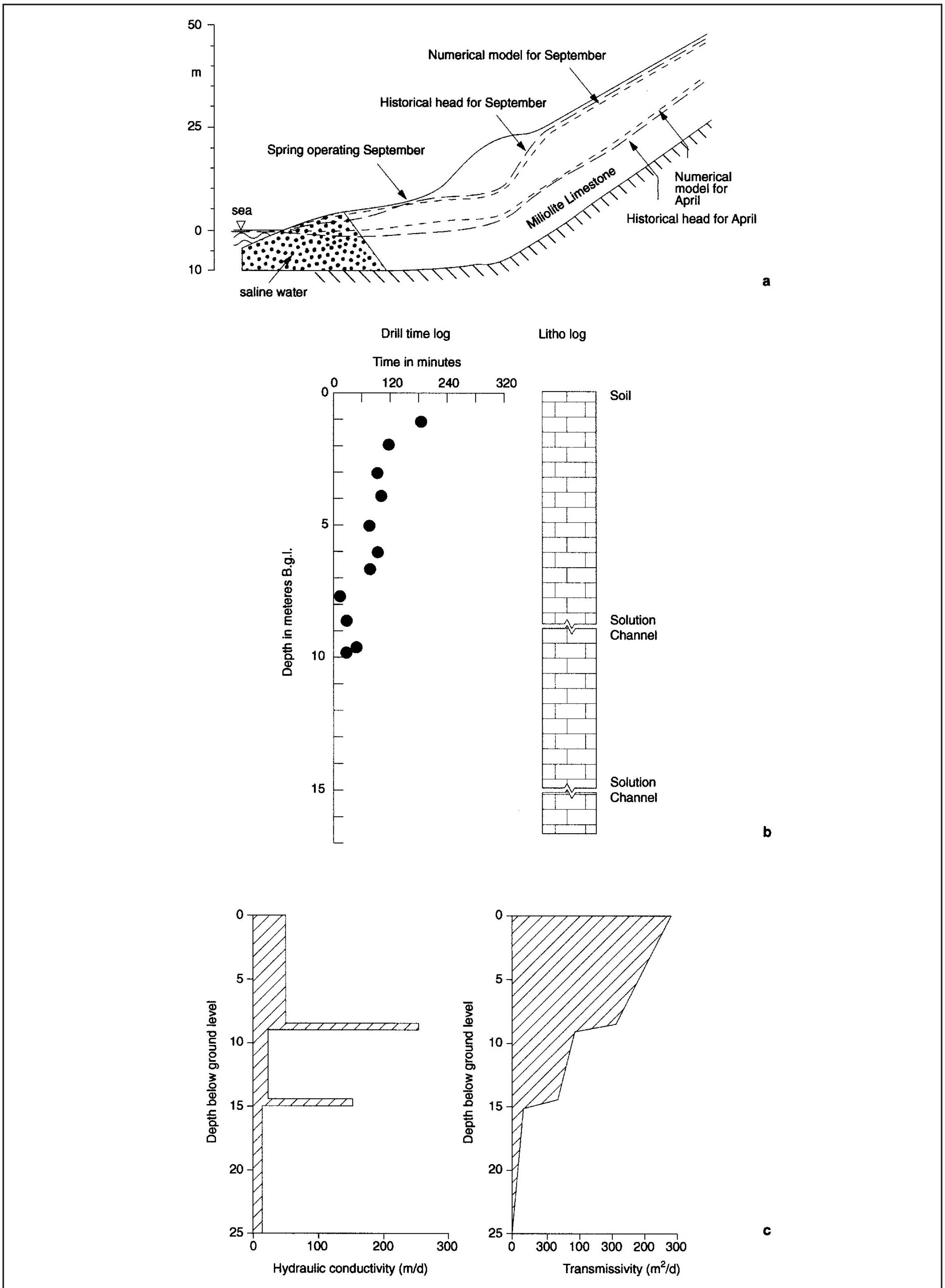
These examples demonstrate that it is often difficult to identify the flow mechanisms resulting from the exploitation of aquifers, since these mechanisms may not be operating when the initial investigations are carried out. When preparing an assessment of the groundwater resources, the various components of the flow processes must be considered. These are discussed below.

11.2.2 Groundwater resource components

(i) Inflows

Recharge is the major inflow to most aquifers; methods of estimating recharge are discussed in Lerner et al. 1990, Simmers 1997. Recharge may result from precipitation and also occur due to return flow from irrigation. In arid areas, the main recharge may be from wadis which flow for only a short period most years. Other sources of recharge are permanent water bodies (such as lakes, rivers, canals, flooded rice fields) and leaking water mains or sewers. It is important to distinguish between the potential recharge which is available at the ground surface and the actual recharge which reaches the body of underground water; the actual recharge may be less than the potential due to factors such as the presence of poorly permeable zones between the ground surface and the permanent water table, or the inability of the aquifer to accept all the potential recharge.

Figure 11.3 Example of Miliolite limestone aquifer: (a) typical cross-section through aquifer system, (b) information about drilling, (c) assumed variation of hydraulic conductivity and transmissivity with depth



Flows from adjacent strata can be another source of inflow, but it is essential to perform an approximate water balance of the adjacent strata to ensure that there is sufficient water available. It is preferable to deduce the inflow from a water balance rather than using a Darcy's Law calculation based on uncertain values of transmissivity and groundwater head gradient to estimate the cross-boundary flows.

Water from storage is a significant input when the outflows from the aquifer exceed the recharge and lateral inflows. When the outflows exceed the inflows, water can only be supplied by a fall in the water table thereby releasing water from storage. If the recharge exceeds the outflows, the excess water will be taken into storage, leading to a rise in the water table.

(ii) *Outflows*

Discharge areas can occur in the lower areas of a groundwater catchment; it can be difficult to determine the actual flow in discharge areas because of the diffuse nature of the outflow, which cannot be measured; often some of the discharge is lost to evaporation.

Flows to adjacent aquifers; comments are similar to those concerning inflows from adjacent strata.

Rivers and springs can provide natural outlets for water. One technique for identifying the groundwater flow into rivers is baseflow separation. However, for investigations using models it is possible to derive a relationship from which the flow to or from a river can be established. The outflow is usually proportional to the difference between the groundwater head in the aquifer and the elevation of the river or spring. Rivers can also act as sources of inflow when the groundwater head falls below the river level; a more detailed discussion can be found in the section on *Flow Processes*.

Rejected recharge may be significant in areas where the surface permeability of the aquifer is low or where the aquifer is already full of water. In many hard rock aquifers, the infiltration capacity is limited and much of the potential recharge is rejected and enters drainage channels and rivers.

Abstraction from an aquifer is the one outflow which can usually be quantified. Even if the pumpage is not measured directly, reasonable estimates of the volume of water withdrawn from the aquifer can be obtained from an examination of the domestic, industrial or agricultural use of the water.

(iii) *Flow processes within the aquifer system*

Nature of aquifer: information about the nature of the aquifer combined with the location and magnitude of the inflows and outflows allows a picture to be developed of the flow processes within the aquifer system. For example, if there are poorly permeable zones in the vicinity of the water table, this may restrict the recharge, while very permeable zones can increase the yield of boreholes.

Influence of very permeable zones: very permeable horizontal zones allow rapid horizontal flow towards a borehole or a spring. However, the question which needs to be asked is: from where does the very permeable zone collect water? Frequently the very permeable horizontal zones attract water by upward or downward flow through poorly permeable zones.

Influence of poorly permeable zones: significant vertical flows can occur through poorly permeable zones even though their permeability is at least three orders of magnitude lower than the very permeable zones. For the poorly permeable zones the total vertical flow can be estimated from the product of a small vertical velocity and a large plan area, whereas for a very permeable zone the higher horizontal velocity is multiplied by a far smaller vertical cross-section.

Fissures, fractures and *faults* can all have a strong influence on the flow processes. Fissures and fractures often enhance yield from a well or borehole. Because of the convergent flow towards a pumped well or borehole, there are high velocities within ten metres of the borehole and there is a significant groundwater head loss in this distance. The higher permeability of fissures and fractures can reduce the head losses in the vicinity of the pumped boreholes. The influence of faults is more difficult to quantify; in some circumstances a fault with a large throw isolates parts of an aquifer system. However, a series of smaller faults may enhance the permeability.

River/aquifer interaction can provide a route for water to leave an aquifer and increase river flows if the groundwater level is higher than the river elevation, or a river can recharge an aquifer if the groundwater level is below the river elevation. A form of river/aquifer interaction is illustrated in Fig. 11.4; in the following discussion the groundwater head can refer either to the water table close to the river or to the groundwater head within the aquifer adjacent to or beneath the river.

11.2.3 Examples of identification and formulation of aquifer flow mechanisms

(i) Weathered-fractured aquifer

The workers who carried out the initial analysis of the pumping test in the Dulapally weather-fractured aquifer assumed that the fractured zone is a separate aquifer from the weathered zone. Consequently when tube-wells were drilled into the fractured zone and cased out through the weathered zone, it was believed that horizontal flow through the fractured zone would dominate. This view was supported by good yields during the pumping tests which were obtained with relatively small drawdowns. Furthermore the Theis method was used to analyse the observation well readings obtained from the pumping test; in this analysis the fractured zone was represented as being confined, with no hydraulic contact with the weathered zone.

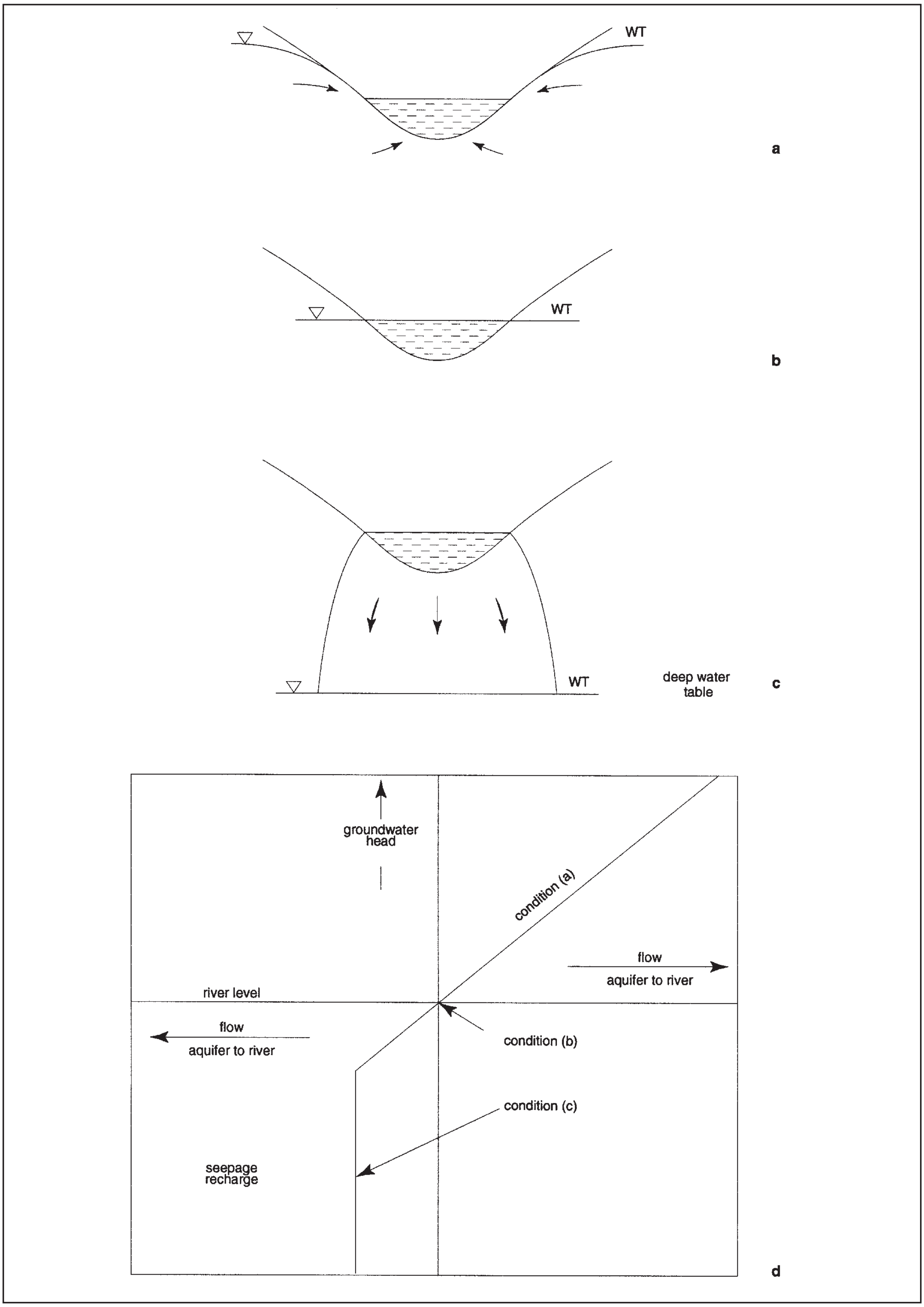
However, some of the field results from the pumping test were ignored. Figure 11.1b shows that, in addition to the rapid fall of the piezometric heads in the observation wells in the fractured aquifer, the shallow observation wells in the weathered zone respond slowly, with the fall of the water table *continuing after pumping stopped!*

To investigate this observed response, a conceptual model (Fig. 11.1c) was developed in which there is a possibility of flow between the fractured and weathered zones. In fact the difference between the drawdowns in the fractured and weathered zones (Fig. 11.1b *proves that vertical flows do occur!* The difference in the groundwater heads increases to a steady value of about 0.8 m; when pumping stops the difference decreases only slowly, demonstrating that vertical flows continue from the weathered zone to the fractured zone to refill the fractured aquifer. This example shows that the difference in groundwater heads between two observation wells at different depths within the aquifer system is of crucial importance in identifying the aquifer flow mechanisms. The quantification of these flows can be achieved using a model as outlined in section 11.3.4.

(ii) Alluvial aquifer

During initial investigations when deep trial boreholes were constructed into the deeper zones of the Mehsana alluvial aquifer, the groundwater heads reflected the higher heads of the common recharge zone, Fig. 11.2a. These higher heads led to the assumption that a new aquifer had been identified. However, as many production boreholes were drilled and the deeper aquifers became heavily exploited and the groundwater heads fell; it then became possible to

Figure 11.4 River/aquifer interaction: (a)-(c) effect of different relative positions between the water table and the river level, (d) assumed river/aquifer interaction



identify the nature of the flow mechanisms. Figure 11.2c shows the hydrographs of a shallow piezometer and a deep observation borehole. An examination of the results shows that:

a. the groundwater heads in the shallow piezometer rise during the recharge period and fall during the dry season,
b. the heads in the deep piezometer fall when heavy pumping for irrigation occurs but rise during the hot season of March to June, when few crops are grown, and continue to rise during the monsoon season,
c. there is a distinctive vertical gradient between the shallow and deep aquifers, indicating that vertical flows are significant and that the vertical resistance to flow is high.

These results allow the important flow mechanisms to be identified, Fig. 11.2d. Diagram (i) shows that when the trial boreholes were drilled, the lateral flow from the common recharge zone was intercepted and a good yield could be obtained for a small pumped drawdown; this only occurred because there was no other pumping from the aquifer. However, when a large number of deep tube wells were drilled, the lateral flows in the aquifer were insufficient to meet the demand and therefore vertical gradients were set up to draw water from the water table in the upper aquifer, diagram (ii). Figure 11.2b illustrates how conditions changed with time within the aquifer system.

Were the conclusions of the initial investigation wrong? With hindsight it can be seen that the flow mechanisms within the aquifer system changed as abstraction increased, with large vertical flows becoming the most important mechanism. However, this could not be identified from the initial tests. When the conditions in an aquifer change from those during the testing stage, predictions of future behaviour can be incorrect.

(iii) Coastal limestone aquifer

The key to the response of the limestone aquifer, Fig. 11.3a, is that there are major solution channels which have developed in the limestone. Figure 11.3b provides information about the drilling of a well in the limestone; major solution channels were identified and mud loss occurred when the drilling reached the first solution channel. These solution channels allow substantial quantities of water to be drawn into a pumped well; a discharge of 5200 m^3/d was achieved in a 0.45 m diameter well for a drawdown of 0.15 m. However, during the dry season when the water table fell below the level of the solution channels, an abstraction rate of 500 m^3/d could not be maintained because of the reduction in transmissivity.

It is not possible to measure the effective permeability of the solution channels but the postulated distributions of the hydraulic conductivities, shown in Fig. 11.3c, explain the large difference in the well yields. When the well water level is at 5 m below ground surface the transmissivity is 720 m^2/d, but when the well water level is 15 m below ground level and below the lowest fissure the transmissivity falls to 100 m^2/d. This varying transmissivity with water table elevation is often the most important factor in understanding the response of aquifers with karstic features (Rushton and Raghava Rao, 1988).

11.2.4 Important questions when identifying resources

The purpose of this section is to list questions which should be considered when identifying the flow mechanisms within an aquifer system prior to quantifying the resources.

a. which flows are more important, horizontal flows, or vertical flows, or both combined?
b. does the aquifer respond quickly to recharge events or changes in abstraction (i.e. within one or two months) or is the response much slower?

c. is there a significant change in the saturated depth (a change of more than 10% is usually significant)?
d. does the actual recharge reaching the water table equal the potential recharge from below the soil zone?
e. does the lowering of the water table cause the recharge to increase or decrease?
f. are conditions within the aquifer system likely to change with time (such as the dewatering of the upper part of the aquifer or the cessation of groundwater flow into streams and rivers)?

11.2.5 Selection of the co-ordinate system

A further important issue is the selection of the co-ordinate system to be used in describing the problem. A number of alternatives are listed below, in each case the groundwater head is expressed as a function of certain co-ordinates, ***z*** is always the vertical coordinate positive in the upwards direction.

h(x), because the flow is predominantly in one horizontal direction, the flow is only a function of the horizontal coordinate (this will not occur if boreholes are used to abstract water from an aquifer).

h(r), radial flow; this assumption can apply to flow towards a borehole in a confined aquifer.

h(x,y), regional groundwater flow formulation; even though the groundwater head is a function of the horizontal *x-y* coordinate system, vertical components of flow are indirectly included, provided the length to thickness ratio of the aquifer system is greater than five and there are no poorly permeable strata within the aquifer system (Connorton, 1985; Rushton and Rathod, 1985).

h(x,z), flow in a representative vertical section; this approach is important for problems such as the flow through earth dams.

h(r,z), combined radial and vertical flow which occurs around a pumped borehole in a layered aquifer.

h(x,y,z), this is a full three-dimensional formulation; for a layered aquifer such as an alluvial or sandstone aquifer an alternative approach is to consider a series of permeable aquifer layers interconnected by vertical flows through the less permeable strata.

An unsatisfactory selection of the coordinate system can mean that important flow mechanisms are not represented; this can invalidate the analysis. Steady state conditions are unusual and therefore the time variant response of the aquifer system should be included; the groundwater head is then written as *h(x,t)*, *h(z,t)*, *h(x,y,t)*, etc. For quick response aquifers it may be sufficient to consider only two or three years but for slow response aquifers such as extensive alluvial or sandstone aquifers, field information over a period of twenty or more years is desirable.

11.2.6 Preliminary flow balances

A vital step prior to the preparation of a mathematical model is the preparation of a water balance which includes inflows, outflows and changes in storage. Reliable estimates can be made of certain inflows (such as recharge) and outflows (such as abstraction) but for other components, such as the change in storage, estimates are based on field measurements of the groundwater head fluctuations multiplied by the estimated specific yield.

A careful selection of the periods over which the water balances are carried out can increase the usefulness of the calculation. In cases where there is a recharge period each year it is helpful to estimate the water balance three or four times each year. This will ensure that account is taken of the balance between recharge and changes in storage It is unlikely that the components of the water balance will sum to zero but any serious out of balances require careful investigations.

A preliminary water balance for the case study of the Miliolite Limestone aquifer can be found in Table 11.1; the basis on which the estimates are made is explained in notes following the table. It is inevitable that this form of water balance will involve uncertainties. Nevertheless the balance will show whether there is an approximate understanding of the flow components. As modelling proceeds it is possible to refine the flow balance.

11.3 Use of models for quantifying groundwater resources

11.3.1 Fundamentals of groundwater modelling

Any groundwater model is based on two principles:

(i) Darcy's Law which relates the velocity of flow to the hydraulic conductivity and the head gradient,

(ii) Continuity conditions, which ensure continuity of flow from inflows, through the aquifer system to the outflows (water budget).

The three-dimensional time-variant equation for saturated groundwater flow based on these two principles is:

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right)=S_S\frac{\partial h}{\partial t} \tag{11.1}$$

To study a particular problem it is necessary to specify the hydraulic conductivities, K_x, K_y and K_z the specific storage, S_S throughout the study area, to identify the location of all the boundaries and to specify flow or head conditions on each boundary. These boundary conditions, which may be external or internal, represent inflows such as recharge or flows from other strata or outflows such as rivers, springs or flows from pumped wells and boreholes.

Table 11.1 Approximate water balance for Miliolite Limestone aquifer; all quantities are in Ml/d

	1982			*1983*		
Notes	*Jan– Apr*	*May–Aug*	*Sep–Dec*	*Jan–Apr*	*May–Aug*	*Sep–Dec*
Recharge[1]	0	130	190	0	490	0
Abstraction[2]	–120	–70	–10	–220	–60	–80
Storage[3]	+130	–60	–130	+240	–400	+70
Springs[4]	0	0	0	0	–5	0
Coast[5]	–6	–12	–30	+3	–10	–30

Notes:

1. Recharge is estimated from a daily soil moisture balance.
2. Abstractions are estimated from the number of groundwater structures, their withdrawal rates, areas of crops grown and the crop water requirements.
3. Storage changes are based on the estimated specific yield of 0.12 multiplied by the average change in water table elevation; a rise in the water table means that water is released from storage to flow through the aquifer and is indicated as '+'.
4. Spring flows are based on field observations.
5. Flows across the coast ('–' means an outflow to the sea) are estimated from the groundwater gradient at the coast and the transmissivity; the transmissivity is higher for outflows since the flow occurs in the higher permeable zones.

There is no direct way of solving Equation (11.1) for practical situations. Two approaches are possible:

(i) reduce the number of time and/or space dimensions until the simplified equation can be solved using a suitable mathematical technique to give an expression for the groundwater head distribution; this is called an *analytical solution*. The main limitation of this approach is that the process of removing one or more of the dimensions means that certain of the important flow processes may not be included.

(i) use a *numerical technique* in which the groundwater head is specified only at nodal points; the differential equation can then be written as a series of simultaneous equations which can be solved using a digital computer. A commonly used method is the finite difference method; other suitable techniques include the finite element method and the boundary element method (Anderson and Woessner, 1991; Boonstra and De Ridder, 1981; Liggett and Liu, 1983; Pinder and Gray, 1977; Rushton and Redshaw, 1979).

11.3.2 Various types of models

The most direct models are based on a calculation using inflow, Darcy's Law and outflow. Alternatively, models may be complex three-dimensional time-variant simulations. In most studies it is advisable to carry out preliminary direct calculations before embarking on complex model studies.

Many mathematical models are based on different approximations. However, it is essential to identify the type of model that is required from hydrological and hydrogeological considerations. Consequently, the focus of this discussion will be the different types of model available as summarised in Table 11.2. This is not a complete list but it does indicate a range of models including analytical techniques in which exact solutions can be obtained and numerical techniques that are usually more flexible. The analysis may assume predominantly one-dimensional lateral or radial flow, two-dimensional flow, multi-layered flow or fully three-dimensional flow.

There are many computer packages available which can be used by workers who have little or no experience in numerical analysis. These packages usually correspond to one of the types of model listed in Table 11.2. Nevertheless it is essential to carry out the formulation phase of the study before selecting the method or package; if the essential features are not included in the model, the model results may be totally misleading.

A model checks the consistency of the assumed components and parameters associated with the inflows, outflows and flow processes. A model cannot be used to avoid the need to collect and interpret field data, but it can be used to indicate the important processes and parameters which should be considered in detail.

11.3.3 Important issues in developing models

Important information about the development of models can be found in a number of references, including Anderson and Woessner (1991), Kinzelbach (1986), Pinder and Gray (1977), Rushton and Redshaw (1979), Wang and Anderson (1982). The issues summarised below highlight questions which must be resolved during the development of a model.

Size of area to be studied: the issue in selecting the size of the study area is whether to investigate a large catchment or a smaller area of special interest. When studying a large catchment there may not be enough detail in areas of interest. On the other hand, the area of interest may not have boundaries on which hydraulic conditions are known. Often the best

Table 11.2 Different types of groundwater models

Type of model	Purpose
i) Darcy's law calculation ***h(x)*** ***h(z)***	Applied in horizontal plane for flow in confined aquifer. Applied in vertical plane for flow through poorly permeable layers.
ii) One-dimensional flow ***h(x,t)***	Based on Darcy's law and continuity, a differential equation can be derived, $\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) = S\frac{\partial h}{\partial t} - q$ with appropriate boundary conditions; S is the confined storage coefficient or the specific yield; this equation can be integrated to obtain steady state heads for horizontal one-dimensional flow. Alternatively the Dupuit approach can be used. Numerical solutions to this equation using finite differences can be used to examine time-variant responses, the effect of varying saturated depth, etc.
iii) Radial flow to well ***h(r,t)***	Analytical solutions such as the Theis equation and the many further developments can be used to examine predominantly radial flow to pumped boreholes in a wide variety of situations. Numerical solutions can be obtained for the radial flow equation and they can include all the conditions represented by the analytical solutions and other factors such as well clogging, non-steady abstraction, conditions changing between confined and unconfined.
iv) Two-dimensional flow ***h(x,y,t)***	Regional groundwater flow may be represented as two-dimensional in plan with the equation $\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) = S_S\frac{\partial h}{\partial t}$ and appropriate boundary conditions. The transmissivity T may be a function of $h(x,y)$; S_y is the specific yield.
v) Two-dimensional flow in a vertical section ***h(x,z,t)***	$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) = S_S\frac{\partial h}{\partial t}$ with appropriate boundary conditions; S_s is the specific storage.
vi) Combined radial and vertical flow ***h(r,z,t)***	$\frac{\partial}{\partial r}\left(K_r \frac{\partial h}{\partial r}\right) + \frac{1}{r}\left(K_r \frac{\partial h}{\partial r}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right) = S_S\frac{\partial h}{\partial t}$ with appropriate boundary conditions.
vii) Multi-layered flow ***$h_1(r,t)$, $h_2(r,t)$***	Two zone radial flow model; radial flow models as described above are inter-connected vertically to represent intermediate low conductivity zones; a wide range of practical radial flow situations can be studied using this approach.
$h_1(x,y,t)$, $h_2(x,y,t)$	Multi-layered systems such as alluvial and sandstone aquifers can be represented as two-dimensional aquifers in plan interconnected through poorly permeable aquitards, etc.
viii) Three-dimensional flow ***h(x,y,z,t)***	Full three-dimensional solutions in which variable hydraulic conductivity distributions can be included; very extensive field $\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right) = S_S\frac{\partial h}{\partial t}$ information is required to justify this approach.

approach is to develop a two-dimensional model for the whole groundwater catchment and then study a smaller area as a three-dimensional model with boundary conditions taken from the whole catchment model (Ward et al., 1987). In aquifers with a lower transmissivity too large an area is often studied. Rushton and Redshaw (1979) introduced a method of estimating the time taken for a change in conditions at a major feature (such as a river or a well field) to spread a specified distance. Consider an unconfined aquifer with the following properties:

- transmissivity, T = $40\ m^2/d$,
- specific yield, S = 0.03.

The time for the effect of a change in condition to spread for 2 km can be estimated from the equation,

time, t = $2.5\,L^2\,S/T = 2.5 \times 2{,}000^2 \times 0.03/40 = 7{,}500$ days or 20.4 years.

If the simulations covers a period of about 20 years it is sufficient to model an area which extends 2 km on either side of the location at which the changes in major features occur. The model does not need to extend to features 10 km from the area of interest, since the effect will take 510 years to have any influence!

Mesh spacing and length of time steps: when a numerical model is used in which heads are calculated at discrete points and at specified time intervals, the mesh spacings and time increments must be defined. The important criterion is that the mesh interval and time step must be selected to give sufficient detail in areas of rapid change in groundwater head. Advice about the choice of mesh interval and time step can be found in the standard references quoted earlier.

Average conditions: it is usually unwise to attempt to model average conditions, since time variant effects are generally of considerable significance. Average conditions (sometimes called a pseudo-steady state) will not be appropriate if:

- significant variations in saturated depth occur due to large groundwater head fluctuations,
- the annual recharge pattern is erratic,
- abstraction patterns vary significantly during the year.

For most groundwater problems the time-variant fluctuations are an important part of the study.

Starting (initial) conditions: it is essential to select suitable starting conditions. Specified heads are usually unsatisfactory for starting conditions (Wang and Anderson 1982); if the specified heads are not consistent with the assumed transmissivities, these specified heads may be equivalent to unrealistic recharge values. The best approach is to use a dynamic balance in which the simulation of a typical year is cycled a number of times until the annual dynamic changes become regular.

Length of time to be studied: groundwater conditions change only slowly; if the model simulation is too short it is possible that the changes in the aquifer conditions due to factors such as increased abstraction may not become apparent during the simulation. Changes due to increased abstraction from an unconfined aquifer often take tens of years to occur. There is no convenient way of identifying the required length of a simulation; the best advice is to run the simulation for several years more than initially planned, to see whether the effects of changes have dissipated.

Unsaturated conditions: above the water table unsaturated conditions are certain to occur. Whether the unsaturated conditions need to be considered depends on the nature of the aquifer system. As rainfall recharge passes through an unsaturated zone a delay is likely, especially if there are zones of low hydraulic conductivity in the unsaturated zone. Unsaturated conditions can also be important beneath leaking canals. In many practical situations a factor to delay or reduce the vertical movement of water in the unsaturated zone can be introduced, rather than modelling the unsaturated conditions directly.

Quality issues: the above discussion has been concerned with groundwater heads and groundwater flows, but the quality of the water and the movement of any contaminant in the aquifer may be the crucial issue. Information about contaminant transport techniques can be found in Anderson and Woessner (1991); new techniques of quality modelling can be found in recent literature. However, one important consideration is that only when the understanding and modelling of the groundwater *flow* is reliable will the predictions of the *contaminant* movement be adequate.

11.3.4 Description of models used for case studies

The following section outlines the nature of the models used for the three case studies.

(i) Weathered-fractured aquifer

Field data for a pumping test in the weathered-fractured aquifer system is available; consequently, the model used to represent this test is a radial flow model. The model must include the following features, Fig. 11.1c:

- fractured aquifer with horizontal hydraulic conductivity, specific storage and vertical hydraulic conductivity;
- weathered aquifer with horizontal hydraulic conductivity, vertical hydraulic conductivity, specific yield and recharge;
- the fractured and weathered zones are interconnected through the vertical hydraulic conductivities;
- water is only pumped from the fractured zone.

There are five unknown parameters (the vertical hydraulic conductivities are combined) which are deduced by a sensitivity analysis using a two-zone model (Rathod and Rushton, 1991). The aim of the modelling is to reproduce the pumping and recovery responses in the weathered and fractured observation wells (Rushton and Weller, 1985). Because of the variability of a hard rock aquifer, it is unreasonable to expect the field response to be represented exactly by a model which cannot simulate the heterogeneous nature of the fractured aquifer. Nevertheless, the general trends during the 7.5 hours of pumping and the subsequent recovery are reproduced by the model, Fig. 11.1d. From the model results, the overall flow processes can be identified as shown in Fig. 11.1e. The model parameters are deduced to be:

- *weathered aquifer*, horizontal hydraulic conductivity = 0.5 m/d, vertical hydraulic conductivity = 0.3 m/d, specific yield = 0.01,
- *fractured aquifer*, horizontal hydraulic conductivity = 5.0 m/d, vertical hydraulic conductivity = 3.0 m/d, confined storage coefficient = 0.0025.

(ii) Alluvial aquifer

Three types of model were used during the study of this alluvial aquifer:

(i) a vertical section model which represented the complex layering of gravel, sands, silts and clays; this model confirmed the importance of the vertical flow components when increased abstraction occurred from the deeper aquifers (Rushton and Tiwari, 1989);

(ii) a large regional model which idealised the aquifer unit as a system of inter-connected layers of higher and lower hydraulic conductivity;

(iii) a small regional model which concentrated on a region of plan area 4 km by 4 km with a mesh spacing of 250 m; the computational work for this small regional model was carried out on one of the earlier personal computers (Kavalanekar et al., 1992). The complex layered aquifer is represented as consisting of the upper water table aquifer, an intermediate poorly permeable clay zone and a lower aquifer from which most of the abstraction occurs. Lateral flows are small compared to the vertical flows and can be represented as small specified flows on a boundary of the lower aquifer. The advantage of this small study area is that individual tube-wells can be represented; this is not possible with a large regional model having a mesh interval of 2 km. The adequacy of the simulation is demonstrated by comparisons between the field and modelled groundwater heads in the shallow and deep aquifers, Fig. 11.2f; the field information is very limited but it is sufficient to show that the model does represent the major flow processes.

(iii) Limestone aquifer

The limestone aquifer can be represented by a two-dimensional regional groundwater flow model, provided that the change of transmissivity with the groundwater head is included at all nodal points. Relationships between transmissivity and water table elevation in the form of Fig. 11.3c must be included at every node.

A second important feature of the mathematical model is the representation of periods of high recharge. When the recharge is high, water moves rapidly down-dip and leaves the aquifer through springs. The adequacy of the models is demonstrated by a comparison between the field and modelled groundwater heads on a typical section during the pre-monsoon and post-monsoon seasons, Fig. 11.3a. The significance of the springs during the post-monsoon period is apparent from the diagrams since the groundwater head profiles intersect the ground surface.

11.4 Using forecasting to identify safe yields

11.4.1 Forecasting and prediction

During the study of groundwater systems, an investigator is often required to comment on future groundwater conditions. These estimates of future groundwater conditions can be grouped into:

- *forecasting* the effect of natural changes and
- *predicting* the effect of man-made changes.

Hydrodynamic methods based on distributed models are generally used for *predicting conditions in the disturbed groundwater regime,* whereas statistical methods are often used for *forecasting the groundwater regime under natural conditions*.

11.4.2 Forecasting*

A hydrogeological forecast has the following specific features:

- a probabilistic aspect involving the determination of the confidence interval of probable predicted values,

* This section has been based on extensive material provided by Professor V. S. Kovalevsky, Institute of Water Problems, Moscow (Russia).

- a temporal aspect and
- a verification aspect involving the reliability of predicted values.

Forecasts may be made for a natural, slightly disturbed or disturbed groundwater regime. For natural conditions, forecasting only requires data on the natural hydro-meteorological features; for disturbed regimes, both the meteorological features and aquifer flow mechanisms must be considered.

The following time scales apply to forecasts:

- *short term forecasts* are made for a period of several days and are based on factors which have already come into action (rainfall, flood water discharge etc.); they reflect the time-lag of the responses to the external impacts,
- *seasonal forecasts* which cover a period of several months are based both on factors which are partially known at the time of forecasting and on long term inter-relations between the regime components,
- *long term forecasts* are often based either on the inertia of the system and the exposed tendencies of its development or on the relations with other long term factors such as solar activity and climate changes.

Forecasts may not relate directly to specific times, but may be in the form of possible magnitudes of variables without indication of the time of their occurrence. Furthermore, the predicted data may relate to global, regional, basin or point scales. In terms of the predicted elements, forecasts can be divided into two types, one parameter (groundwater head, temperature, or specific discharge for a certain period of time) or combined parameter forecasts (groundwater chemical composition, intra-annual groundwater level distribution).

The choice of the forecasting method depends on the availability of initial information, the duration of the forecast, the accuracy and the purpose. Available methods include the method of hydrogeological analogy, the balance method, probabilistic statistical methods and dynamic stochastic methods.

(i) Forecasting based on hydrogeological analogies

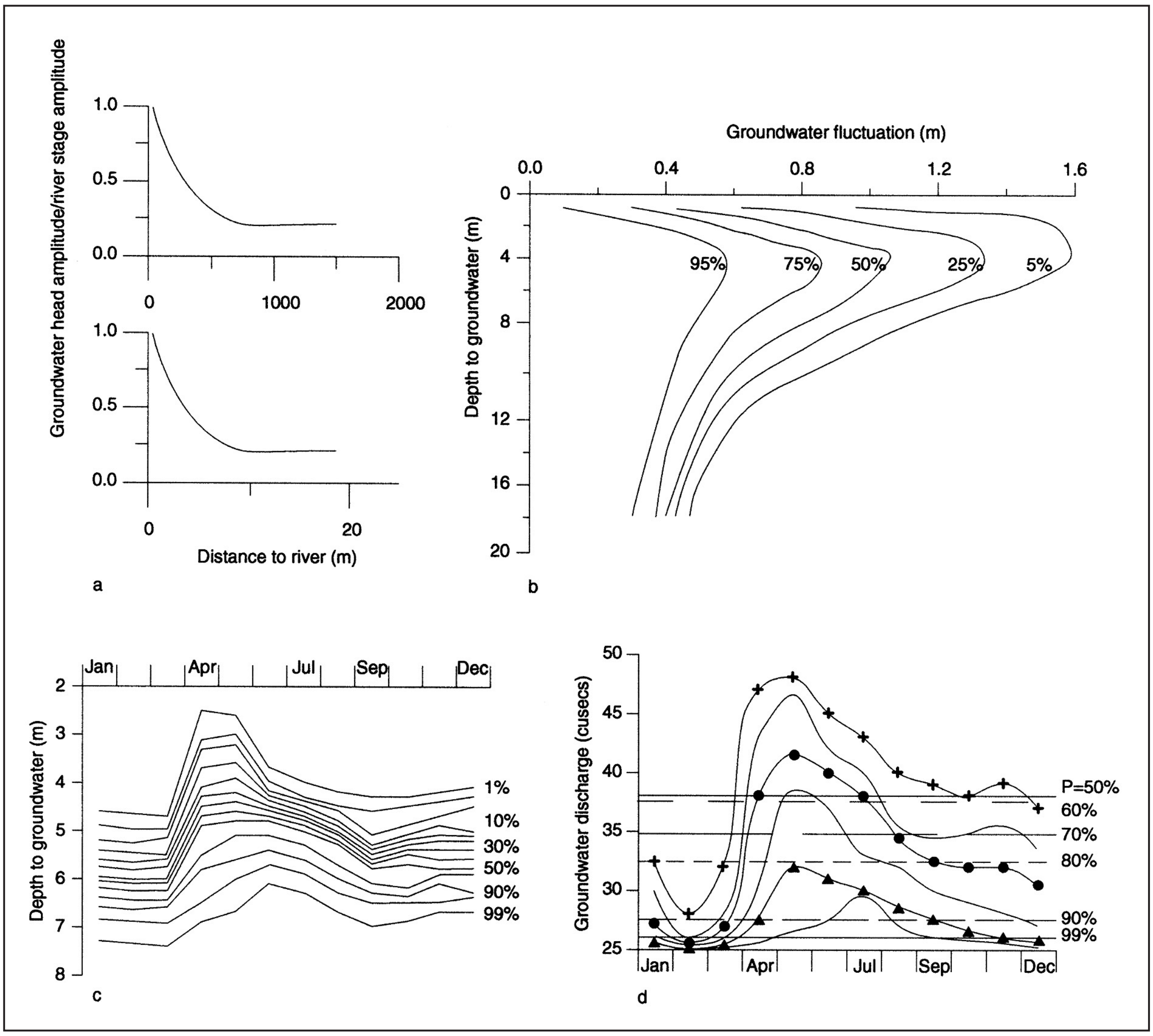
The hydrogeological analogy may be used for initial approximate estimates of the scale of possible changes. Relationships in both time and space can be used by considering the hydrogeological subdivisions of regions by groundwater regime formation conditions (i.e. distinguishing regime types, sub-types, kinds and varieties) it is possible to:

- extend the data on the groundwater regime and forecasts from an area which has been studied to an area which has not been studied,
- evaluate the probabilistic nature of the seasonal and long term variability in groundwater heads, the variation in amplitude of the groundwater heads and
- determine the most probable minimum and maximum groundwater heads, temperatures and chemical composition of the groundwater.

Quantitative relationships between pairs of parameters may be used for approximate forecasts. Typical examples include:

- relationships between the amplitude of the fluctuation of groundwater heads, h_g, and the depth of water in a river at a gauging station, h_r, expressed as a fraction of unity, (Fig. 11.5a);
- relationships between the amplitude of the fluctuations of groundwater heads and the water table position or the character of the inter-annual level regime in years with different water availability (probability) (Fig. 11.5b-c);
- normal annual relations between groundwater flow and precipitation in the form of coefficients, *K*, which equal the percentage ratio of precipitation to groundwater

Figure 11.5 Examples of forecasting: (a) relationships between groundwater levels amplitude/river stage amplitude and the distance to the river, (b) between groundwater level fluctuation amplitude and depth to ground water, (c) graph of the inter-annual groundwater regime, (d) graph of groundwater discharge in years with different probability of water availability



discharge ($K = 100\ P(\text{mm})/Q(\text{mm})$); the relation between the groundwater discharge and the annual water availability (Fig. 11.5d) may be used for predicting the possible temporal variability in groundwater discharge to streams.

Long series of observations can be used as an analogy to obtain statistical parameters for probabilistic estimates for areas with a short series of observations. The similarity of groundwater regimes in areas with short or long series of observations can be proved by the similarity of physical conditions. Consequently the method of hydrogeological analogy can be used for areas with insufficient hydrogeological information, to indicate the nature and scale of the processes.

(ii) Probabilistic statistical methods of forecasting

Probabilistic-statistical methods can be applied to forecast the natural and slightly disturbed groundwater regimes (Brown et al., 1972; Bredehoft, 1982). Probabilistic techniques are required because of the interaction between the groundwater regime and a large number of different but essentially unpredictable factors, which are mostly climatic. For example, the magnitude of the groundwater recharge depends on the magnitude and intensity of the precipitation and also on the air humidity deficit, soil moisture content, freezing of the aeration zone, air temperature, evapotranspiration and, if appropriate, snow melting and thaw periods. It may be impossible to obtain all this information on a regional scale for the period of the forecast; consequently the forecast is based on the most significant factors.

Often it is difficult to find a long series of observations in the region where the forecast is required and therefore a probabilistic statistical technique is used. For forecasting in moderate latitudes it is usually possible to predict mean annual and extreme groundwater levels and discharge rates from:

- depths to the water table at the start of the forecast,
- total mean monthly precipitation for the spring, autumn and winter periods,
- total mean monthly temperatures with an air moisture deficit for the spring, autumn and winter periods,
- duration of thaws or a total positive temperature for the winter thaw period.

The contribution of these factors is different in different regions and for various groundwater level forecasts. It is necessary to identify the important factors for a particular forecast; this can be achieved by correlation techniques and good judgement. More information can be found in Brown et al. (1972).

The longer the period covered by the forecast, the lower the accuracy. The optimum period for forecasts of the minimum levels is 3 to 4 months, and for the maximum spring levels it is 2 to 3 months. As the duration of the forecast increases, the meteorological factors become more important and the role of the preceding history becomes less important. Further information can be found in Kovalevsky (1976).

(iii) Balance methods of forecasting

Balance methods of forecasting are based on the main groundwater components which determine the groundwater head response with time. The groundwater balance equation can be written as

$$\mu \cdot \Delta h = \frac{(Q_1 - Q_2)}{F} + w \cdot \Delta t \qquad (11.2)$$

where

μ = water yield or saturation deficit of the soil or rock (the storage coefficient),
Δh = change in groundwater head in a flow element between wells,
Δt = time interval for the calculation,
Q_1, Q_2 = groundwater inflow and outflow in the flow element,
F = plan area of the flow element,
w = rate of infiltration recharge of groundwater from above.

The water balance equation for the soil-rock prism extending from the top of the confining layer to the water surface is:

$$\Delta C + n \cdot \Delta h = \left[W_a + \frac{(Q_1 - Q_2)}{F} \right] \cdot \Delta t \qquad (11.3)$$

where:

ΔC = the change in moisture reserve in the zone of aeration,
n = total soil moisture capacity,
w_a = water exchange between zone of aeration and atmosphere (evapo-transpiration).

Combining the above equations

$$\Delta h = \left[W_a \cdot \Delta t - w \cdot \Delta t - \Delta C\right] \cdot \; V_{vol} \qquad (11.4)$$

$$\frac{\Delta h \cdot V_{vol}}{W_a \cdot \Delta t} = 1 - \frac{W \cdot \Delta t + \Delta C}{W_a \cdot \Delta t} \qquad (11.5)$$

where V_{vol} (= $n - \mu$) is the soil moisture content within the capillary fringe.

These equations allow a forecast to be made of the rise in the water table, Δh, if data are available concerning the water exchange between the zone of aeration and the atmosphere and the groundwater recharge. Such forecasts are only realistic if there are long series of data for these components. Forecasts based on lumped parameter approaches are discussed by Anderson and Burt (1985).

(iv) Forecasts based on the inertial properties of aquifers

The longer term response of an aquifer to meteorological influences is not direct. Because horizontal groundwater velocities are typically tens of metres per year, the precipitation which infiltrates into the aquifer at some distance from a river may take many years to reach the river. As the groundwater moves towards the river it will receive additional recharge each year hence the river response is the result of precipitation over a significant number of years. This may be the reason for the weak correlation between precipitation in one year, and groundwater heads or flows.

The capability of an aquifer system to integrate the effects of the precipitation over a number of years and the degree of the inertia of the system is unique because it depends on the seepage properties of the water-bearing rocks and the drainage area. This groundwater inertia may be assessed by the correlation between groundwater heads and the summed effect of the precipitation in preceding years.

It is advisable to use as an annual precipitation the total effective precipitation (the precipitation for the cold season being the main source of groundwater recharge). The extremes of the curve of these correlations gives the optimal number of years whose precipitation should be taken into consideration in forecasting. This procedure is necessary when it is not possible to use the values of preceding groundwater levels as one of the factors in which the precipitation of the preceding stage is taken into account. Such variants emerge, for example, when forecasting the groundwater regime and resources for the remote future, using various climate change scenarios.

Although the integrating capability of aquifers to reflect the precipitation for the preceding years may be large, the greatest contribution may be due to the recent history. All this predetermines the possibility of using as the method of forecasting the extrapolation of regular features established for the period of earlier observations, preserving the former conditions of the groundwater regime for the period of forecasting.

Methods based essentially on the inertial properties of the aquifer include the following:

- identifying and extrapolating linear and non-linear trends,
- distinguishing significant harmonics in the structure of the series and extrapolating future groundwater head and discharge variations as a total of these harmonics,
- forecasting groundwater heads as a Markovian process.

Strictly speaking, forecasts based on these methods do not require a knowledge of the physics of the processes, since they are based on properties of the time series. However, the reliability of the forecast depends on the extent to which the trends can be substantiated physically. Linear and non-linear trends in groundwater head and discharge variations may be caused by human activities such as reclamation, urbanisation, construction of hydraulic structures, groundwater abstraction, technological changes, or by natural climate changes. Linear or non-linear trends should be identified and the statistical reliability of the trends should be considered.

(v) Probabilistic forecast estimates

Probabilistic forecasts of the groundwater regime may be made if calendar (time-based) forecasting is impossible or inadvisable. Probabilistic forecasts allow characterisation of extreme, mean or other values with given probabilities, without an indication of the time of their onset. Forecasts based on probability concepts depend on:

- statistical analysis of long term groundwater observations,
- computation of statistical properties of groundwater heads and discharge,
- finding the law of groundwater head and discharge distributions,
- using empirical and theoretically calculated probability graphs,
- groundwater head and discharge values of 50% or 95% or other recurrence intervals can be deduced from the probability graphs.

It is essential to estimate the possible errors in the computed values. The accuracy of all statistical estimates depends on the length of the series. Higher accuracies result from longer series. If the series are short, they should be extended using standard techniques or using series analogues.

11.4.3 The use of models for predictions

The reason behind the development of a groundwater model is usually to obtain a means of predicting the effect of exploiting an aquifer or modifying the exploitation to minimise the effects of falling water tables. At the outset it must be stressed that the use of a model for predictive purposes does not mean that precise forecasts will be made of the groundwater conditions at a particular time in the future; the variability and unpredictability of the recharge usually prevents precise forecasts. Nevertheless it is possible to consider the likely consequences of any changes; the findings will be in terms of likely changes in groundwater levels, river flows and changes in the quality of water. There are a number of important issues to be considered when predicting future aquifer responses:

- the purpose of the prediction may be to identify the *safe yield* of an aquifer; this can be defined in terms of the *dynamic storage*, whereby the volume of water which can be removed equals the volume of stored water which can be replenished by the end of the annual recharge season. If the annual recharge shows significant variations it may be acceptable to make use of some of the *static storage* to smooth out the fluctuations in recharge. The general principle behind the safe yield concept is that, in the long term, the withdrawal from the aquifer should be such that a long term water balance may be maintained.
- an alternative approach involves the *mining* of groundwater. In deep alluvial and sandstone aquifers, the dynamic storage may be less than 1% of the total storage, consequently some mining of the groundwater reserves may be acceptable. In semi-arid areas with deep aquifers where groundwater is used for irrigation, mining of groundwater is almost certain to occur. However, as the mining of groundwater proceeds, the water table falls, the cost of lifting the groundwater from depth increases and the efficiency of the wells and boreholes is likely to decrease so much that shallow wells are abandoned and the drilling of new deep boreholes becomes necessary. These severe consequences of the mining of groundwater must be considered.

- there are also *environmental consequences* of changing the use of an aquifer. Examples include the settlement of land due to the mining of groundwater (Ramnarong, 1983) and the drying up of rivers and lakes. On the other hand, excess recharge due to irrigation or urban development can lead to waterlogging; typical examples are the waterlogging in the Indus Basin or in cities such as Riyadh (Rushton and Al-Othman, 1994). A further environmental consequence can be a deterioration in the quality of pumped water due to recent or trapped saline water being attracted to the pumped boreholes.

11.4.4 Methodologies in the use of models for predictive purposes

This section outlines certain of the issues which need to be resolved when using a model for predictive purposes.

What are the inputs? Recharge to the aquifer has to be specified and it is advisable to use historical rainfall and evapotranspiration values so that the forecasting can show how aquifer conditions would differ under alternative abstraction regimes. If the purpose of the study is to predict the effect of climate change, then rainfall and evapotranspiration modified according to the predicted climate changes should be used in the recharge estimation.

Do conditions in the aquifer system change? Care must be taken when the model is used for conditions other than those for which the model was tested New mechanisms can become important, as demonstrated by the case studies discussed in this chapter. Typical changes in conditions include reductions in transmissivities, dewatering of fissures, changes between confined and unconfined conditions, modifications to the recharge mechanisms, an alteration to the efficiency of boreholes, or changes in the nature of rivers or streams.

What are the starting conditions? For certain situations it is acceptable to use the same starting conditions as for a historical run. However, if mining of groundwater occurs with a steady decrease in the water table elevation, the simulation should start from the current conditions, to predict future trends. Unsatisfactory starting conditions are the cause of many misleading predictive runs.

What are the outputs? Often it is helpful to plot differences between the predicted results with changed conditions and a baseline which may be a historical run or a predictive run using the current abstraction policies. modifications to the flow balance within the aquifer system also provide useful insights.

For how many years should predictive simulations continue? For aquifers with large volumes of stored water, changes in the conditions within the aquifer can be slow, often taking decades. consequently, it is advisable to run predictive simulations for very long time periods until a new equilibrium is approached.

11.4.5 Examples of predictions

Each of the models developed for the case studies has been used for forecasting.

(i) Weathered-fractured aquifer

The model described in Section 11.3.4 was developed to represent a pumping test with 7.5 hours of pumping and the subsequent recovery. This model can also be used to obtain estimates of the likely responses due to long term pumping. However, because the predictive runs are far longer than the historical testing, the results should be used with caution. In predicting likely responses, three situations have been considered to determine how the aquifer can be exploited safely; the aquifer properties are recorded in Section 11.3.4.

(a) a large-diameter well in the weathered zone with a diameter of 5.8 m and a depth of water at the start of the irrigation season of 8.0 m.

(b) a tube well with a diameter of 0.15 m tapping only the fractured zone with a submersible pump, thereby allowing large pumped drawdown,

(c) a dug-cum-bore well in which a bore is drilled into the fractured zone through the base of the large diameter well of example (a); the pumped drawdown is restricted since there must always be sufficient water in the bottom of the large diameter well to provide suction for the surface pump.

The effective spacing between the wells is set at 400 m; consequently the area of aquifer from which any of the structures can draw water is 160,000 m^2. The aim is to pump water for 100 days to maximise the irrigation potential. The results are as follows:

(a) The yield of the large-diameter well is poor; if the duration of pumping is 7.5 hours, the pumping rate must not exceed 150 m^3/d if the abstraction rate is to be maintained for the full 100 days of the growing season.

(b) If a submersible pump is used with the borehole taking water from the fractured zone, the yield is greatly increased; in fact a discharge rate of 350 m^3/d can be maintained without difficulty for 15 hours per day for 100 days. The maximum pumped drawdown would be 13.5 m but significant drawdown would also occur in the weathered zone. In fact the weathered zone would almost be dewatered, with drawdown greater than 9.3 m. This would mean that shallow dug wells in the weathered zone would become dry. Another serious consequence is that the recharge needed to refill the weathered aquifer during the next monsoon rains would be 94 mm; this intensity of recharge occurs only in the wettest years. Prediction (b) represents precisely what happened when tube wells were drilled in the fractured zone of the Dulapally aquifer. Because so much water was pumped from the aquifer and because the spacing between the tube wells was often less than 400 m, the tube wells failed after one or two years and also the weathered zone was dewatered, causing many of the large diameter wells in the weathered zone to fail too. The problem is that the submersible pumps in the tubewells are too powerful at removing water from the aquifer.

(c) To use the aquifer efficiently, the ability of the fractured zone to transmit water should be used but the drawdown should be limited; this can be achieved using a dug-cum-bore well. When the abstraction is 350 m^3/d for 7.5 hours, the maximum drawdown in the well is 6.6 m, which is within the capability of a suction pump. The maximum drawdown in the aquifer is 5.0 m close to the well; with this order of drawdown some deepening of domestic dug wells would be necessary. In most years the recharge would reach the 47 mm required to ensure that levels recover following the monsoons. Consequently a dug-cum-bore well with a surface pump should be used with this form of layered aquifer.

(ii) Alluvial aquifer

Extensive predictive runs have been carried out in the Mehsana alluvial aquifer; all have shown that a significant reduction in the volume of water pumped from the deep aquifers is required to stabilise the drawdown.

One particular concern is that, as the water table continues to fall, it moves into the zones with a higher clay content having a lower specific yield. This condition of a decreasing specific yield can be included directly in the model by modifying the specific yield from the standard value of 0.08 to 0.01 when the water table in the overlying aquifer falls more than 50 m below ground level. After six years the predicted groundwater heads in the main aquifer fall to 110 m below the ground level; with this substantial decline in heads there would be a serious risk that unconfined conditions would develop in the deeper aquifers.

(iii) Limestone aquifer

Canal irrigation has extended to an area close to the Miliolite Limestone aquifer; one predictive scenario is to use the surplus canal water during the monsoon season for artificial recharge, both to improve the yields of the existing wells and to flush out the saline water in the coastal zone. Field experiments demonstrated that the wells are very efficient at accepting the recharged water but the trial did not continue for long enough for the overall success of the scheme to be proved.

The likely outcome of implementing an artificial recharge scheme was explored using the mathematical model which had been verified against the historical field response. When artificial recharge was simulated in the model:

- the first benefactors were the farmers with wells downstream from the recharge sites; the yield of their wells increased significantly,
- farmers with wells upstream or at some distance from the artificial recharge sites would receive little benefit from the recharged water,
- the next stage is that much of the recharged water would escape from the springs,
- there would be no reduction in the salinity of the groundwater within 2 km of the coast; this saline water was drawn in by pumping and will not be flushed out by the artificially recharged water which takes easy routes out of the aquifer rather than moving to the coastal area to force out the poor quality water.

11.5 Concluding remarks

Extensive experience has been gained in the assessment of groundwater resources and the prediction of likely future responses. The progress in recent years has been due to improved field techniques combined with developments in groundwater modelling. The key to reliable assessments and predictions is the imaginative use of the wide variety of hydrogeological techniques combined with the skilful use of groundwater models. Groundwater assessment and prediction requires a team effort in which skills and techniques are combined to lead to an understanding of the flow processes within the aquifer system.

11.6 References and additional reading

ANDERSON, M. G.; BURT, T. P., 1985. *Hydrological Forecasting*. Wiley, New York.

ANDERSON, M. P.; WOESSNER, W. W., 1991. *Applied Groundwater Modelling*. Academic Press, 381 pp.

ANDERSON, P. F.; MERCER, J. W.; WHITE, H. O., 1988. Numerical Modelling of Salt Water Intrusion at Hallandale, Florida. *Groundwater,* Vol. 26, pp. 619–30.

BALEK, J., 1989, *Groundwater Resources Assessment.* Developments in Water Science 38, Elsevier, 249 pp.

BEAR, J., 1979. *Hydraulics of Groundwater*. McGraw-Hill, 567 pp.

BOONSTRA, J.; DE RIDDER, N. A., 1981. *Numerical Modelling of Groundwater Basins; A User-Oriented Manual*. ILRI Publ. 29, Wageningen, 226 pp.

BOUWER, H., 1978. *Groundwater Hydrology*, McGraw-Hill, 480 pp.

BREDEHOFT, I. D. (ed.), 1982. *Groundwater Models,* Vol. 1. UNESCO, Paris.

BROWN, R. H.; INESON, J.; KONOPLYANTSEV, A. A.; KOVALEVSKY, V. S., (eds.), 1972, *Groundwater Studies*, UNESCO, Paris.

CONNORTON, B. J., 1985. Does the Regional Groundwater-Flow Equation Model Vertical Flow? *J. Hydrol.*, Vol. 79, pp. 279–99.

DE MARSILY, G., 1986, *Quantitative Hydrogeology*, Academic Press, 440 pp.

DRISCOLL, F. G., 1986. *Groundwater and Wells, Second edition*, Johnson Division, St. Paul, Minn., 1089 pp.

FREEZE, R. A.; CHERRY, J. A., 1979. *Groundwater*, Prentice-Hall Inc., 604 pp.

KAVALANEKAR, N. B.; SHARMA, S. C.; RUSHTON, K. R., 1992. Over-Exploitation of an Alluvial Aquifer in Gujarat, India. *Hydrol. Sci. J.*, Vol. 37, pp. 329–46.

KINZELBACH, W., 1986. *Groundwater Modelling*. Developments in Water Science 25, Elsevier, 333 pp.

KOVALEVSKY, V. S. 1976. *Long Term Fluctuations in Groundwater Levels and Discharge.* Nauka, Moscow. (In Russian.)

LERNER, D. N.; ISSAR, A. S.; SIMMERS, I., 1990. Groundwater Recharge, *International Contributions to Hydrogeology,* IAH, Vol. 8, 345 pp.

LIGGETT, J. A.; LIU, P. L., 1983. *The Boundary Integral Equation Method for Porous Media Flow*. Allen and Unwin, London. 255 pp.

MARINO, M. A; LUTHIN, J. N., 1982. *Seepage and Groundwater*. Elsevier, 489 pp.

MCWHORTER, D. B.; SUNADA, D. K., 1977. *Groundwater Hydrology and Hydraulics*. Water Resources Publications, 290 pp.

PINDER, G. F.; GRAY, W. G., 1977. *Finite Element Simulation in Surface and Sub-Surface Flow*. Academic Press, New York.

RAMNARONG, V., 1983. Environmental Impacts of Heavy Groundwater Development in Bangkok, Thailand. *Intern. Conf. on Groundwater and Man,* Australian Government Publishing Service, Canberra, Vol. 2, pp. 345–50.

RATHOD, K. S.; RUSHTON, K. R., 1991. Interpretation of Pumping from a Two-Zone Layered Aquifer Using a Numerical Model. *Groundwater,* Vol. 29, pp. 499–509.

RUSHTON K. R.; AL-OTHMAN, A. A. R., 1994. Control of Rising Groundwater Levels in Riyadh, Saudi Arabia. In: *Groundwater Problems in Urban Areas,* Thomas Telford, London, pp.299–309.

RUSHTON, K. R.; RATHOD, K. S., 1985. Horizontal and Vertical Flow Components Deduced from Groundwater Heads. *J. Hydrol.*, Vol. 79, pp. 261–78.

RUSHTON, K. R.; TOMLINSON, L. M., 1979. Possible Mechanisms for Leakage between Aquifers and Rivers. *J. Hydrol.*, Vol. 40, pp. 49–65.

RUSHTON, K. R.; WELLER, J., 1985. Response to Pumping of a Weathered-Fractured Granitic Aquifer. *J. Hydrol.*, Vol. 80, pp. 299–309.

RUSHTON, K. R.; RAGHAVA RAO, S. V., 1988. Groundwater Flow through a Miliolite Limestone Aquifer, *Hydrol. Sci. J.*, Vol. 33, pp. 449–64.

RUSHTON, K. R.; REDSHAW, S. C., 1979. *Seepage and Groundwater Flow*. Wiley, Chisester, 332 pp.

RUSHTON, K. R.; SRIVASTAVA, N. K., 1988. Interpreting Injection Well Tests in an Alluvial Aquifer. *J. Hydrol.*, Vol. 99, pp. 49–60.

RUSHTON, K. R.; TIWARI, S. C., 1989. Mathematical Modelling of a Multi-Layered Alluvial Aquifer. *J. Inst. Eng. India,* Vol. 70, part CV2, pp. 47–56.

RUSHTON, K. R.; WELLER, J., 1985. Response to Pumping in a Weathered-Fractured Granitic Aquifer. *J. Hydrol.*, Vol. 80, pp. 299–309.

SIMMERS, I. (ed.), 1997. Recharge of Phreatic Aquifers in (Semi-)Arid Areas. *Intern. Contrib. Hydrogeol.*, IAH, Vol. 19, Balkema, Rotterdam.

TODD, D. K., 1980. *Groundwater Hydrology*. John Wiley and Sons, New York, 535 pp.

WALTON, W. C., 1991. *Principles of Groundwater Engineering*. Lewis Publishers, Michigan, 546 pp.

WANG, H. F.; ANDERSON, M. P., 1982. *Introduction to Groundwater Modelling, Finite Difference and Finite Element Methods*. W. H. Freeman & Co., San Francisco, 237 pp.

WARD, D. S.; BUSS, D. R.; MERCER, J. W.; HUGHES, S. S., 1987. Evaluation of Groundwater Corrective Action of the Chem-Dyne Hazardous Waste Site Using a Telescopic Mesh Refinement Modelling Approach. *Water Resources Research,* Vol. 23, pp. 603–17.

ZEKTZER, I. S. (ed.), 1983. *Principles of Hydrogeology, Vol. 2, Hydrogeodynamics.* Nauka, Novosibirsk. (In Russian.)

12 Groundwater management

12.1 Introduction

12.1.1 Scope

As with any natural resource, the *management* of groundwater involves reconciling a supply-based approach (managing the resource) with a demand-based approach (managing exploitation and utilisation), in the light of the objectives of the management actors whose decisions are based on a number of criteria subjected to various constraints and applicable on the basis of different methods. The following brief survey of groundwater management will therefore look at: the concept of the resource to be managed, the physical conditions and exploitation strategies for which a strategy is to be defined, and the socio-economic conditions. The latter are the managers and their objectives, management constraints and criteria, management decision methods and aids, and management tools.

12.1.2 Groundwater resource management – preliminary remarks

As is the case with water resources in general, the groundwater resource is a *combined physical and economic concept:* it is water found underground which is seen from a utilitarian point of view in terms of the possibility and benefit of drawing on it. It constitutes a natural 'supply' to meet the demand for water.

The groundwater resource is a *multidimensional concept:* it is defined by its location, its occurrence over time, its size, properties, conditions of accessibility, the effort required to mobilise it and, therefore, its cost, all of which are also to be considered in the context of demand. The resource is thus assessed using a variety of criteria, and it would be simplistic to measure it solely in quantitative terms on the basis of the flow. In terms of user criteria, the groundwater resource – as a source of water supply competing with others (surface water, in particular) – has different advantages as well as some disadvantages and limitations, which are summarised in 12.1.

Table 12.1 Advantages and disadvantages of groundwater as a source of supply, compared to the use of surface water

Spatial distribution	An extensive resource facilitating catchment near points of use, thus minimising conveyance costs, but to satisfy strong demand a number (how many depends on output) of catchment areas will be required
Availability over time	A permanent resource with steady discharge volume, less affected by the weather than surface water, and therefore affording greater security of supply. Natural storage which does not require regulation works
Resource evaluation	An invisible resource whose precise evaluation may require the use of rather sophisticated methods which are more expensive but also quicker than those used for surface water resources
Natural properties	Constant or little changing properties making it easy to correct certain deficiencies for certain uses (potability, process water): hardness, Fe and Mn content. Increased salinity at medium and great depth
Vulnerability to pollution	Deep-lying water is generally free of risks, except those posed by faulty well drilling and deliberate injection. Underground 'phreatic' water is more indirectly exposed to the risks of diffused or localised pollution than surface water, but suffers from more lasting effects (being less resilient)
Production costs	Investment and operating costs are lower on average than for surface water, but fairly diversified, depending on the local characteristics of the aquifer (depth, capacity). Operating costs reflect variations in energy costs (except in the case of gravity catchment). Decreasing yield with increase in the overall rate of exploitation of a given aquifer
Operational flexibility	Possibility to install or close at any time additional production wells or water abstraction equipment, and therefore better adapted to react to changing demand than surface hydraulic development structures; investments can be spread in time and returns will come quicker

The groundwater resource must be described, evaluated and managed within precisely defined physical frameworks: the *aquifer systems* (see also Chapter 10), the natural units of groundwater management whose extent, complexity and behaviour condition the exploitation, conservation and management of the resources.

Because of the continuity that exists between the aquifers and the surface water systems of a catchment basin, and because of the possible interactions between groundwater exploitation and surface water management, the groundwater resources are not independent from the surface water resources in terms of their evaluation or their management. Hence the need to *integrate* groundwater management into overall water management. This integration does not exclude the profitable utilisation of distinctive groundwater characteristics which often complement those of surface water.

The groundwater resource, described as being a 'supply model', can rarely be evaluated in terms of units (average producible quantities, costs, external effects, etc.). Its evaluation cannot be dissociated from *predictive management* and the assessment of the feasibility of different exploitation scenarios, taking different constraints and user criteria into account.

Unlike other underground resources, groundwater is not a mineral 'raw material' like any other. It is mobile and, therefore, should be considered not in terms of legal ownership but rather of exploitation rights. As a result of the transmission of effects in the aquifers, the impact of actions on groundwater (withdrawals, excess of inflow, pollution) extends as far as the physical boundaries of the aquifer system. The management of a unit of groundwater resources should, therefore, be community based, involving all the stakeholders concerned.

12.2 Physical conditions of groundwater management

12.2.1 Flow management and storage management

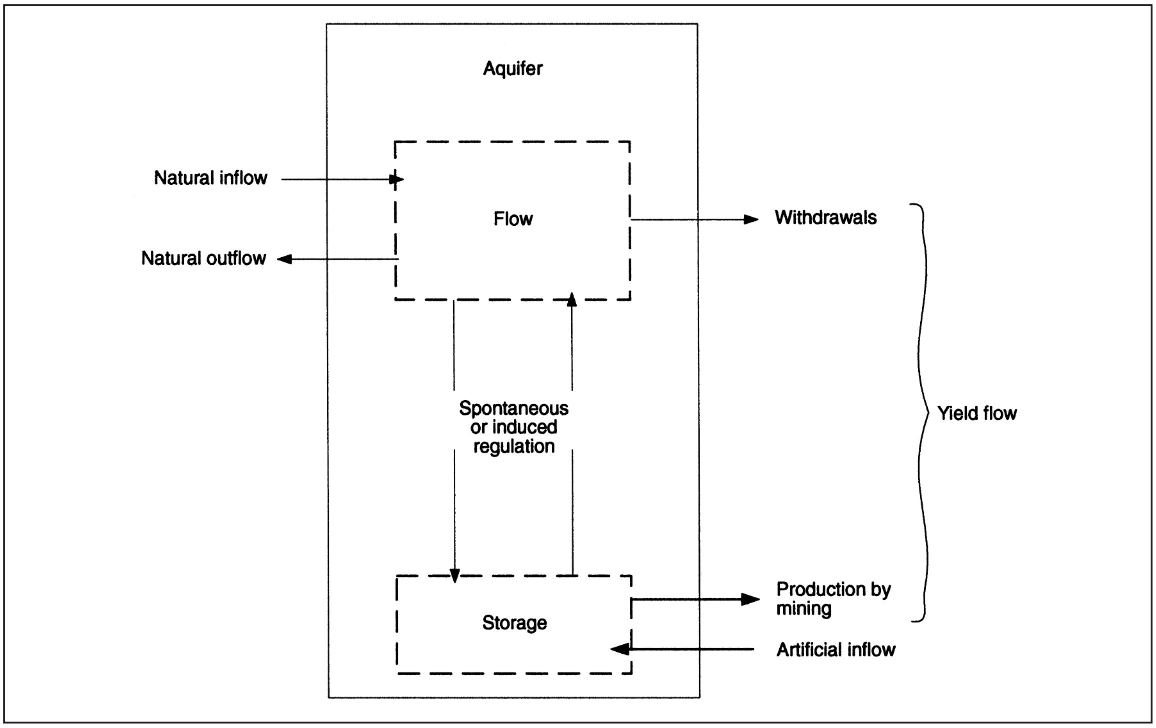
As an aquifer system combines the functions of conveyance and storage, its management should concern both flow and storage, but it can combine the two in very varying degrees.

In general, an aquifer system provides a renewable resource managed as a flow that is regulated to a greater or lesser degree by storage variations. The possibility of amplifying these variations provides a degree of freedom of action in relation to the aquifer. It makes the strategy of withdrawals more or less independent of the natural inflow, i.e. more regular or irregular than the latter – on a perennial basis as well – depending on variation in demand. In certain cases, the aquifer storage itself may become a non-renewable resource, which may be mined.

The respective and more or less combined roles of flow and storage in groundwater resource management (quantitatively speaking) can be expressed by the diagram of Figure 12.1.

The effect of an aquifer's storage and the variation in storage on the rate of water production is therefore of prime importance in the definition of the actual or planned exploitation strategy.

Figure 12.1 The roles of flow and storage in groundwater management



12.2.2 Overall exploitation strategies of an aquifer

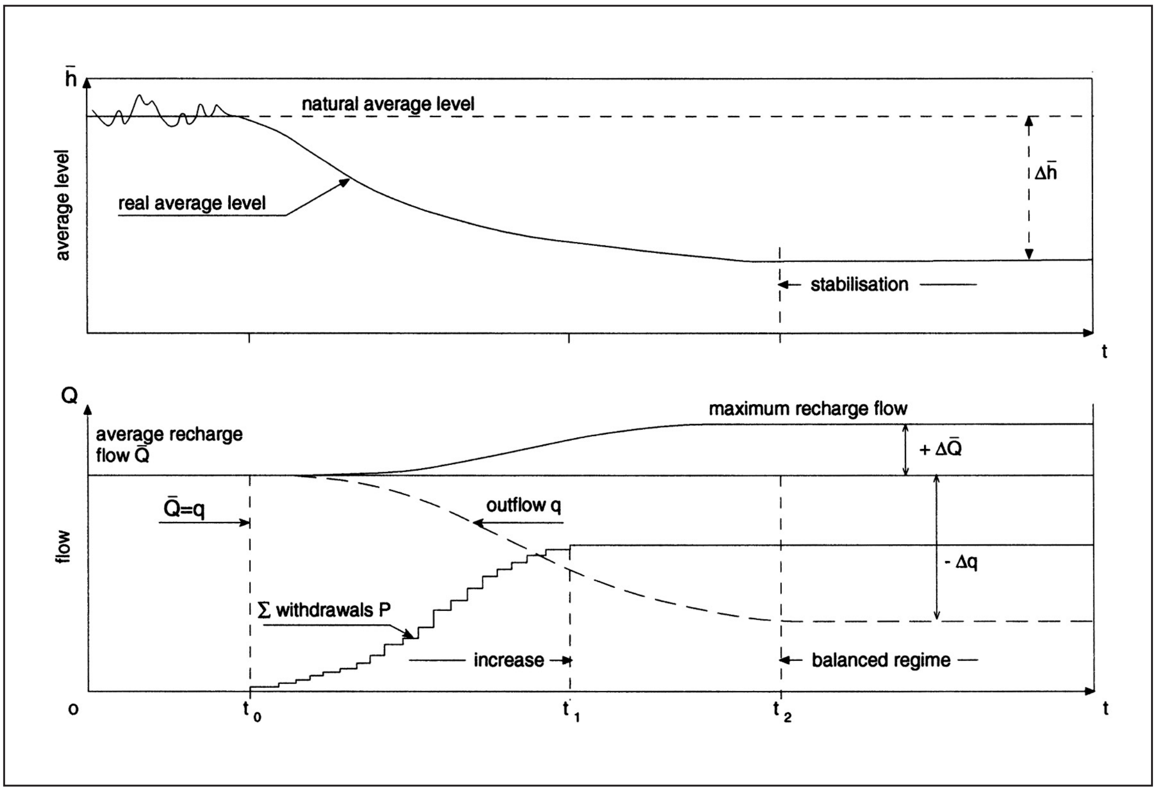
Depending on the degree and reference period of the dynamic balance established between an aquifer's natural flows and the flow imposed by exploitation – depending, therefore, on the role played by storage variation – three types of exploitation strategy are conceivable and practicable, albeit under certain constraints.

(i) *A dynamic balance strategy*

A dynamic balance strategy, may be 'imposed' by an average withdrawal rate that is less than or equal to the average inflow (possibly increased by boundary effects), without prejudice to possible seasonal or even interannual variations (e.g., increase during periods of drought).

The storage decreases during an initial phase of imbalance, but then stabilises and assumes a regulatory role (on an annual or perennial basis). The natural regulatory function may be subject to the requirement that a minimum flow rate is maintained at the discharge boundaries (e.g., spring, flow or low-water level of main stream) or the need to preserve the freshwater/saltwater balance in a coastal aquifer (Fig. 12.2).

Figure 12.2 **Exploitation under a dynamic balance strategy after a necessary transitional phase: diagram of changing average levels – global indicator of the state of the storage – and flows. P is less than Q. The stabilisation of levels (on average) is deferred (in t_2) in relation to that of withdrawals (in t_1), After a withdrawal phase from t_0 to t_2, there is a time lag until the levels stabilise; the effect of the withdrawals in t_1 results in the levels at t_2**

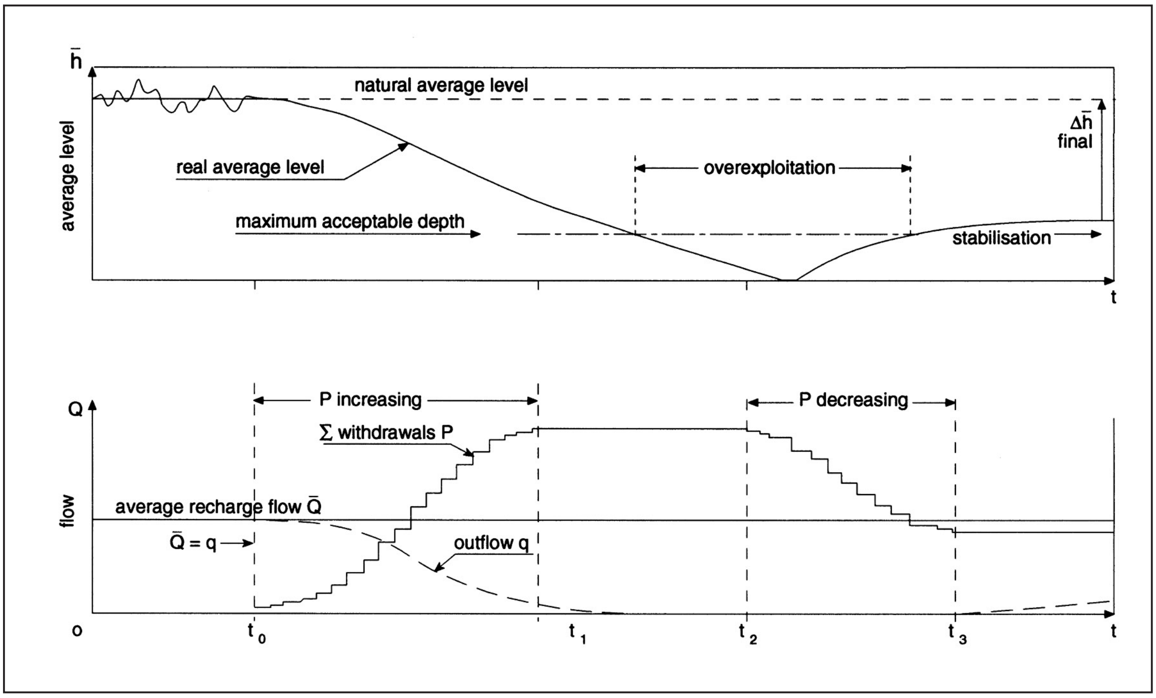


(ii) *A temporary non-equilibrium strategy*

A temporary non-equilibrium strategy, whether 'managed' or unavoidable, has an average withdrawal rate higher than the average inflow (even if augmented by boundary effects), irrespective of whether the withdrawals are increasing or stabilised. There is a time lag before equilibrium is restored.

Withdrawals from storage make a large, and sometimes dominant, contribution to water production, for a period of variable length which is limited either by external constraints (strategy 1) – or as a result of the reduced capacity of the wells (e.g., drawdowns which are excessive or restricted by the depth of the aquifer's substratum). In a later phase of possible restoration of equilibrium – involving a reduction in withdrawals or sometimes stimulated by an artificial or induced inflow increase – groundwater storage can be partially recovered and stabilised (on average), as in strategy 1 (Figure 12.3).

Figure 12.3 Exploitation under a prolonged imbalance strategy followed by a return to balance by reducing withdrawals: diagram of changes in average level and flows, with P > Q. Imbalance strategy occurs from t_0 to t_3 during the phases of growth, followed by stability then decrease in withdrawals P, with dewatering from t_0 to t_2, despite the almost total reduction in residual outflow q. There is a tendency towards restoration of balance in t_3, with partial reconstitution of groundwater storage, withdrawals now being at a rate close to the initial recharge flow. Overexploitation would consist in a lowering of levels below the maximum acceptable depth

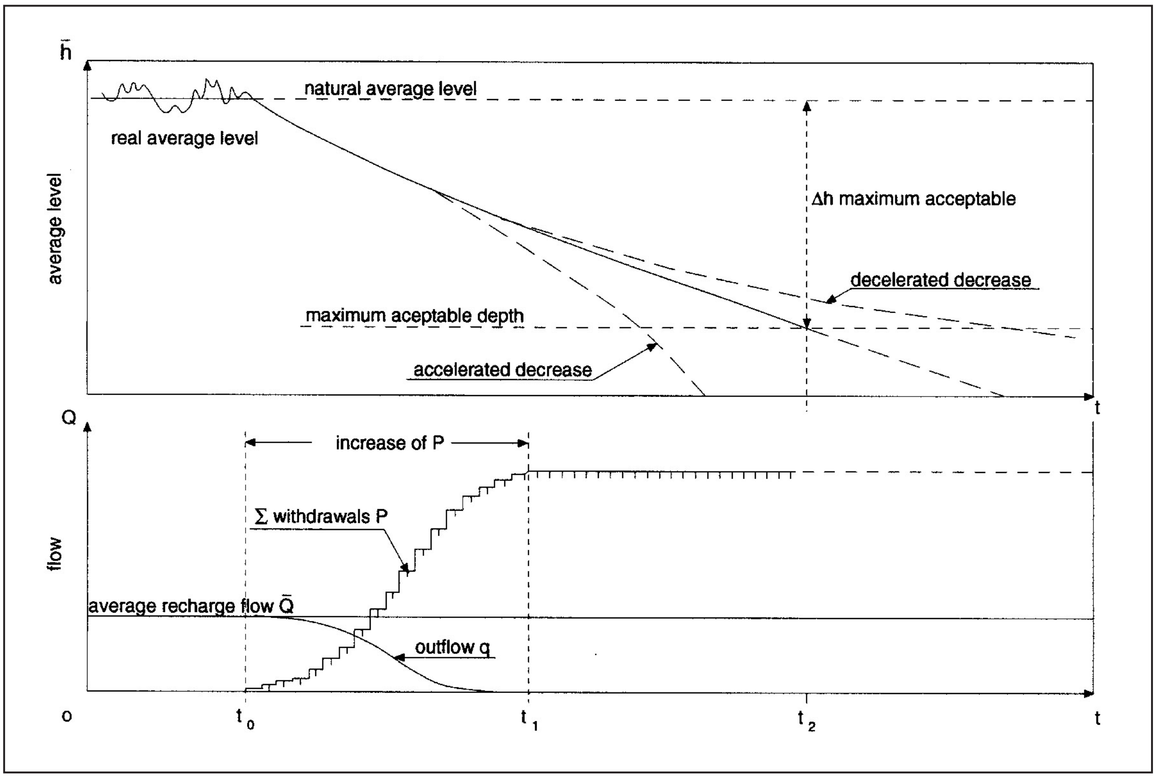


(iii) Depletion or permanent non-equilibrium strategy

Depletion or permanent imbalance strategy, that is the mining of groundwater at a withdrawal rate that may or may not increase, and that from early on is above – often well above – the average inflow rate.

Withdrawals from storage provide most of the water produced. In the long-term, the exploitation is limited when the drawdown becomes excessive, but equilibrium is not regained returning to a balanced strategy, because the recharge of the groundwater storage is too slow and is sometimes hindered by the irreversible degradation of storage capacity trough compaction of the aquifer as a result of the pressure decrease (Fig. 12.4).

Figure 12.4 **'Mining' of stored groundwater storage, under a depletion strategy; diagram of changes in levels and flows (P > Q). The time limit of the exploitation, depends on the rate of withdrawal P and the characteristics of changes in the levels. It is reached when the drawdown levels reach the maximum depth which is practicable (given the position of the boundary of the aquifer), or acceptable from the point of view of pumping costs**



(iv) *Overexploitation: surplus or excessive exploitation?*

Rather than the neutral sense of temporarily excessive withdrawals from the average natural flow of an aquifer, which is covered by the hydraulic concept of imbalance, overexploitation is more usually defined in a derogatory way as excessive exploitation, a management error or failing which has to be prevented or stopped. It is the kind of exploitation which has harmful consequences, either for the exploiters themselves or for third parties (disregard of internal or external constraints), whether or not the exploitation strategy is balanced in the shorter or longer term. Quantitatively speaking, it is a kind of exploitation which can hinder the long-term reproduction of the resource and exceed the limits of resource renewal to the detriment of future yields, without this being an intentional strategy. This type of exploitation can also lead to unacceptable production cost overruns and quality degradation.

The term overexploitation should not be applied neither to temporary large-scale withdrawals from an aquifer with an irregularly recharged aquifer that make judicious use of the regulatory capacity of its storage. Nor should it be applied to the deliberate 'mining' of an aquifer not receiving recharge. Whether the term overexploitation has a negative connotation always depends on the management objective for an aquifer and the relevant exploitation strategy. Hence an exploitation strategy characterised by temporary or permanent imbalance should only be described as 'overexploitation' if the imbalance is unintentional and has harmful consequences, and is therefore contrary to the chosen strategy.

12.2.3 Types of aquifer systems and management conditions

Of course, individual exploitation strategies cannot be indiscriminately applied to any given type of aquifer. Whatever the development objective may be, each particular type of aquifer requires a particular exploitation strategy (Table 12.2).

The variety of hydrogeological conditions and the resulting size, structure and complexity of aquifer systems implies that there is a matching variety of physical management conditions, as well as approaches to groundwater resources. This diversity can, however, be narrowed down to the most common types, presented in Table 12.2. The two most important defining characteristics for this 'typology' are: the flow storage ratio and aquifer/surface stream relationships. Both can show very wide variations:

- The ratio of average flow and average storage, in other words the capacity of the storage to regulate the overall discharge rate of an irregularly recharged aquifer, expressed in terms of duration of overall replenishment, varies from less than one year for thin and shallow alluvial aquifers, to the order of 10^4 to 10^5 years for deep-lying aquifers containing what is often termed 'fossil' water.
- The aquifer/river(s) relationship ranges from a very strong, continuous and permanent connection (thin alluvial aquifers that have an unclogged connection with the river), to complete independence (deep-lying confined aquifers and certain coastal aquifers).

12.3 Socio-economic conditions – management actors and objectives

12.3.1 Management actors

Depending on an aquifer's size and accessibility, the usually numerous economic actors may have the practical and financial means as well as the legal right – land-owners' rights – to exploit the aquifer and possibly find it to be a profitable source of water supply. They are either direct users (households, farmers, industries, etc.) or producers and distributors of water as a commercial economic product (the supply of public drinking water). In addition to the groundwater operators, there are many more actors who can directly or indirectly influence the aquifers exploitation strategy or its properties. They comprise:

- occupants of (i) the land containing the aquifer whose impacts stem from their effluents and their use of the land, e.g., agriculture with the use of fertilisers, waste disposal), or (ii) the users of the underground, e.g., quarries of sand and gravel deposits, mining, underground storage.
- engineers who dam, gauge, embank or otherwise influence the streams linked to the aquifer.

A typology of these different people can be based on whether they are users and non-users of the groundwater, exploiters or developers, or both combined (Table 12.3). The analysis of the 'system of economic actors' involved, of their individual objectives, modes of action and actual or potential conflicts of interest is a preliminary exercise which is as important to management as the analysis of the aquifer system itself.

12.3.2 Management levels

The fact of using or influencing one and the same aquifer unites numerous actors even though their various acts of exploitation or activities impacting on the water strategy or properties have different individual motivations. Usually, the actors are not aware of participating in the *de facto* management of a common property, and when mutually harmful influences are felt, 'user

Table 12.2 Types of aquifer system and management conditions

Type of aquifer system	*Relationship with surface water*	*Sensitivity*	*Appropriate exploitation strategy*	*Restrictive factors*
Alluvial system (often unconfined) Limited size and storage Weak regulatory capacity	Strong relationship Aquifer/river exchanges possible in both directions	Sensitivity to regulation of streams and land occupation Vulnerable to pollution from the surface and streams	Strategy (1) Short-term balance. Over-balance possible from river-induced recharge Flow management common	External constraints: clogging of banks, conservation of minimal flow from springs and in rivers
Phreatic system of plains and plateaux and karst systems Significant storage(storage often > 10 annual flows) with good regulatory capacity Strong hydro-dynamic inertia	Often one-way connection: river drainage (except for bank storage and river loss in karst areas) Unequal density depending on hydrographic network	Variable recharge: Sensitive to drought and to land occupation Vulnerable to pollution, especially when diffuse	Strategies (1) and (2) Perennial balance. Management of the flow and regulating storage	External constraints: - conservation of spring discharges and minimal stream flow
Multi-layered system with interconnected phreatic and semi-confined aquifers Significant storage (Storage > 100 to 1,000 annual flows) partial regulatory capacity Variable inertia	Unequal connection restricted to shallow unconfined aquifers	Sensitive to land occupation (especially the shallow unconfined aquifers) Varying vulnerability to pollution. Exposed to risks of exchange between aquifers with water of unequal quality resulting of drilling	Strategy (2) Final equilibrium possible after prolonged dewatering phase. Management of flow and storage moving towards flow management	External constraints (for near surface unconfined aquifers): - conservation of spring discharges and minimal stream flow Internal constraints (for confined aquifers): - maximum draw-downs acceptable
Deep-lying confined aquifer Large storage (> 1,000 - 10,000 or more annual recharge) but no regulatory capacity It cannot be mobilised without dewatering Low inertia	Independence (except in case of discharge by an artesian spring)	Insensitive to climatic conditions, not vulnerable to pollution from the surface. Exposed to risks of exchange between aquifers as a result of perforation of the casing	Strategy (3) Global imbalance (final equilibrium in some areas) More sustainable dewatering with stabilisation (or adjustment) of withdrawals Management of storage and some-times management of sectoral flow (near open boundaries)	Internal constraints: - maximum acceptable draw-down, - risk of intrusion of brackish water

Table 12.3 Classification of the actors in groundwater development, exploitation and management

	Users, beneficiaries	*Intermediaries, non-users*
Developers who are not exploiters	Mining industries, users of subterranean space, land occupants, surface operators responsible for hydraulic development, etc. impacting on groundwater (strategy, properties, etc.)	Authorities and public bodiescontracting authorities
Developers who are also exploiters	Businesses and associations of irrigators, industries with their own water supply	Producers/distributors of drinking water, industrial water, irrigation water
	With development of subterranean dams, artificial recharge, etc.	
	Mining industries, users of subterranean space, dewatering operators	
Exploiters who are not developers	Users-exploiters of water (without development): households, industries with their own water supply, irrigators. Exploiters and users of geothermal energy	Producers/distributors of drinking water, irrigation water (without development). Exploiters of geothermal energy, distributors (heat network)

conflicts' arise. The aquifer is like an unrecognised 'condominium' without any spontaneous ad hoc regulation. By the same token, if the exploitation or use of an aquifer as a whole has harmful effects on a third party – for example, the users of dependent surface water – no individual among the community of exploiters acknowledges a collective responsibility. None of the aquifer's stakeholders assumes the role of manager, and, no individual acts at system level or, hence, at the level of global management.

On the one hand, there is the physical unity of an aquifer system, and on the other a multiplicity of various actors with different and, sometimes, diverging interests. This being the case, the management of the aquifer requires that the behaviour of the actors is guided by applying pressure or providing incentives in different forms. In contrast to the management of a dam which is operated by its manager, the management of an aquifer is, of necessity, partially indirect in character, and at two levels:

(a) that of the actions relating to the groundwater, which are carried out by a set of actors specific to each aquifer system. Such actions may be direct and intentional or are the indirect results of various activities,

(b) that of the actions aimed at the behaviour of the actors. This corresponds to a management authority vested with appropriate powers.

This management at two levels applies to water management in general. The particularity – and difficulty – of groundwater management lies in the fact that:

- the actors at level (a) do not have 'resource management' as a spontaneous objective and their individual fields of action do not, in general, correspond to the relevant level of an aquifer system, although they alone directly operate the groundwater and the effects of their acts accumulate and are felt within the system as a whole;

- the management authorities at level (b), when they exist, can define relevant and collective management objectives and are well suited to the required level, but, on the other hand, they do not, as a general rule, have any means of direct action on the groundwater. The only way they can take action in regard to the exploiters and 'developers' is by using various indirect 'management tools'.

12.3.3 Management objectives

The objectives and decision-making criteria at the two management levels differ.

(i) Individual actors

The individual actors are mainly interested in micro-economic production and consumption objectives; groundwater is, at best, a factor, a 'raw material', a vector, or merely an 'object' at the receiving end of impacts and 'external effects'. The direct exploiters of groundwater – to secure water for their own use or to produce water for distribution – manage their water production individually on the basis of their own criteria. They do not manage the resource itself as long as production does not suffer because of its repercussions on the environment. They do not have 'resource management' objectives, but do have the implicit objective of individually preserving the conditions sustaining their water production. The same is true of the actors who remove groundwater not in order to use it, but to remove it from the underground that is to be occupied or exploited (dewatering of mines, land drainage, etc.), and for whom the water is not a resource but an obstacle to be removed in order to achieve the aim of mineral production or use of underground space.

(ii) Management authorities

It is at the level of management authorities that decisions are taken concerning the aquifer as a whole and the objectives of general interest (not all of which are necessarily compatible). Consequently, genuine macro-economic management objectives can be defined and placed in order of priority:

- The prevention and arbitration of conflicts between individual co-exploiters (or people whose activities have an impact) in the same aquifer system, while ensuring respect for external constraints. This means the prevention of conflicts between the community of groundwater users and a wider community of users of surface or subterranean water and space: in short, the minimum and most common objectives of water policing, in regard to groundwater.
- The conservation of the capacities and 'accessibility' of groundwater, including its initial dynamic levels, which implies that the exploiters who were first to arrive have an advantage.
- The conservation of the potential resource in terms of both quantity and quality, in the interests of the present and future community of exploiters, or the prevention of excess exploitation that would damage the renewal possibilities or the properties of the water, i.e. the prevention of 'overexploitation' (see sub-section 12.2.2(*iv*)).
- The allocation of the resource, with priority being given to certain present and future demands (e.g., drinking water supply), which can imply 'reservations'. This means arbitration that is more planned or State-managed, in the general interest and on a long term basis, or is brought about by a market mechanism for water rights.

- Intensifying the use of the resource if it is considered to be 'underexploited', by using groundwater, rather than other water resources to satisfy certain demands. Getting the most out of the resource; that is, not leaving 'unused' a production factor 'unused' that could contribute to socio-economic development if it could be linked to the objective of allocating the resource, as a matter of priority, for the most beneficial uses.

Given that groundwater is usually an integral part of the overall water resource, these different groundwater management objectives are themselves dependent on the water management objectives in general, and must sometimes be reconciled with subsurface management objectives (mineral resources, subterranean space), and land management objectives. In the last analysis, they are dependent on the wider socio-economic objectives of the community concerned.

The objectives may be incompatible, e.g., conservation management is not compatible with the kind of management which promotes development and therefore requires transformations and redistribution. Consequently, the choice of priority objectives is the primary management option.

12.4 Management constraints and criteria

Groundwater resource management is subject to constraints linked to the appreciation of the internal and external effects of the exploitation, in the light of different criteria. The internal effects, the inevitable repercussions of exploitation on the exploitability of the resource, engender internal constraints stemming from cost/benefit analyses based on the criteria of the exploiter. The external effects or impacts can engender external constraints stemming from arbitration between the interests and objectives of the users of the water produced and those of other actors.

12.4.1 Internal constraints

The internal constraints determine the limits of overall or sector exploitation in keeping with exploitability criteria based on the dimensions of the resource and the water demand mentioned above: discharge, location, conditions of access and mobilisation, water properties, cost. But the criteria of the exploiters and those of a management authority are not defined at the same levels, and can differ:

- for the exploiter, the most important considerations are local productivity, the imposed practical methods of exploitation, the quality of the water produced, security of production, direct production costs (to be minimised in a micro-economic perspective);
- for the management authority, the most important considerations are the conservation of the resource in terms of both quantity and quality, an equitable distribution of the conditions of access and productivity, and the overall cost of exploitation (to be optimised in a macro-economic perspective).

It is always necessary to classify and grade these criteria in order to determine management constraints. If, for example, the most important criterion is that of cost, in the form of the maximum unit production cost acceptable, this means pumping from the maximum depths practicable and, as a result, the fixing of a ceiling for withdrawals. The conservation of the dynamic levels of groundwater without exceeding a maximum acceptable depth established for each individual zone is the constraint most often generated by the criteria. Where the management of a non-renewable resource is concerned, if the criterion of duration of production is paramount, it is the chosen duration which determines the ceiling constraint on annual withdrawals per sector.

12.4.2 External constraints

The external constraints are based on the desire or the need to reduce or prevent external effects which are detrimental to other economic actors, users of dependent surface water, land occupants, or, more generally, to the environment (e.g., preservation of aquatic ecosystems, which are dependent on outlets). They can stem directly from the obligation to maintain 'reserved discharges' on the surface. These constraints are the result of negotiated consensus and arbitration between the objectives and sectional interests of the different parties concerned, based on socio-economic criteria which are in the general interest. They can also lead to financial mechanisms of internalisation. The constraints also very often impose the maximum admissible depths of groundwater levels.

The external constraints combine with the internal constraints, but take priority: in this way, they can impose additional limits on the direct use of groundwater. Both the external constraints and, sometimes, the internal constraints can find legal expression in regulations (prohibitions, obligations). Conversely, groundwater managers can impose constraints in order to conserve and/or protect the resource. These result in external constraints on other actors:

- on developments and/or the use of surface water when a development or use is likely to impact on the groundwater strategy or properties in a manner detrimental to its users;
- on various activities relating to land and subsoil occupation, e.g., prevention of subsidence caused by dewatering, and prevention of pollution, particularly in the sanitary buffers around catchments of water intended for human consumption.

12.5 Management decision methods and aids

The management decisions concerning a given aquifer depend on whether the management is direct or indirect. The decisions may be strategic (choice of strategy and general plan of exploitation) or tactical (choice of machinery and technical modes of exploitation and guidance of the production), They can be based on predictive management techniques which are themselves primarily based on hydrodynamic or optimisation modelling, and management control techniques.

12.5.1 Modelling and predictive management

The principles and ways of the hydrodynamic modelling of aquifer systems are dealt with in Chapter 11 and are the subject of an abundant literature. The models are used as tools for study and analysis, checking the consistency between the observed data, hypotheses and calculated results, and can also help to optimise data collection (Pfannkuch, 1975). They also simulate behaviour, so that forecasts can be made of the consequences of actions envisaged on the aquifer system, including exploitation scenarios, and exercises in predictive management can be carried out. The results obtained provide useful indications about feasibility and comparative data on different variants of the exploitation projects in well-defined time frames.

The primary exploitation answers from such models concern forecasts of changing drawdown rates of the piezometric levels, decreasing discharge of natural outlets, and sometimes the inferred impact on salinity. Moreover the modelling results make it possible to estimate foreseeable production costs in terms of the depth variable, or to compare the calculated future states of the system within the established constraints. Such models are thus outstanding aids to management decision-making and their use has become commonplace (Pfannkuch, 1975; Bachmat et al., 1980; Hogh Jensen, 1987). The model must be sufficiently representative to make a reasonably valid simulation of variations in the state of the system, even when the situation differs significantly from the situation used to calibrate the model. The conceptual schematisation must be relevant – particularly in the case of multilayered systems – and the adopted storage coefficients must be valid, because they have a large influence on the results of the model. It is advisable to establish margins of uncertainty (confidence intervals) for the numerical results.

A hydrodynamic simulation model can also constitute the infrastructure of an economic management model, especially for the purposes of optimisation, and be generally integrated into a basin management model. Lastly, if the models are periodically updated and adjusted in the light of the improved knowledge resulting from the operations themselves, they can also guide management by providing revised projections.

12.5.2 Forecasting unit production costs

Unit production costs and their evolution, in constant monetary units per cubic metre of water, are an essential ingredient of the comparative feasibility studies of projected exploitation plans. They are used in particular as the basis for predictive cost/benefit analyses. The direct costs borne by the exploiters are calculated on the basis of fixed terms (investments) and of variables linked above all to the operation:

- average envisaged productivity per well, pumping station, or spring,
- desired total yield, possibly in stages, and number of wells programmed to achieve this result,
- expected average life of wells and pumps,
- costs of initial wells and their equipment and of the future replacement costs,
- time allowed for amortisation of investments,
- operating costs (fuel, maintenance, etc.), based in particular on pumping levels inferred from dynamic depth levels calculated at different times.

12.5.3 Forecasting external costs

The exploitation of groundwater can have harmful external or internal effects or impacts, as examined in Chapter 11. A forecast of the resulting costs should also be included in the feasibility studies and cost/benefit analyses of projects, whether or not it is used to incorporate such costs in the direct costs borne by the exploiters. Depending on the types of impacts, these external costs are not all readily calculable in monetary units but their likely magnitude can be indicated.

- Decreases in productivity of previous drilled wells, in particular of traditional modes of capture, can be evaluated on the basis of compensation costs, alternative solutions, or the costs of the resulting losses in production (in agriculture, etc.).
- Degraded water properties can be evaluated in terms of the costs of remedial measures, or by those of solutions involving alternative sources of supply.
- Land subsidence or drainage harmful to vegetation can be evaluated on the basis of the costs of remedial measures or reductions in land value (depreciation of assets).

12.5.4 Optimisation methods

Simulation and predictive calculation methods enable the different exploitation options and scenarios to be compared in terms of their costs and expected results. However, they do not deal with the question of determining the optimal solution. The multicriteria methods used for analysing management options provide a new and efficient approach (Tecle and Duckstein, 1992), that augments the function-objective and single-criterion based optimisation techniques, such as linear and dynamic programming (Wanakule et al., 1986). Their principles and ways of application to aquifer management, are not dealt with here (but see Duckstein et al., 1993; El Magnouni, 1993; El Magnouni and Treichel, 1994).

The management of a given aquifer is rarely based on a single objective or criterion, for example, to maximise discharge or duration of production, or to minimise cost. Instead a multiplicity of criteria are involved. In most cases, each criterion has to do with an aspect of the aquifer's behaviour; the criteria do not have a common scale (for example, of monetary values), however. A hierarchy of preferences has to be established and formalised for the criteria. For example, priority can be given to the quantity of water produced, but not at any cost or at the risk of an unacceptable deterioration in quality or an extremely short duration of production. Information on the preferences both inherent in and established between the criteria should logically enable the establishment of an optimal solution, also known as the 'best compromise'. The interconnection between the simulation model of an aquifer and a model of the value system of the decision-maker (constraints, criteria and preferences) also constitutes a multicriteria 'management model' of the aquifer (El Magnouni, 1993). But that implies the existence of a management authority in which the power to take decisions is a centralised.

12.5.5 Management control

To check whether the changes in an exploited aquifer are in keeping with forecasts and the programme, the management authority should have regularly updated data on the condition of the aquifer and on the amount of water withdrawn. This requires the monitoring of two types of significant variables: variables of condition (e.g., piezometric levels, salinity of the water produced, natural discharge outlets if relevant) and decision variables (e.g., withdrawals, overall state of wells).

(i) Monitoring of piezometric levels

The methods and techniques of monitoring groundwater levels by piezometric networks and the analysis of the data obtained are dealt with in Chapters 3 and 10. In addition to providing insight into aquifer dynamics, such monitoring provides essential information for management control. Groundwater levels are, after all the most significant state variable; they indicate the states of flow and of storage, and reveal all the impacts. In order to monitor the exploited aquifer, a piezometric network should have reference observation points in zones not subject to influence from the groundwater exploitation as well as at points close to the well fields.

In the case of deep aquifers where specially drilled piezometers are rare, because of the high costs, the following can often be used as piezometers:

- unexploited unproductive water wells,
- boreholes used for mining or oil exploration which pass through the aquifer and can be partly filled up and perforated at the desired depth.

The frequency of measurement should be adapted to the type of aquifer and operation strategy. Monthly or more frequent monitoring can be useful in the case of a low-capacity aquifer which is intensely exploited with seasonal variations. Annual monitoring is enough for high-capacity confined and unconfined aquifers.

(ii) Monitoring of outlet discharges

The monitoring of the discharges from springs and gaining streams, identified as outlets of the exploited aquifer, provides essential information, particularly for evaluating the possible impacts of the exploitation on boundary flows. Such monitoring, using traditional hydrometric techniques, should be specifically adjusted to low discharge periods.

(iii) Monitoring of the salinity of yields

It is advisable to sample and analyse the main physical and chemical variables of quality (resistivity, chloride, sulphate content, etc., temperature) of yields annually from a considerable number of the abstraction points.

(iv) Monitoring of water production

In cases of direct management, the operating authority regularly quantifies the amount of water withdrawn. In other cases water production monitoring can be based on several methods, some of which are more precise and costly:

- the installation of meters which are read periodically (including the measurement of 'artesian' well discharges) is difficult to carry out in practice when there are many wells or boreholes and these, provide moderate yields on average;
- periodic surveys, in which indirect indicators (energy consumption, area irrigated, etc.) are measured or recorded.

Making it mandatory for the operators to declare their withdrawals can facilitate a quantifying of withdrawals. It does not exempt the management authority from some degree of inspection, however.

(v) Data bank and information

A data bank of computerised geological references, possibly linked to a geographical information system (GIS), is the modern way of filing all the data collected periodically and validated. It should enable tables, graphs and updated maps to be produced, including reports on levels and yields. these are the management control data on the operation of the managed aquifers. The output could be published and distributed periodically as a news-sheet among users as well as regional and national administrative bodies responsible for water, the environment and economic development.

12.6 Management instruments

Collective management systems similar to the surface hydraulic management systems used to manage a catchment basin are rarely used for the direct technical management of an aquifer, except for specific cases such as subterranean dams, artificial recharge installations, etc. The actors directly exploiting or impacting on groundwater are not individually involved in resource management. This management devolves on authorities, who set up indirect management instruments which guide the behaviour of such actors so that their decisions will all be as consistent as possible with the management objectives established in the general or community interest. These indirect management instruments are primarily of two kinds: regulatory and economic.

12.6.1 Regulation

Legislation consistent with the water laws of the country subject individual and collective operators to certain obligations:

- either a simple strategy for statistical purposes with a *declaration* of withdrawals and/or pumping units being exploited, installed or abandoned by the user, and a kind of 'registry' kept by the management authority. There is usually a discharge or depth threshold ;
- or, in addition, a prior *authorisation* strategy requiring a licence to install a catchwork, and/or to withdraw water; here again with application of thresholds.

Other legislation can restrict or prohibit acts by land or subsurface users which are likely to lead to a deterioration in the natural properties of the groundwater, with possible zoning (a particular case being the sanitary perimeters established to protect the public catchworks of water for human consumption). A water 'policing' authority or water or subsurface administrative body can be responsible for applying such legislation.

12.6.2 Financial incentives

Financial incentives, also established within a legal framework, can be used to guide and change the behaviour of the private and collective economic actors operating or impacting on groundwater, in the way desired by the management authority. Such incentives can function in two ways:

- Positive incentives or aid: stimulation of exploitation by investment credit, subsidies, 'free' insurance against prospecting and well-drilling risks, tax refunds on energy sources for pumping; including aid given to actions to lessen pollution.
- Negative incentives or taxation: restriction on or dissuasion from exploitation by the establishment of royalties on withdrawals above a number of ceilings which are adjustable according to zone, season or user category, in the context of a resource allocation policy. Also, taxation of acts which generate pollution, on the basis of 'the polluter pays' principle..

Lastly, the effectiveness of the measures can be enhanced by providing all the actors concerned with information, in order to ensure greater understanding of problems and explain objectives which are in the common interest and the reasons behind the measures taken. In general, it is advisable that the powers to make regulations and provide financial incentives and information be vested in the same management authorities. In all cases, the remit of these authorities should cover whole physical units of management (aquifer systems), irrespective of the size of such systems and whether or not they are divided between several administrative and political districts.

12.7 The future of groundwater management

Groundwater management is likely to become more of a reality in the future, to be more integrated with water and land management, and, at times, to be more ambitious.

12.7.1 Towards groundwater management implementation

In many countries groundwater management still largely comprises statements of intention and the installation, rather than the implementation of regulatory instruments. This is because the applicability and application of these instruments require resources and motivation, but these are often lacking. As long as groundwater management is only dealt with in terms of describing such instruments and attributing the responsibilities for their application, and defining administrative 'powers', it will remain hypothetical. It is foreseeable – or in any case, desirable – that in future groundwater 'administration' should put greater emphasis on actual water management which, should be in some form 'co-management'. This will involve more aware and better informed managers, common and priority objectives, and management authorities adapted to each unit formed by natural structures.

12.7.2 A more integrated management

The management objectives should not be restricted to those of the exploiters or direct users of groundwater, but be consistent with groundwater management objectives in general – groundwater as resource and environment – and with those relating to land and subsurface occupation and use. This could imply a variety of priorities, depending on the individual case and zone:

- production of water by the maximised intensive exploitation of groundwater, preferable to withdrawals from main streams, especially when such surface water is poor in quality or governed by low-quality objectives; or, on the contrary, the long-term conservation of current exploitation conditions;
- production of high-quality water, implying controls on land occupation;
- conservation of reserved surface discharges and preservation of the quality of perennial surface water dependent on groundwater runoff: extension to groundwater forming the upper waters of a 'reserved discharge' stream, and quality objectives assigned to the latter;
- land use (agricultural production, urbanisation, etc.) with impacts on the groundwater strategy and/or properties that can be tolerated up to a certain point;
- final dewatering of a subsurface space, implying the permanent transformation of the local groundwater strategy.

This will mean wider consultation, encompassing a variety of partners with, at times, contrasting motivations and interests, and, in general, wider political arbitration, with the particular objective of taking account of long-term collective interests (future generations).

12.7.3 A more ambitious form of management

In certain cases, groundwater management will evolve towards a form of aquifer management that is more ambitious than the mere extensive or intensive exploitation of natural flows, and involves more active mobilisation of groundwater storage and actions relating to aquifer recharge:

- increasing use of the exploitation strategies of the second type described above, using the multi-annual regulation capacities of certain aquifers, particularly to compensate for surface shortfalls in runoff during dry years (replacement resource or minimal flow support), sometimes based on spring tapping at different levels or on subterranean dams. The regulatory function of aquifers containing fairly significant and manoeuvrable quantities of water can be put to greater use, together with surface dams, to control irregular flows, especially in countries where the useful capacities of surface reservoirs are rapidly deteriorating as a result of sedimentation;
- development of artificial recharge, often combined with the preceding strategy;
- utilisation and maintenance, even stimulation, of the self-purifying capacities of certain aquifers in the 'management of the properties' of water, particularly in systems which reutilise waste water;
- reservation of certain soils for the production of water of excellent quality. 'Hydroculture' might take precedence over agriculture, taking over from it within the framework of a land development policy and an appropriate agricultural policy;

- exploitation of non-renewable resources, managed in a long-term perspective and a view to the preparation of alternative sources. One way this could be achieved is by investing some of the profits gained from using the abstracted water, to enable changes in the water economy.

12.8 References and additional reading

ANONYMOUS, 1991. *Sustainable Use of Groundwater Problems and Threats in the European Communities*. E.C.E., RIVM-RIZA, Rep. No. 600025001 to Ministers Seminar on Groundwater, The Hague, 26–27 Nov. 1991, 80 pp.

ASCE, 1972. *Groundwater Management*. ASCE Manuals and Reports on Engineering Practice, No. 40, New York, 216 pp.

BEAR, J.; LEVIN, O., 1967. The Optimal Yield of an Aquifer. In : Management of Aquifers. In : *Alimentation artificielle et aménagement des nappes aquifères.* Internat. Assoc. Hydrol. Sciences, Proceed. Sympos. Haifa, IASH/UNESCO-DHI, Publ. IASH, No. 72.

BACHMAT, Y. ET AL. 1980. Groundwater Management. The Use of Numerical Models. Amer. Geophys. Union, Water. Res. Monogr. 5, 127 pp.

CUSTODIO, E.; GURGUI, A., (eds.), 1989. *Groundwater Economics*. United Nations International Symposium and Workshop, Barcelona, Oct. 1987, Elsevier, 625 pp.

DUCKSTEIN, L.; TREICHEL, W.; EL MAGNOUNI, S., 1993. *Multicriterion Analysis of Groundwater Management Alternatives.* University of Paris-Dauphine, Cahier du LAMSADE No. 114, Paris.

EL MAGNOUNI, S., 1993. Méthodologie d'aide à la décision pour l'évaluation et la gestion multicritère des ressources en eau souterraine. Doctoral thesis, Univ. Nancy, Nancy, 134 pp.

EL MAGNOUNI, S.; TREICHEL, W., 1994. A Multicriterion Approach to Groundwater Management. *Water Resources Research,* Vol. 30, p. 1881–96.

ERHARD-CASSEGRAIN, A.; MARGAT, J., 1983. *Introduction à l'économie générale de l'eau.* Masson, Paris, 361 pp.

HUBERT, P., 1984. Chap. XI : Aspects de la notion de ressource en eau, et Chap. XIV : Gestion des eaux souterraines. In : *Eaupuscule ; une introduction à la gestion de l'eau.* Ellipses, Paris, 192 pp.

HOGH JENSEN, K., 1987. *The Value of Groundwater Models for Planners and Decision-Makers.* UNESCO, Tech. Doc. in Hydrology, Project IHP.III 2-4(b), Paris.

LLAMAS MADURGA, M. L., 1975. *Motivaciones extraeconómicas en la utilización de las aguas subterráneas; la hidroesquizofrenia.* National conference on general and applied hydrology, Zaragoza, Spain, 1975.

LLAMAS, R.; BACK, W.; MARGAT, J. , 1992. Groundwater Use; Equilibrium between Social Benefits and Potential Environmental Costs. *Applied Hydrogeology,* Vol. 2, pp. 3–14.

MANDELS, S., 1979. Problems of Large-Scale Groundwater Development. *Journal of Hydrology,* Vol. 43, p. 439–43.

MANDELS, S.; SHIFTAN, Z. L., 1981. Groundwater Resources; Investigation and Development. In: *Criteria for the Regional Exploitation of Groundwater,* Acad. Press, New York, 369 p.

MARGAT, J.; SAAD, K. F., 1983. Concepts for the Utilization of Non-Renewable Groundwater Resources in Regional Development. *United Nations Natural Resources Forum,* Vol. 7, pp. 377–83.

MARGAT, J., 1985. Groundwater Conservation and Protection in Developed Countries. *IHA, Report 18th Congress: Hydrogeology in Service of Man, Vol. XVIII, Part 1, Keynotes paper,* Cambridge, pp. 270–301.

MARGAT, J,. 1992. The Overexploitation of Aquifers, in: Selected Papers on Aquifer Overexploitation. *I.A.H. Selected Papers on Hydrogeology,* Vol. 3, pp. 29–40.

PFANNKUCH, H. O., 1975. *Study of Criteria and Models Establishing Optimum Level of Hydrogeologic Information for Groundwater Basin Management.* Water. Res. Research Center, Univ. Minnesota, Minneapolis, Bull. 81, 165 pp.

SAHUQUILLO, A.; ANDREU, J.; O'DONNELL, T., (eds.), 1989. *Groundwater Management, Quantity and Quality.* Symposium 2–5 Oct., Benidorm. IAHS Publ. No. 188, 2 Vols., 1015 pp.

SCHNEIDER, H., 1977. *Management of groundwater resources in arid zones*. Conf. Different Strategies, Development and Management of Deserts, Vol. 11, Sacramento, 25 pp.

TECLE, A.; DUCKSTEIN, L., 1992. Concepts of Multicriterion Decision-Making. In: *Decision Support Techniques for Integrated Water Resources Management,* UNESCO, Paris.

UN/EEC, 1983. *Groundwater Protection Strategies and Practices*. UN/EEC, Seminar, Athens, Oct. 1983, Proceed. Publ. Minist. Energy and Natural Resources, Athens, 501 pp.

UNESCO et al., 1983. *Groundwater in Water Resources Planning.* Proceed. Internat. Symposium, Coblenz, Aug.–Sept., 3 Vols., UNESCO, IAH, IAHS, Nat. Comm. Fed. Rep. Germany.

WANAKULE, N.; MAYS, L. W.; LASDON, L. S., 1986. Optimal Management of Large-Scale Aquifer: Methodology and Application. *Water Resources Res.,* Vol. 22, No. 4.

WIENER, A., 1972. Role of Groundwater and Surface Water in the Management of a Basin. In: *The Role of Water in Development,* McGraw-Hill, New York, pp. 385–93.

WRIGHT, C. E. (ed.), 1980. *Surface Water and Groundwater Interaction.* Studies and Reports in Hydrology, UNESCO, Paris, Vol. 29, 123 pp.

Selected references on case studies of groundwater management

ANONYMOUS/K.F.U.P.M. 1987. *Groundwater Resources Evaluation in Saudi Arabia and Long-Term Strategic Plan for Fresh Groundwater Use.* King Fahd University of Petroleum and Minerals, Dahran.

ANONYMOUS/UNESCO, 1972. *Étude des ressources en eau du Sahara septentrional, Algérie-Tunisie*. UNESCO, Final Report REG 100, 7 vols., technical annexes, Paris.

COLLECTIVE WORK, 1981. *L'exploitation des réserves d'eau souterraine en zones arides et semi-arides*. Doc. CEFIGRE, Sophia-Antipolis, Sept. and Bull. BRGM, Sect. III, 1982, No. 2, Orléans. This work contains articles on a variety of case studies: aquifers in large sedimentary basins in Algeria-Tunisia, Libya-Egypt, Senegal and Mauritania, Arizona and Australia.

COLLECTIVE WORK, 1983. *Groundwater Protection Strategies and Practices*. UN-ECE Seminar, Athens, Greece. This seminar's proceedings include reports on groundwater management in different countries: Austria, Bulgaria, Canada, Federal Republic of Germany, Finland, France, Greece, Hungary, Netherlands, Poland, Switzerland, Ukraine, USSR.

FOSTER, K. E., 1977. Managing a Finite Groundwater Supply in an Arid Area; the Santa Cruz Basin Example in Arizona. In: *Proceed. Conf. On Alternative Strategies for Desert Development, May,* Sacramento, Ca., 21 pp.

HABERMEHL, J., 1980. The Great Artesian Basin, Australia. *B.M.R. Journal of Australian Geology and Geophysics,* Vol. 5, pp. 9–18.

MARGAT, J., 1990. *La gestion des eaux souterraines/France*. Prospective studies group, seminar, 'Évaluation des politiques publiques de l'environnement', meeting of 15 February 1990, Paris, 18 pp.

SALEM, O. M., 1992. The Great Manmade River Project; A Partial Solution to Libya's Future Water Supply. *Water Resources Development,* Vol. 8, pp. 270–8.

SCHEFTER, J. E., 1985. *Declining Groundwater Levels and Increasing Pumping Costs in Floyd County, Texas; A Case Study*. National Water Summary, 1984. Hydrologic Perspectives, USGS, Water Supply Paper 2275, Washington.

SHAHIN, M., 1991. Groundwater Management. In: *Assessment of Groundwater Resources in Egypt,* IHE, Report Series No. 23, Delft, 71 pp.

SOLIMAN, M. M., 1984. Groundwater Resources Planning and Management in the Libyan Sahara. In: M. M. Soliman (ed.), *Groundwater Management in Arid Regions,* Contrib. for the Intern. Course on Groundwater Hydrology, UNESCO/Ain Shams University, Cairo, 2 Vols., 275 and 136 pp.

13 (A) The influence of changes in hydrogeological conditions on the environment (B) Groundwater quality protection

13A The influence of changes in hydrogeological conditions on the environment

13A.1 Introduction

Many artificial disturbances in the environment result from changes in hydrogeological conditions. Prolonged fluctuations in groundwater levels and changes in the chemical composition of groundwater can lead to:

- substantial disturbances in the functioning of groundwater and surface ecosystems,
- activation of geodynamical and engineering geological processes (e.g. karst, landslides, subsidence, erosion, flooding, etc.),
- changes in soil and cryogenic processes and surface water flow,
- seismic activity.

Because these negative environmental impacts occur in regions where the groundwater regime is disturbed, forecasting them is associated with forecasting the disturbed groundwater regime. Therefore, the purpose of hydrogeological investigations is not only to study and forecast external effects on the groundwater, but also the influence of changing hydrogeological conditions on the environment (Kovalevsky, 1994). Understanding these consequences requires monitoring of the groundwater system and other related environmental components. It is essential to predict the development of critical and disastrous situations to the environment caused by groundwater, and to apply preventive measures. Three types of disturbance of the groundwater regime have been identified:

- artificial discharges or increases in the outflow components of the groundwater balance that are characterised by lowering of the groundwater levels;
- artificial recharge of groundwater or increase in the inflow components characterised by groundwater level rises, and
- changes in the chemical balance of the groundwater, characterised by changes in the groundwater quality.

High rates of groundwater abstraction and the resultant decline in groundwater heads take place in regions of centralised water abstraction for domestic and industrial purposes, in mine water discharge areas, in the regions of oil and gas fields, and in irrigation and land drainage zones. First of all, the exploitation disturbs the groundwater regime and balance. The groundwater level decline may lead to a decrease in groundwater recharge, infiltration and evaporation. The formation of cones of depression may lead to a reduction in the yield of wells, the disappearance of springs and a reduction in groundwater discharge to rivers. Under the influence of groundwater withdrawal, salt water encroachment may occur, sea water intrusion into aquifers is observed, and in desert areas salt water contours approach the fresh groundwater lenses. Changes in the water temperature in cones of depression is also observed (Kovalevsky, 1976). All these factors change the salt balance and the quality of the groundwater and often make the groundwater unfit for use.

All these processes can be studied, and there are many types of models to forecast them (e.g. Goldberg, 1976; Gavich, 1980).

13A.2 The environmental impact of groundwater withdrawal

13A.2.1 Effects on the relation between groundwater and surface water

Groundwater development changes the interaction of surface water and groundwater; groundwater discharge to rivers either decreases or the rivers begin to recharge the aquifer. As a result the surface water resources decrease. The flow of small rivers decreased by 70–80 percent in Belarus, by 10–15 percent in the Moscow region (Russia), and some rivers of Kazakhstan have fallen dry. The water withdrawal in lakeland areas also leads to reduction of water depths, drying, and the loss of recreational potential and fish production. For example , the decrease of cold water discharge into Lake Pleshcheevo (Russia) has led to an increase in the temperature of the bottom layer and the water of the lake. This has intensified the decomposition of silt, enriched the lake water with organic matter, and reduced the oxygen content, resulting in a decline of the productivity of some rare fish in this region. Marshland reclamation is also accompanied by enrichment of water with organic matter, iron, manganese, and nitrogen compounds. The acidity of water increases and the solubility of heavy metals, including toxic ones, increases. The base flow decreases, and deflation and spontaneous ignition of peat land may occur.

13A.2.2 Land subsidence

One widespread consequence of intense groundwater development is the consolidation of sediments, resulting in land subsidence and fractures that cause the deformation of buildings, roads, etc.; Konoplyantsev and Yartseva-Popova (1983) mention examples in United States of America, Japan, Mexico, Italy, The Netherlands, and other countries. Subsidence rates range from a few millimetres to 75 cm per year, and the resultant subsidence can be several metres (e.g. up to 8.8 m in California, United States of America). The subsided volume in Texas (United States of America) is 22 percent of the volume of groundwater pumped from sandy and clayey sediments. Subsidence is accompanied by a change in the state of both water-bearing and confining layers. Subsidence is caused by the release of the hydrostatic pressure. At first the compaction is due to deformation by compression of the layers; this is succeeded by plastic deformation that is accompanied by the displacement of rock particles. The interaction of particles is increased by the forces of molecular attraction, resulting in an increase of the amount of bound water. These processes are irreversible and ground surface levels recover only partially when water withdrawal is terminated or decreased. Subsidence can be simulated by the numerical solution of equations of two-dimensional unsteady flow and equations of loamy soil deformation.

13A.2.3 The influence on karstification

Groundwater development in carbonate rocks very often results in the intensification of karst processes, formation of sinkholes and collapse of the ground surface, destroying buildings and damaging to the land and infrastructure. Lowering the piezometric levels of confined or semi-confined groundwater bodies to below river level may result in the aquifer being recharged with river water rich in free carbon dioxide, which may bring about the intensification of the dissolution of carbonates. Some pollutants and micro-organisms in the groundwater may act as catalysts of the karstification process. For example, as a consequence of the exploitation of a semi-confined aquifer in the Moscow area its piezometric level dropped by 10−70 m below the levels of the river and of the phreatic groundwater and turned this region into a recharge area. The aggressive river water and polluted phreatic groundwater encroached on the semi-confined aquifer and changed the hydrochemical situation:

- the groundwater salinity increased 1.5–2 times,
- the calcium content increased 15–20 times,
- the oxygen content increased 2–10 times,
- the pH decreased from 7.0−8.0 to 4.9–5.5,
- the free carbon dioxide reached values of 20 mg/l or more,
- the salt discharge by the pumped water increased to 100 t/year per km^2,
- the strontium content, characteristic for the leaching process, increased from 0.1–0.21 mg/l to 16–18 mg/l,
- the residence time, estimated by the tritium method, decreased from about 100 years to 2–12 years,
- the helium content decreased 15–20 times.
- The river water that reached the semi-confined aquifer caused an increase of the number of micro-organisms in the groundwater to 14,000/ml divided over 30 species, including:
  - four species of filamentous bacteria,
  - five species of diatoms,
  - seven species of green algae,
  - three species of green-blue algae,
  - three species of euglenes,
  - infusoria,
  - shell amoebae.

The carbon dioxide content reached 200 mg/l in zones with a high content of micro-organisms that produce weak acid (Kovalevsky and Zlobina, 1988). The leaching of carbonates in the presence of wastewater of a thermal power plant increased 4–5 times, because the high groundwater temperature resulting from the thermal pollution promoted an increase of the acidity and of the calcium deficiency (Zlobina, 1988). Thus, water withdrawal may create not only hydrodynamic, but also thermal, hydrochemical, and hydrobiological anomalies. These factors may intensify the karst processes, the appearance of sinkholes in areas where they were unknown during the first decades of groundwater exploitation.

Karstification resulting from of groundwater exploitation has been numerically simulated at the theoretical level only (Mercado, 1977).

13A.2.4 Effects on plants and animal life

Large scale pumping or drainage of groundwater inevitably results in a lowering of the water table (or phreatic surface) and the capillary fringe, expansion of the zone of aeration, a change in the soil moisture content and consequently an impact on the environment. This impact may be positive or negative (Kovalevsky, 1994).

We know numerous examples of the dewatering of forest areas in regions with artificial drainage of groundwater, resulting in a transformation of the vegetation and of the quality and productivity of the forest. The moisture-loving vegetation (hydrophytes), e.g. willow, reed and reed mace, are the first to perish in river valleys in areas that have an arid climate and deep-rooting phreatophytes, e.g. salt cedar, honeysuckle and dog-rose show considerably depressed growth. On the other hand, when the groundwater development leads to dewatering of wetlands in the humid zone, the yield of grasses, especially on flooded meadows can be substantially increased because low yielding species are replaced by much higher-yielding ones. A few examples:

- On artificially dewatered lands of the Vologda Region (Russia) sedge (150–300 kg/ha) has been replaced by high quality meadow grass (2000–2200 kg/ha).
- On flooded land only wilted bushes, low birch, and poor quality pine can grow, while on dewatered lands mixed coniferous/deciduous forests of good quality can thrive.
- The natural species succession of forests in the piedmont region of the Caucasus and in the region adjoining the Aral sea is: willow/alder/poplar/oak and hornbeam. Drastic artificial lowering of groundwater levels has caused these forests to disappear.

The correlation between the productivity of forests and other plant communities and the depth to the water table or to the capillary fringe for areas with different climatic and landscape conditions can be the basis for predicting the environmental consequences of water withdrawal and land drainage. The decline in groundwater level and the resulting desiccation of the forest litter is leading to locust plagues in the southern parts of the former USSR and to the proliferation of the encephalitic tick in the northern parts. On the other hand, better drainage leads to a reduction of the population of certain rodent species carrying febrile diseases like tularaemia and leptospirosis.

The groundwater development in small river valleys (e.g. the Ural river, Russia) may result in a decrease in flood rates, and the cessation or reduction of the flooding of floodplains and oxbow lakes. This influences the reproduction of fishes and adversely affects fishery activities.

Drainage operations in Belarusia, Lithuania and Karelia have led to:

- an increase in groundwater salinity (to 300–3,000 mg/l),
- ablation of soils (up to 15 t/km^2 per year),
- appearance of ravines,
- downcutting by rivers and watercourses,
- shrinkage and spontaneous ignition of peat,
- increase of karstification processes in the beds of canals carved into limestone..

13A.2.5 Influence on seismicity

One of the most controversial questions is the possible influence of changes in hydrogeological conditions on seismic activities, comparable to the seismic activity that occurs in the wake of oil and gas exploitations. Valeisha (1983) considers that one of the earthquakes in Tashkent is related to the large thermal water development that resulted in the hydraulic head being lowered by 150 m. The 1985 earthquake in Mexico may also be related to intense groundwater exploitation. Shtengelov (1980) suggests that most of the severe earthquakes in the Crimea

(Russia) coincide with sharp natural drops of groundwater levels; he also found a relationship between high seismic activity and long term pumping tests.

13A.3 The impact of man-induced groundwater level rise on the environment

13A.3.1 General

Artificial groundwater recharge is a widespread cause of a disturbed hydrological regime characterised by rising groundwater levels and the water table subsequently stabilising at a higher level than before the disturbance. The increased recharge may be due to:

- field irrigation systems,
- leakage from canals, drinking water mains and sewerage systems,
- industrial wastewater disposal,
- seepage losses from reservoirs, ponds and sludge tanks,
- street flushing.

Another cause of the rise of the water table is deterioration of the natural subsurface outflow. It may be caused by:

- subsurface obstructions (e.g. tunnels, embankments, foundations),
- filling up of gullies,
- permanent rise of surface water levels (dammed rivers).

The effects of the rise of the water table depend on climatic, geological and other local conditions, but they can be summarised as follows:

- flooding of settlements and urban areas, agricultural land and forests,
- waterlogging and salinization of soils,
- changes in the groundwater and salt balance,
- changes in the engineering properties of rocks that may result in:
  - changes in bearing capacity,
  - greater susceptibility to landslides, subsidence, seismicity, etc.,
  - changes in the ecological situation.

13A.3.2 Effects of waterlogging

Waterlogging is probably the most widespread and serious ecological effect, resulting in a disturbance of the water, salt, air and temperature regimes in the root zone of the natural and planted vegetation, causing a decrease in yield and productivity. The critical depth to water depends on the type of soil, the climate and the sensitivity of the vegetation to high soil moisture and increasing soil salinity. The impact of water table rise on the functioning of ecosystems depends on the relation between the depth of the root system and the capillary fringe, and the water requirements of plants at different vegetative stages. Each type of vegetation has an optimal groundwater level, below this level the root system dries up and above it air exchange deteriorates and the root system is asphyxiated or will rot away. The changes in productivity and composition of the vegetation are an indicator of the soil moisture conditions.

Over-moistening and hydromorphism of the former automorphic soil layer results in an increase of the acidity of the soil, soil gleying, change in humus content of the soil and in the exchange of cations. This brings about a drop in soil temperature, because of the heat expenditure for phase transitions during the evaporation of surplus soil moisture and the melting of frozen soils, etc. This leads to a change in seepage properties of water bearing formations (Maslov, 1970). Changes in the water, air and thermal regimes of soils, in the related redox conditions and in the chemical composition of the soil moisture, affect the structure of the soil. Changes in the formation of humus, in the amount of organic substances present and in the minerals and toxic compounds lower the biological activity of soil micro-organisms and soil productivity. In turn, this favours the above mentioned transformations of surface ecosystems in the flooding zone and their biological productivity.

Considerable changes may occur in soils of the humid zone of northern Russia where the climate makes the environment especially vulnerable to disturbances of the groundwater regime. In the permafrost areas, waterlogging will change the cryogenic situation. The warming effect of the rising groundwater causes the degradation of the frozen ground and reduces the thickness of the permanently frozen soil layers.

An undue increase of the moisture content of soils and rocks induces changes in their physical and engineering properties. When the specific weight and consistency index increase, this causes a change in the general deformation modulus, internal friction, binding angle and shear stress parameter. This results in an increase of the sensitivity to vibration, clay plasticity, and a decrease in the binding of soil particles, resulting in a decrease in the carrying capacity of the soils, the compression of subsiding soils, fine sands turning into quicksand, and disturbance of the structural bonds in soil.

13A.3.3 Effects on agricultural land

Land reclamation and especially surface water irrigation under semi-arid conditions strongly affects the groundwater balance of the involved area; unavoidable results are:

- changes of the annual groundwater regime,
- changes in the hydrochemical composition ,
- changes in the interrelationship between the phreatic and the (semi-) confined aquifers and in the groundwater drainage conditions,

and in the later stages:

- waterlogging and subsequent secondary salinization of the soil.

The methodology of predicting the hydrodynamic and hydrochemical groundwater regime under irrigation is well known (Averyanov, 1978; Baron et al., 1981; Katz and Pashkovsky, 1985).

To counteract waterlogging and secondary soil salinization, artificial land drainage, e.g. by a horizontal subsurface pipe drainage system (ILRI, 1974) is required when, at the end of the wet season, the phreatic level is less than 1–2 m below the surface.

13A.3.4 Effects of surface water reservoirs

In the areas downstream of surface water reservoirs, groundwater levels may drop and consequently meadow-alluvial soils may transform into podzolic and even dry-river-valley soils. They will become overgrown with forests and bushes, the sod will be destroyed, dunes may form and soil fertility will decrease. Upstream of the reservoir, where the groundwater levels rise, comparable transformations will occur.

Reservoir construction usually contributes to the groundwater level rise and waterlogging of the area adjoining the reservoir, where flooding, undercutting and reworking of the lower part of the shore, and erosion of the old landslide slopes reduces the soil stability, reactivating old landslides and sometimes causing new ones. Models for the simulation of the effect of surface water reservoirs on the groundwater regime have been developed by Abramov (1978) and Shestakov (1965).

The intensification of subsidence processes in waterlogging and groundwater discharge zones are not rare. In the area of the Kakhasvskoe Reservoir (Russia), subsidence effects occurred in a strip that is 100–300 m, in some places 600 m, wide. Along tens of kilometres of the shore landslides occurred in a strip 70–100 m inland.

It should be noted that most negative ecological consequences of the changes in the hydrogeological regime around reservoirs can be averted by draining the land properly, so as to avoiding waterlogging.

13A.3.5 Effects on karstification

Changing groundwater levels in limestone results in changes into the depth of the zones where karst formation is most intense, as can be observed in the areas around the Verkhnekamskoe and Bratskoe reservoirs (Russia) (Pechorkin et al., 1980).

13A.3.6 Biological effects

Waterlogging may pose a health risk (Elpiner and Boer, 1990):

- Areas with groundwater at or very close to the surface may become breeding places for vectors of many diseases, e.g. malaria mosquito (anopheles), and ideal habitats for the propagation of worms (ascarides, etc.);
- Groundwater recharge by sewage may result in a deteriorating sanitary situation, in the contamination of groundwater and soil, and in the outbreak of epidemics such as cholera;
- People living in damp dwellings in areas with poor drainage show high rates of respiratory diseases, tuberculosis, rheumatism and of stomach and intestinal diseases.

13A.3.7 Effects on seismicity

An increase in formation pressure may lead to high seismic activity. An example is the deep well injection of 21,000 m^3 per month of industrial wastewater in the aquifer at a depth of 3,671 m in the Denver area (Colorado, United States of America). This injection caused earthquakes with epicentres near the injection well. After the injection was stopped, the number of earthquakes decreased sharply. A similar reaction to deep well injection has been observed on the Russian Plateau in the Dimitrovgrad region. The storage in groundwater reservoirs of large masses of water for irrigation in the agricultural areas of the northern Crimea has also resulted in seismic activity (Lushchik et al., 1988).

The diversity of technical impacts on seismic activity may be explained by the specificity of tectonic movements. The artificial increase in loading is in the descending block, where it stimulates seismic activity, in the ascending block it restrains it. Consequently, water withdrawal in the ascending block will promote this process. It must be remembered that not every deep well abstraction or injection of water results in earthquakes, because the hydrodynamic factor is the trigger rather than the cause of the seismic activity.

13A.3.8 Effects in urban areas

The consequences in urban areas are complicated and varied. High rate groundwater withdrawal, groundwater outflow by subway and deep drainage structures, regulated storm runoff, snow removal, interception by asphalt roads and roofs, may cause an increase of the output and a decrease of the input components of the groundwater balance. On the other hand, groundwater inflow via embankments and deep foundations, leakage from water supply mains and sewage water disposal systems, industrial wastewater leakage, leakage from surface water storage ponds, street washing, deep percolation of river water, etc. bring about an increase of the groundwater balance input component and hamper the outflow. The summation of the effects may give a different outcome, even for different parts of a single city.

Depending on the lithology of the soil and rocks, waterlogging may result in land subsidence or in a rise of the surface as a consequence of swelling. Uneven subsidence on a small scale may result in the deformation and eventual collapse of buildings and infrastructures.

Predictions of waterlogging development in urbanised areas are made both by modelling and by extrapolation of stable trends in groundwater levels.

13A.4 References and additional reading

ABRAMOV, S. K., 1978. *Predicting and Preventing of Groundwater Flooding of the Territories under Construction*. Stoiizdat, Moscow, 177 pp.

AVERYANOV, S. F., 1978. *Control of Irrigated Land Salinization*. Kolos, Moscow, 288 pp.

BARON, V. A.; BETSINSKY, T. A.; BOTOVA, K. D.; KATS, D. M.; PAFYONOVA, L. I.; PETROV, P. G., 1981. *Prediction of Groundwater Regime in Irrigated Areas.* Nedra, Moscow, 246 pp.

ELPINER, L.I.; BOER, S. A., (eds.), 1990. *Manual for Predicting Medical and Biological After-Effects of Hydrotechnical Constructing*. USSR Acad. Sc., Moscow, 172 pp.

GAVITCH, I. K., 1980. *Theory and Practice of Using Simulation in Hydrogeology*. Nedra, Moscow, 359 pp.

GOLDBERG, V. M., 1976. *Hydrogeological Predictions of Groundwater Quality in Well Fields*. Nerd, Moscow, 152 pp.

ILRI, 1974. *Drainage Principles and Applications*. ILRI Publication No. 16, Wageningen, 4 Vols.

KATS, D. M.; PASHKOVSKY, I. S., 1985. *Reclamative Hydrogeology*. Nauka, Moscow, 147 pp.

KONOPLYANTSEV, A. A.; YARTSEVA-POPOVA, E. N., 1983. *Land Surface Subsidence Due to Groundwater Table Lowering.* VIEMS, Moscow, 48 pp.

KOVALEVSKY, V. S., 1976. *The Study of Groundwater Regime under Exploitation*. Nedra, Moscow, 200 pp.

——, 1994. *The Impact of Groundwater Condition Changes on the Environment.* Nauka, Moscow, 138 pp.

KOVALEVSKY, V. S.; ZLOBINA, V. L., 1988. Location of concealed areas with intensified karst-suffosion processes using helium survey. *IAH XXI Congress, Guilin, China, Proc. part II,* pp. 779–85.

LUSCHIK, A. V.; LISICHENKO, G. V.; YAKOVLEV, E. A., 1988. *Groundwater Regime Formation in the Areas of Active Geodynamic Processes*. Naukova Dumka, Kiev, 162 pp.

MASLOV, V. S., 1970. *Groundwater Regime in Oversaturated Soils and its Regulation.* Kolos, Moscow, 150 pp.

MERCADO, A., 1977. The Kinetics of Mineral Dissolution in Carbonate Aquifers as a Tool for Hydrological Investigation. *J. Hydrol.,* Vol. 35, pp. 375–84.

MINKIN, E. L., 1973. *Groundwater and Surface Water Interrelation and its Significance in Solving some Hydrogeological Water Management Problems*. Stroiizdat, Moscow, 103 pp.

PECHORKIN, I. A.; PECHORKIN, A. I.; KACHENOV, V. I., 1980. *Theoretical Basis of Geological Process Predictions on Reservoir Shores.* PGU, Perm, 85 pp.

SHESTAKOV, V. M., 1965. *Theoretical Basis for Assessment of Groundwater Lowering Aflux and Drainage*. MSU, Moscow, 250 pp.

SHTENGELOV, E. S., 1980. *On the Impact of the Upper Jurassic Aquifer on the Crimea Seismicity.* Water Resources, No. 2, pp. 57–65.

VALEISHA, V. O., 1983. Formation of Hydrogeodynamic Regime under the Impact of Counter Regional Loads on the Aquifer. Moscow, thesis MSU, 51 pp.

VERIGIN, N. N., 1975. *Filtration from Ponds and Reservoirs,* Kolos, Moscow, 303 pp.

ZLOBINA, V. L., 1988. *The Impact of Groundwater Exploitation on Developing Karst-Suffosion Processes*. Nauka, Moscow.

13B Groundwater quality protection

13B.1 Introduction

Groundwater, a renewable and finite natural resource, vital for man's life, social and economic development and a valuable component of the ecosystem, is vulnerable to natural and human impacts. Earlier, little attention was paid to the protection of groundwater quality, mainly because people were unaware of the threats to this hidden resource. The idea that groundwater is well naturally protected by geological environment, separated by soil and unsaturated zone from pollution impact, and thus not vulnerable to human activities survives a very long time and has serious consequences on groundwater quality in many countries.

Groundwater vulnerability to pollution as a side effect of human impacts was recognized as a significant world wide environmental problem during the nineteenfifties. In the nineteensixties the conceptual approach to the groundwater resources protection based on monitoring, mapping, modelling and remedial technology was developed. In the nineteenseventies groundwater protection policy and management were formulated, and appropriate pricing policy based on the 'polluter pays principle' was defined as part of water legislation. Sustainable groundwater use and development and environmentally sound management of groundwater protection became important in national water planning, policy and strategy in the course of the ninteeneighties. A holistic concept in water resources policy and management emphasized on the International Conference of Water and the Environment in Dublin (1992), significantly influenced the approach to the development, use and protection of water resources in the nineteennineties. The implementation of a holistic approach is particularly important for the management of transboundary aquifers. Their development and protection policy is a serious international hydro-political problem; the solution of which will proceed in the twenty-first century.

13B.2 Groundwater quality protection strategy

Groundwater protection strategies should be based on environmentally sound management and on the concept that prevention of pollution is always less expensive than groundwater rehabilitation, a costly, long-term and technically demanding task.

Groundwater protection should not be an emergency action only taken up when pollution problems arise. The concept of a groundwater protection strategy is translated into a long-term multidimensional programme that accommodates and integrates the environmental, technical, economic and social aspects involved. This programme is implemented, coordinated, and managed by governmental authorities. It is supported by relevant legislation, requires regular supervision and inspection, includes the education and training of responsible experts and technical staff, and the education and information of the general public (Vrba, 1994). The principal technical and capacity building criteria for the successful implementation of an aquifer protection strategy are:

- the availability of comprehensive analyses of the aquifer systems based on good knowledge of the groundwater system's unsaturated and saturated properties (the geological, physical and chemical parameters, the flow net, aquifer boundary conditions and the level of vulnerability);
- the effective operation of groundwater quality monitoring systems on national, provincial and local levels and the analysis of the data collected;

- the identification and listing of the historical, present and potential pollution sources and the evaluation of their nature, extent and real or potential impact on groundwater quality;
- a research programme supporting the development , improvement and innovation of methods and techniques for groundwater protection, quality conservation and pollution remediation;
- the existence of institutional structures that have the power and human and financial resources for the creation, co-ordination, and implementation of a comprehensive groundwater protection strategy;
- a legislative basis, regulatory statutes, codes and standards for the implementation of a groundwater protection and quality conservation programme;
- an effective inspection and control system, based on relevant legislation, ‘the polluter pays’ principle and on implementation of repressive or stimulating financial instruments;
- availability of qualified, trained, experienced and motivated professional personnel to implement groundwater protection programmes;
- a public information education and involvement groundwater protection programme, particularly regarding the ‘consumer pays’ principle.

Groundwater protection strategy is effective if the above criteria are applied in an coherent manner. Their partial or inconsistent application will not lead to a long term successful protection of the groundwater resources.

13B.3 Groundwater quality protection policy

Not all groundwater resources need to be protected comprehensively, because they possess different environmental and socio-economic values and not all of them are of the same vulnerability. To protect all groundwater resources to the same extent would be economically unsustainable, hydrogeologically pointless and unrealistic in terms of management and control. Comprehensive protection throughout the national territory of a resource as diffuse and vast as groundwater certainly is, would be too costly, would provoke conflicts in the exploitation of other natural resources, and could lead to an undesirable restriction of economic development. Groundwater protection policy should be constructive and not based on bans and restrictive measures only.

Groundwater protection policy should be:

- based on relevant legislation,
- integrated with the protection of remaining components of the hydrological cycle and other environmental elements,
- co-ordinated with land use activities and industrial development,
- based on the groundwater resources, their vulnerability and water supply needs,
- linked with the social policy,
- reflective to the cultural and historical traditions of the society.

The criteria supporting establishment of groundwater policy depend on:

- the value (quantity and quality) of the groundwater and its vulnerability;
- the current and expected demands for groundwater use and related protection in a given area;
- the importance of groundwater for the ecosystem (e.g. wetlands).

The classification of groundwater using the above criteria is always a complicated task, it can be controversial and requires good knowledge of the hydrogeological, economic and social aspects of the area concerned. An example of protection criteria is from the United States of America (Table 13B.1) in which three classes of groundwater are defined.

Table 13B.1 Classes of groundwater in the United States of America (EPA, 1984)

Class	*Basic criteria*	*Level of groundwater protection*	*Remarks*
I. Special groundwater	Highly vulnerable, irreplaceable or ecologically vital	Extremely high	Significant water resources value
II. Current and potential drinking water and waters having other beneficial use	All other groundwater use or available for drinking or other purposes	High to moderate prevention of contamination based on technological remedies rather than through restrictions	Majority of usable sources of ground-water in the United States of America
III. Groundwater not considered potential sources of drinking water and of limited beneficial use	Heavily saline or heavily polluted	Usually low migration to class I or II ground-water or discharge to surface water must be precluded	Limited beneficial use

A three-level classification system for groundwater quality has also been applied in the Czech Republic:

Class I Groundwater suitable as a source of drinking water without any treatment (only disinfectants and mechanical acidification when necessary);

Class II Groundwater which requires treatment that is justifiable financially as a source of drinking water (chemical acidification, removal of iron, manganese etc.);

Class III Groundwater unsuitable as a source of drinking water because of its natural properties, or a high level of contamination due to human impacts. Improving its quality is technically demanding and costly.

The policy of protecting groundwater as a drinking water source reflects the economic, social and cultural level of a society and its environmental philosophy and thinking. Gone are the days when groundwater protection policy focused only on the technical, economic and financial aspects while ignoring environmental and social attributes. Therefore, the objective of a modern policy on sustainable groundwater protection must be:

- timely identification and analysis of potential conflicting issues and constraints;
- formulation, quantification, analysis and validation of competitive factors with respect to land use and groundwater values;
- hierarchical screening with the aim of finding a balance between groundwater quality protection and conservation, economic development and social and health implications with a view to both the short term and long term prospects.

13B.4 Groundwater protection management

The objective of groundwater management is to ensure the quality, safety and sustainability of groundwater used, in the first and foremost place as a drinking water source, but also as part of the natural heritage and as a source of irrigation and industrial water. Groundwater protection management is not a universal or anonymous process; it is carried out and applied to particular aquifer systems having their specific natural properties, influenced by various human impacts, and therefore it varies from place to place.

Groundwater protection management must take into account: a degree of ignorance, uncertainty and unpredictability concerning the groundwater system's behaviour and properties, and the cumulative human impacts that are frequently difficult to identify. The effects of the latter cannot be predicted accurately over a longer term. The uncertainties in defining internal and external influences on groundwater systems are associated with certain risks. Therefore risk assessment and risk analysis are part of groundwater protection management.

Sustainable groundwater protection management is based on a holistic approach, reflecting the social and economic value of groundwater of groundwater and integrity of aquatic and terrestrial ecosystems, considering the close connection between groundwater, surface water and other components of the hydrological cycle in relation to land use. At the same time, attention should be given to both quantitative and qualitative aspects in view of sustainable development and environmentally sound protection of groundwater resources. Holistic groundwater protection management should be based on participatory approach, involving planners, policy and decision makers, managers, water users and the general public.

Two categories of groundwater protection management can be considered: general protection of groundwater resources and comprehensive protection around public groundwater supplies.

13B.4.1 General protection of groundwater

General groundwater protection is based on the assumption that all effectively accessible groundwater resources are, or could be, tapped for drinking or other purposes now or in the future, and therefore their protection is desirable. The water authorities should bear the responsibility for the protection of all usable groundwater resources.

Implementation of general protection of groundwater resources calls for the following activities:

- investigation of the groundwater system and determination of its vulnerability;
- identification, and listing the existing and potential pollution sources;
- operation of groundwater quality monitoring systems;
- control over human activities, in particular in vulnerable areas of groundwater basins and aquifers;
- stipulation and implementation of legislative measures for the protection of groundwater resources, and the control over the pollution sources.

In several European countries, recharge areas of important groundwater basins and aquifers have been delineated and protected within the general protection of groundwater resources. These areas cover several hundred square kilometres; human activities within these areas are under control and partially restricted by law.

13B.4.2 Comprehensive groundwater protection

The main purpose of delineating groundwater protection zones is to protect drinking water supply wells or well fields from pollution and provide the population with water that meets the standards for drinking water. In many countries, groundwater resources used for public drinking water supplies are comprehensively protected by protection zones. In the United States of America. they are referred to as wellhead protection areas, usually comprising two or three levels of protection. Wellhead protection is an obligatory part of groundwater protection management programmes, and is based on the relevant legislation. Protection of public water supply wells requires:

- delineation of protection areas for each abstraction well or spring as protection zone I and designation of protection zones II and III, which cover vulnerable and recharge areas of water supply systems, that are sensitive to pollution or other human impact;
- designation and operation of groundwater monitoring systems;
- an inventory of all potential sources of groundwater pollution;
- cooperation between local and provincial authorities, water work companies, land users, (particularly farmers) and water users (public);
- establishment of a specific plan of management of groundwater and land use activities in protection zones, with the scope to maintenance a good quality of drinking water;
- evaluation of economic losses due to the limitation of human activities (particularly farming) in groundwater protection zones and determination of the amount and the responsibility to compensate;
- implementation of institutional, legislative, technical and control measures and regulations in protection zones;
- elaboration of an emergency plan for providing alternative sources of drinking water in case pollution of water supply systems occurs.

First level protection zones protect the well and its immediate surroundings from physical damage and direct pollution. Their extent is usually small=several tens of square meters is the maximum and human activities are excluded.

The second and third level groundwater protection zones are mostly extensive (several hundreds of square metres to several square kilometres) and include the discharge areas, the cones of depression (zone of influence around pumping wells, the recharge areas and other vulnerable parts of the water supply system of interest. In several European countries second level zones cover areas having a delay or residence time of 50 to 60 days. It should be emphasized that the delay time has been determined so as to protect water supply wells from the risk of microbial contamination of groundwater, but it may become inadequate for viruses and certain persistent chemical pollutants. The third level zone protects groundwater quality in water supply wells from persistent chemical pollutants. It is the most extensive zone but there are less restrictive measures. Van Waegeningh (1985) recommends a 10 to 25 year residence time for the delineation of groundwater protection zones.

13B.4.3 Delineation of groundwater protection zones

In general, the level of restrictions and prohibitions in groundwater protection zones decreases with the distance from a well or wellfield. The second and third level protection zones cover significant areas, frequently including arable land. Over-protection of water supply wells is therefore not desirable because restrictive measures or exclusion of land from farming lead to economic losses. On the other hand, under-protection of wells and well fields may cause groundwater pollution requiring long-term and costly remedial action.

Although the environmental-social-economic system and the physical (soil and water) system, are independent of each other, their mutual relations must be coordinated and integrated with the objective of deriving benefits from utilisation of soil and water resources while conserving the quality of the environment. Management of the soil-aquifer system requires a specific approach in each region because the natural conditions and intensity of human activities, including farming, differ locally and regionally. Comprehensive protection and management of groundwater quality achieved through stringent control over agricultural activities will unfavourably affect farmers' production targets.

Objective integration of the soil/water users' interests and allocation of the benefits and costs between the water and agricultural sectors is the key factor in the management strategy of effective utilisation of soil and water resources. Under all circumstances the distribution of costs and benefits is a very sensitive and critical point, and should be dealt with in a timely and responsible manner.

It is widely held that protection zones should be as small as possible but as large as necessary. Expressed in financial terms, higher input costs on accurate delineation of protection zones will result in reduced operational costs for well and wellfield protection. Since water supply systems operate for a long time, the operating costs involved in their protection should be as low as possible. A sophisticated approach to protection zone definition is therefore preferable. There is no question about groundwater protection zones having a positive influence on the quality of groundwater (Table 13B.2).

Table 13B.2 NO_3 content in groundwater, in mg/l; regional monitoring system in central Bohemia (Czech Republic)

	NO_3 content		
	1968	*1986*	*Annual increase*
Type of monitoring station		*mg/l*	
Wells and boreholes without protection zones	24.0	43.0	1.1
Public wells and well fields with protection zones	21.0	28.0	0.4

13B.5 Groundwater pollution control

13B.5.1 Influence of natural processes and human impacts

Groundwater contains a broad range of dissolved solids in varying amounts (major, minor and trace constituents). The natural composition of groundwater is controlled by the rock medium in which groundwater moves; contact time of water with rock; a complex of geochemical (dissolution, precipitation, hydrolysis, adsorption, ion exchange, oxidation, reduction), physical (dispersion, advection, filtration) and microbiological (microbial metabolism and decomposition, cell synthesis) processes; the presence of gases (especially carbon dioxide); soil properties; climate and the impacts of other components of the hydrological cycle (e.g. sea water intrusion). The potential solution and precipitation of minerals and rocks depend on the mechanism of reactions between water and rock.

The properties of natural groundwater may not always meet the criteria for drinking water, nor be suitable for farming or industrial needs, because groundwater quality can deteriorate by natural processes and human impacts. Influence of natural processes on groundwater quality is long term and difficult to control. Human impact can significantly influence and accelerate natural processes in groundwater/rock systems. The most important causes of groundwater quality deterioration are different kinds of pollution, and over-exploitation of aquifers that change the groundwater flow dynamics (Chapter 13A). Groundwater pollution is understood to be a process whereby due to human impact water suddenly or gradually changes its physical, chemical or biological composition and ceases to meet the criteria set for drinking water. If it contains hazardous or toxic compounds water becomes dangerous for people and other living organisms and ecosystems. The vulnerability of a groundwater system to changes in its quality depends on hydrogeological conditions, on the above mentioned chemical, physical and biological processes in soil, rock, and groundwater, and the type, extent and intensity of pollution.

Pollution of groundwater is a hidden and long term process and, due to the mostly slow movement of the pollutant and the groundwater, a period of several weeks, months or sometimes even years will elapse between contaminant leakage and groundwater pollution. The groundwater quality is most frequently affected by:

- heavy metals present in liquid and gaseous wastes (leakage of industrial waste water,
- mining water or processing water, waste water from uncontrolled landfills, etc.),
- organic chemicals (particularly uncontrolled leakage of hydrocarbons, halogenated hydrocarbons and chlorinated hydrocarbons),
- inorganic and organic fertilizers as a main source of nitrate pollution of groundwater under arable land,
- different kinds of pesticides (herbicides, insecticides and nematicides) used in agriculture and forestry, and
- pathogenic bacteria and viruses (leakage from septic tanks, sewerage systems, animal sludge applied to farm land, etc.).

Various criteria are used for classification of groundwater pollution. The most commonly used classification is based on the extent of pollution – point (local), non-point (diffuse), line – regional or continental; kind of pollutants – physical, chemical, biological, radioactive; and source of pollution – urban, industrial, agricultural, traffic, mining, land waste disposal. An overview of groundwater pollution sources is given in Table 13B.3.

Among the most frequent point pollution sources with impact on groundwater quality are industrial sites, mining areas and waste disposal sites. According to the EEA (1995) the potential pollution of groundwater by point sources does not cover more than 1% of the European territory. However, point pollution sources mostly occur close to urban areas or rural settlements and have serious impact on the quality of public or domestic groundwater supplies. Oil products, heavy metals and various organic compounds are the most prominent pollutants of the groundwater system.

Urban and rural areas are sources of multi-point and heterogeneous pollution of groundwater. Insufficient handling, treatment and management of household wastes and waste waters, industrial effluents, uncontrolled waste disposal sites, rain and melt waters and salt water intrusion in coastal areas are the main sources of multi-point pollution of municipal groundwater (Jackson et al., 1980; Mathess 1982; Vrba, 1985; RIVM 1992 and others). In rural areas the most frequent source of groundwater pollution are unsewered sanitation systems, which may cause degradation of groundwater quality in domestic and public water supply wells. Pathogenic micro-organisms, chlorides, nitrates, household detergents and disinfectants are the main pollutants of groundwater in rural settlements.

The only widely occurring groundwater pollutants reported in respect to non-point or diffuse pollution are nitrates originating from organic and inorganic fertilizers applied to arable land. In Europe and United States of America the groundwater quality is much more affected by diffuse nitrate pollution, than in other continents. In several areas with intensive farming activities the nitrate levels in shallow aquifers are above 50 mg/l. According to Stannes and Bourdeau (1995) the nitrate content reaches EU target value (25 mg/l) on 87%, and drinking water standard (50 mg/l) on 22% of the shallow aquifers under agricultural soil in Europe. In

Table 13B.3 Sources of groundwater pollution

Extent of pollution	*Source of pollution*	*Main pollutants*
Point	Industry	Heavy metals (Pb, Zn, Cd, Cr), arsenic, phenols, petroleum products and additives, high BOD, suspended solids, PAHs, synthetic organic and organometalic compounds
	Mining	Heavy metals, salts (chloride, sulphate), low pH, high TDS, cyanide, PAHs, petroleum products
	Waste disposal sites, including deep disposal wells	Heavy metals, ammonium, sulphate, chloride, phenols, various biodegradable and non-biodegradable organics, faecal pathogens
	Radioactive waste	^{3}H - Tritium, ^{90}Sr, ^{137}Cs, ^{239}Pu, ^{129}I, ^{226}Ra, toxic metal
	Animal husbandry	High suspended solids, BOD, total nitrogen, chloride, faecal pathogens
Multipoint	Urban areas	Heavy metals (Pb, Zn), ammonia, chloride, sulphate, petroleum products, chlorinated hydrocarbons, surfactants
	Rural settlements	Ammonia, nitrate, chloride, sulphate, surfactants, iron, manganese, faecal pathogens
	Military areas	Petroleum products, heavy metals
Non point (diffuse)	Agriculture crop and root-crops farming, cattle breading, irrigation	Fertilizers (organic and inorganic): nitrate, ammonia, chloride, phosphate, sodium, potassium, faecal pathogens, salinity Pesticides: organochlorine compounds (aldrine, heptachlore, carbamate insecticides (atrazine), polyphosphate, organometallic compounds (fungicides)
Line	Roads	High suspended solids, salts, petroleum products, solvents
	Railways	Petroleum products, organic chemicals
	Oil pipelines	Petroleum products
	Sewerage systems	High suspended solids, nutrients, chloride, high BOD, faecal pathogens
	Streams	Nitrate, ammonia, iron, manganese, phenols
Regional (continental)	Acid deposition	Aluminium, low pH, nitrates, sulphates
Coastal areas	Salinization	Sodium, magnesium, chloride, sulphate, high salinity and TDS

United States of America, particularly in the mid-continental United States Corn Belt, where nearly 60% of nitrogen fertilizers of the whole United States of America are applied, high aquifer pollution (150 mg/l NO_3-N) can be found in many regions (Spalding and Exner, 1991). A NO_3-N content from 40−60 mg/l is also reported in groundwater in several irrigated valleys in California and other irrigated areas of the United States (Keeney, 1986). However, high contents of nitrates in shallow wells are mostly a consequence of poor construction and siting of wells located closely to animal corrals and cattle feeding areas.

Acid atmospheric emissions (sulphur dioxide – SO_2, nitrogen oxides – NO_x) are transported hundreds of kilometres across the continents and their chemically converted products (sulphuric and nitric acids) are potential sources of regional transboundary acidification of the soil and water bodies. The influence on the groundwater quality (lowering of pH values, increasing content of aluminium sulphates and heavy metals) has been recognized in some industrial regions in Europe (Holmberg, 1987; Stanners and Bourdeau, 1995).

Natural processes and man-made factors involving groundwater composition and quality, are described in details by Matthess (1982), RIVM (1992) and others. Section 13B.6 comprises references to selected case studies, as examples of different kinds of pollution and remediation (Al-Agha, 1995; Terao et al., 1993; Ramon, 1993; Vellner, 1993; Hyde et al., 1993; Christe et al., 1995; Fried, 1991; Vrba, 1991; Früchtenicht et al., 1995).

13B.5.2 Point pollution control of groundwater

No generally valid methods for identifying pollutants below the surface can be established, because each pollution episode is highly individual in nature, and the hydrogeological environment differs from case to case. Point pollution control based on mapping, monitoring and modelling (simulation and forecasting of pollutant transport) includes the following steps:

- identification of the pollution source and extent (plume and front); use of remote sensing methods is very effective;
- design and establishment of a site-specific monitoring system and programme;
- determination of the composition, properties, amount and duration of the pollutant spill;
- assessment of hydrogeological conditions and delineation of the risk area.

Point pollution is usually a time-limited problem, due to the possibility of identifying and insulating the pollutant source. Once the source is insulated, and pollution plume defined remediation of pollution can be launched. Decisions for the selection and implementation of clean-up techniques and rehabilitation processes are based on the properties and mobility of the pollutant, the age and extent of the pollution (unsaturated zone only, part or all of the aquifer), and the vulnerability and attenuation capacity of the affected groundwater system.

The most frequent source of point pollution of groundwater, i.e. oil hydrocarbons, can be tackled with various physical, chemical and biological methods. When only the unsaturated zone is contaminated, effective methods are: excavation of polluted soil, or injection through boreholes or galleries of water enriched with oxygen, hydrogen peroxide, ozone or with nutrients that stimulate the growth of autochthonous micro-organisms. In the case of soluble oil compounds (petrol, aviation kerosene) washing the contaminated rock medium with hot water, ventilation by injected air (to remove easily volatile hydrocarbons) or vacuum pumping (to remove light oil hydrocarbons in the liquid phase) are clean-up methods that can be employed successfully.

The remediation methods used for cleaning up a saturated aquifer depend on the intensity of pollution and the solubility of the oil hydrocarbons. When the groundwater table is covered by a layer of oil compounds, scavenge pumping is applied. When the content of dissolved and emulsified hydrocarbons is low, the contaminant is collected from the groundwater surface using a strip-type or mop-type mechanical skimmer. Some other methods that can be used with success for remediation of oil polluted aquifers are: vacuum pumping (in the case of oils with a low boiling point), stripping of light fractions of oil hydrocarbons, biodegradation of oil hydrocarbons in an aerobic medium, anaerobic degradation of aromatic hydrocarbons. It should be emphasised that the clean up of water systems polluted by oil hydrocarbons is always a long, and technically and financially demanding process, with uncertain results.

13B.5.3 Non-point pollution of groundwater

Non-point pollution, also referred to as diffuse pollution, of groundwater is usually related to agricultural activities, particularly the massive application of fertilisers and protective chemicals, or irrigation return flow on farmland. At present the only widely occurring groundwater pollutants, reported in respect to non-point agricultural contamination, are:

- nitrates originating from organic and inorganic fertiliser applied to arable land;
- potassium and phosphorous compounds derived from fertilizers and pesticides that accumulate in soil and the unsaturated zone. Because of their lower solubility in water and lower mobility, they constitute a potential threat to groundwater;
- atrazine, which has been identified in many regions of the world, especially in areas with fruit and vegetable plantations on sandy soils over shallow vulnerable aquifers.

Current fertiliser management practices result in the input of nitrogen significantly exceeding nitrogen uptake. Data from several countries show that only about 50 percent of the nitrogen applied as fertiliser is removed with the produced grain.

Incorrect fertilising regimes also lead to disturbances in the dynamic stability of the soil organic matter, and accelerate mineralization and nitrification processes. Eventually they percolate into the groundwater system as inorganic nitrogen compounds and organic carbon-nitrogen compounds – mainly fluvic acids and saccharides.

Nitrate contents in groundwater bodies are not stable. Short-term cyclic changes in nitrate concentrations depend mainly on natural conditions (particularly precipitation). The long-term upward trend in nitrate contents in groundwater reflects human, largely farming influences (Fig. 13B.1). The vertical movement of the nitrate front to deeper parts of an aquifer is also frequently observed in areas under intensively farmed arable land (Fig. 13B.2).

Pesticide management is one of the most important tasks of agriculture in the environmental sphere. The effects of pesticides on the groundwater system are governed mainly by:

- the kind, properties and amount of pesticides and their potential to leach through the soil to the groundwater,
- the meteorological conditions at the time of pesticide application and the application techniques used,
- the type of vegetation and its ability to take up water through the root zone,
- the soil's physical and chemical properties, its structure and texture, amounts of organic matter and clay material, soil water potential and hydraulic conductivity,
- the vulnerability of the groundwater system, thickness and permeability of the unsaturated zone and type of aquifer (phreatic, semi-confined and confined), and its properties.

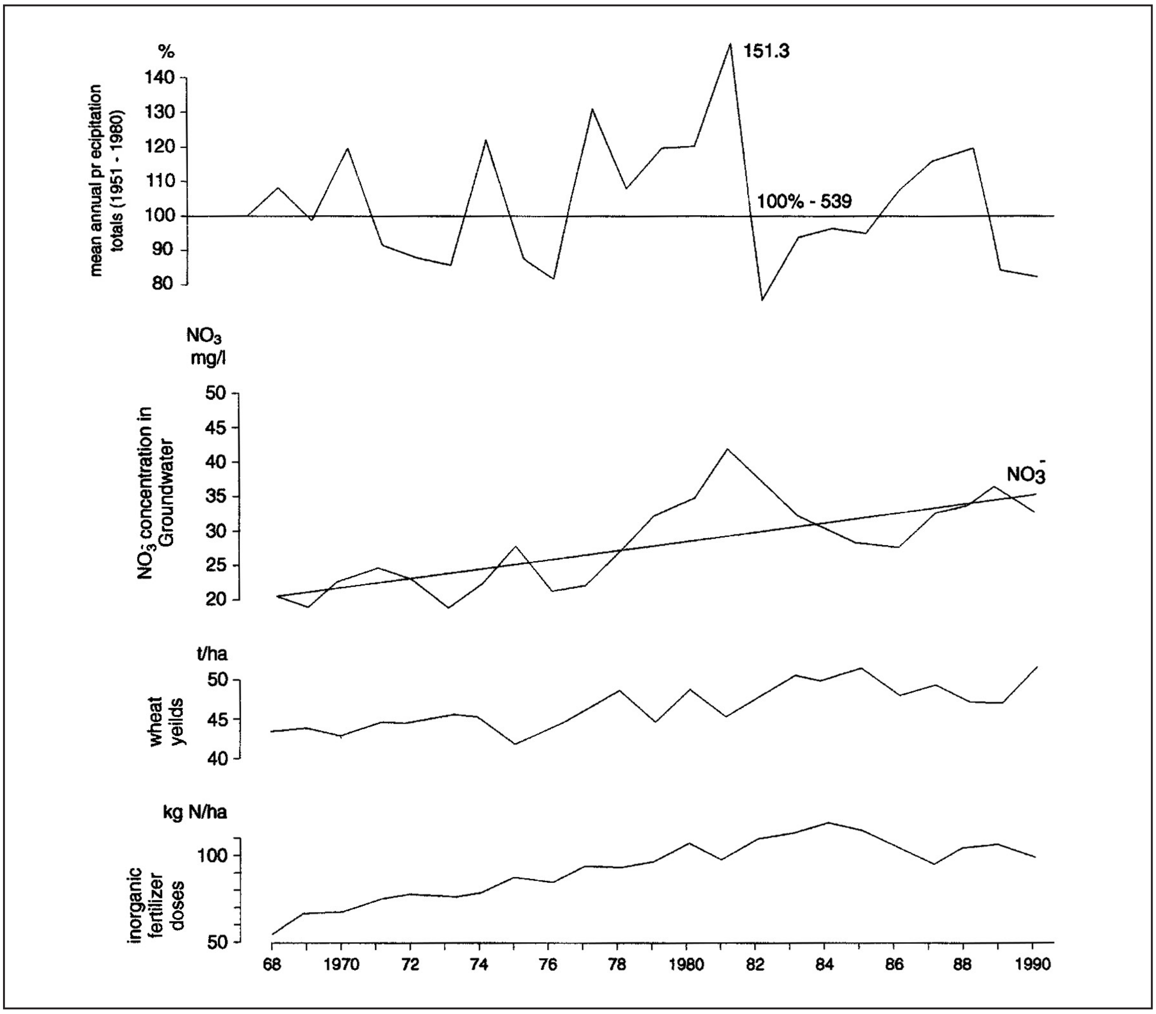
Of special importance is the choice of the pesticide type. Physical and chemical characteristics, toxicity, solubility, persistence, mobility, adsorption, breakdown products, dynamics of residuals and adverse environmental effects should be evaluated when selecting the type of pesticide to be applied.

13B.5.4 Non-point pollution control

Currently, non-point source contamination control systems, and control measures are concerned chiefly with the pollution of groundwater by nitrogen compounds (Jackson et al., 1980; Zwirnmann, 1981; Landreau, 1982; Vrba, 1985).

Because of the great areal extent of diffuse nitrate pollution, the application of 'isolate source policy' and subsurface clean-up techniques are ineffective. The protective measures conducive to improvement of the groundwater quality consist of the management and control of nitrogen input to the plant/soil system. These can be achieved by restricting or prohibiting

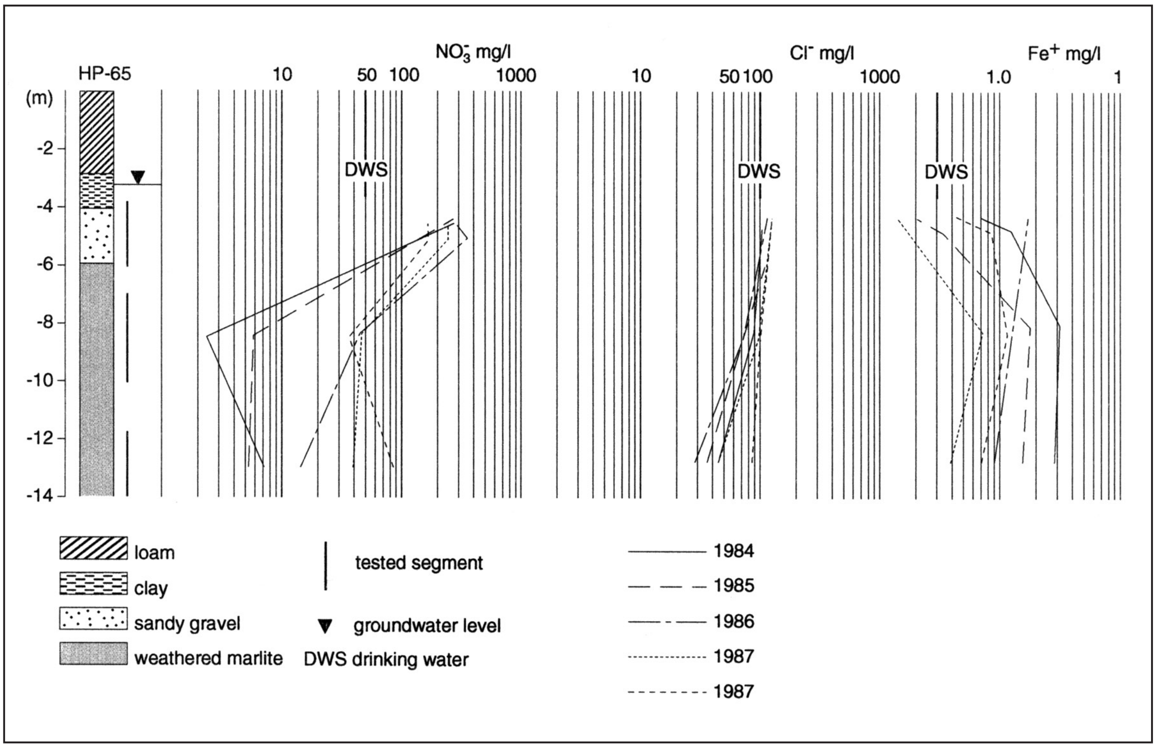
Figure 13B.1 Impact of nitrate pollution on groundwater in the Czech Republic



farming activities in recharge and vulnerable areas of the groundwater system, or by changing agricultural practices. The latter may be related to the selection of suitable fertilisers and determination of the doses, times and techniques of their application; selection of suitable plants (crops); and replacing monocultures by a crop rotation system. Other agricultural activities may have to be modified through regulation of animal production and extension of farmland, or expansion of grassland at the expense of arable land. The control measures depend above all on the steps taken in the agricultural sector. In the sphere of water management the control measures can be focused on symptomatic actions only, not on eliminating the cause of pollution.

A modern approach to pest control lies in the preventive sphere, in the integrated protection of crops against diseases, weeds and pests. The preventive step in integrated protection entails the policy of licensing new pesticides, permanent pest control, soil preparation, selection of plants and sowing procedures, choice of highly resistant kinds of cereals and other crops, and optimal nitrogen and other fertiliser doses. If a disease, pest, or weed has already been identified, the time and dose of pesticide and the technique of its application should be defined with regard to the type of crop. Before the pesticide is applied, the data recorded on weeds and diseases in past years and the qualified forecasts of the potential occurrence and intensity of the disease, pest, or weed in question should be evaluated. When selecting a pesticide, priority

Figure 13B.2 Changes in hydrogeological profiles in a shallow aquifer in 1984–1989; monitoring borehole HP-65 (Central Bohemia, Czech Republic)



should be given to those types that do not produce a need for additional chemical treatment; the synergetic effects of pesticide mixtures should be utilised. Biological and biochemical substances or rapidly degradable types of pesticide (predators, hormone stimulators, pheromones, etc.) should be applied so as to reduce adverse affects on the groundwater system.

13B.5.5 Impact of groundwater on human health

The physical, chemical and biological composition of groundwater may have positive as well as negative effects on human health. Entry of pollutants into a groundwater system through human impacts may be the cause of alimentary diseases, which can grow to epidemic proportions, with potentially fatal consequences. In this respect, groundwater quality is of particular importance, as groundwater accounts for nearly 100 percent of drinking water supply for the population of less developed countries. In advanced industrial countries, as, for instance, in Europe, groundwater usually provides more than 50 percent drinking water. Due to its mobility and ability to transport, transform and absorb pollutants, groundwater is becoming one of the most potentially dangerous contaminating media. It has been reported that in less developed countries, polluted water may cause 80 per cent of diseases.

By comparison with surface water, groundwater's self-purification potential is markedly lower and lessens with the aquifer's depth depending on the declining amount of dissolved oxygen. In spite of that, contact between groundwater and sediments when these have great ion-exchange potential, may cause distinct changes in space and time in their physical and chemical composition, biodegradation and microbial activity.

The adverse effects of pollutants on human health via the food chain, groundwater included, impinge on social and economic spheres (sickness and death rates, migration of population, lower working output, impact on people's mental state, etc.). The health risks posed by different kinds of pollutants in groundwater should therefore be the subject of continuous control and evaluation, because they may assume enormous significance for present and future generations.

Elevated nitrate levels in groundwater are particularly monitored in areas of intensive agriculture. High doses of fertilisers may result in the accumulation of unmetabolised nitrates in plants such as spinach, carrot, kohlrabi and beet which are especially liable. There is a well-known and close relationship between high nitrate contents in groundwater consumed for drinking and alimentary methaemoglobinaemia, which can be fatal for infants under six months of age (through bacterial action nitrates are reduced to nitrites which cause the haemoglobin in blood to change into methaemoglobin, which is unable to transport oxygen). The adverse health effects (gastric cancer, birth defects, cardiovascular disease, effects on the thyroid gland) as a consequence of long-term consumption of water with high nitrate contents are currently being studied. Health risks can also arise when high nitrate contents are combined with pesticides, or when their residues form carcinogenic nitrosamines. Exposure to nitrate in drinking water cannot be implicated or excluded as a causative factor for certain types of cancer (Weisenburger, 1991).

In view of their effects on human health, pesticides should be monitored to ascertain their persistence in soil, ecological-toxicological and neurotoxicological effects, penetration into the food chain, and biotransformation (their intermediate products may be more dangerous for man than the original compounds). At first, pests were combated using salts of arsenic, mercury, copper and zinc, as well as sulphur and barium chloride; later dinitro-compounds, triocianates and, most importantly, chlorinated hydrocarbons, especially DDT, HCH and its isomers lindane, aldrin, dieldrin and heptachlor were introduced. Even low concentrations of the above substances are detrimental to the environment, and chronically toxic. The use of most chlorinated hydrocarbons in farming has been banned because they can cause acute or chronic conditions and pathological processes, and adversely affect the health of future generations. However, their residues in groundwater persist, and have been identified locally and regionally in concentrations of 0.001 mg/l and higher, with polychlorinated biphenyl (PCB) occurring in concentrations up to milligrams.

A broad spectrum of organic compounds is observed in most groundwaters. So far, they have only locally shown up in harmful concentrations in groundwater, in shallow vulnerable aquifers beneath intensively farmed arable land or as a consequence of uncontrolled leakage. Organic compounds of phosphorus pose a risk through their acute toxicity; organic fungicides are a potential source of poisoning and, due to their persistence in soil, cause long-term degradation of the biosphere and hydrosphere; carbamates are toxic substances that are potentially carcinogenic and possess mutagenic and teratogenic properties. Mercurial fungicides are also highly toxic. Triazine, carbamate and similar herbicides pollute groundwater, even at trace concentrations.

The metal contents in groundwater resulting from human activities should be monitored because of their toxic effects on the human organism (its liability to genetic damage), and the ability of these compounds to accumulate in the human body due to their long biological half-life. Cadmium enters the food chain through crops or water. The uptake of cadmium by roots depends mainly on the soil pH. The presence of cadmium in groundwater has been monitored in particular in certain intensively farmed regions. Enhanced levels of lead in groundwater have been identified in areas affected by exhaust gases of dense road traffic and in areas where metallic ores are mined. An increased intake of lead results in anaemia, fatigue and irritability. Reports also mention lead's effect on the growth of long bones during the development of young animals.

Fluorides are frequently present in groundwater and their high content is associated with calcium deficiency. Long-term consumption of groundwater containing more than 2 mg/l of fluorides causes dental fluorosis particularly in children. In concentrations of less than 1 mg/l, fluoride in drinking water reduces solubility of tooth enamel, supports the growth of teeth, and has a beneficial effect on human health.

Uncontrolled spills of liquid human waste from septic tanks and sewerage systems, animal sludges, and fertilising with farm manure (above all pig manure) over shallow, vulnerable aquifers can be a source of bacterial, viral and parasitic contamination of groundwater. Health risks are considerable in such areas, particularly from domestic and shallow wells in rural areas. Diseases can grow into epidemics with potential fatal consequences. Intake of water contaminated in this way may lead to alimentary diseases, above all typhoid, paratyphoid and other salmonelloses, enterovirosis, yersiniosis, etc. The World Health Organization estimated in 1999 that water related diseases caused 3.4 million deaths in 1998, more than half of them children. Other estimates (Van der Hoek et al., 1999) are even higher especially for diarrhoea (5 million deaths).

13B.6 References and additional reading

AL-AGHA, M. R., 1995. Environmental Contamination of Groundwater in the Gaza Strip. *Environmental Geology,* Vol. 25, pp. 109–13.

CANDELA, L.; AURELI, A., 1998. *Agricultural Threats to Groundwater Quality*. UNESCO, CIHEAM, Zaragoza, 251 pp.

CHRISTE, R.; FLURY, F.; HESSENAUER, M.; JEANMIN, P. J.; TURBERG, J., 1995. Groundwater Impact Assessment for the N16 Motorway in Northwestern Switzerland. In: *Groundwater Quality; Remediation and Protection,* IAHS Publication No. 225, pp. 459–70.

EPA, 1985. *Groundwater Monitoring Strategy*. United States Env. Prot. Agency, Washington, 33 pp.

FRIED, J. J., 1991. Nitrates and Their Control in the EEC Aquatic Environment. *NATO ASI series G: Ecological Sciences,* Vol. 30, pp. 3–12..

FRÜCHTENICHT, H.; HELD, T., 1995. *Microbiological* in Situ *Remediation of an Aquifer Contaminated with Waste Oil from a Refinery*. IAHS Publication No. 225, pp. 411–20.

HOLMBERG, M., 1987. Assessing Aquifer Sensitivity to Acid Deposition. In: *Vulnerability of Soil and Groundwater to Pollutants,* Proceedings and Information No. 38, TNO, RIVM, The Hague, pp. 373–80.

HYDE, G.; PEDERSEN, K. E.; ANDERSEN, E., 1993. Aquifer-Wide Groundwater Restoration and Protection; from Concept to Practice, an Example from Denmark. In: *Groundwater Quality Management,* IAHS Publication No. 220, pp. 471–8.

JACKSON, R. E., 1980. *Aquifer Contamination and Protection*. UNESCO, Paris.

KEENEY, O. R., 1986. Sources of Nitrate in Groundwater. *Critical Review in Environmental Control,* Vol. 16, pp. 257–335.

LANDREAU, A., 1982. Fertilisants azotés et qualité des eaux souterraines. In: *Cartographie des zones vulnerables*. B.R.G.M., Orléans.

MATTHESS, J., 1982. *The Properties of Groundwater*. Wiley-Interscience Publications, New York, 406 pp.

RAMON, S., 1993. Experiments on Agricultural Restrictions Carried Out in Protection Areas by the Agence de l'Eau Rhin-Meuse. In: *Groundwater Quality Management,* IAHS Publication No. 220, pp. 433–8.

RIVM, 1992. *The Environment in Europe: a Global Perspective*. Report No. 481505001, RIVM, 119 pp.

SPALDING, R. F.; EXNER, M. E., 1991. Nitrate Contamination in the Contiguous United States. In: *Nitrate Contamination,* NATO, ASI Series G, Vol. 30, pp. 13–48.

STANNERS, D.; BOURDEAU, P., (eds.), 1995. *Europe's Environment.* European Environmental Agency, Copenhagen, 676 pp.

TERAO, H.; YOSHIOKOJ, Y.; KATO, K., 1993. Groundwater Pollution by Nitrate Originating from Fertiliser in Kakamigahara. *Selected Papers on Environmental Hydrogeology,* IAH/Heise, Hannover, Vol. 4, pp. 51–62.

VAN DER HOEK; KONRADSEN, F.; JEHANGIR, W. A., 1999. Domestic Use of Irrigation Water; Health Hazard or Opportunity? *Water Resources Dev.,* Vol. 15, pp. 107–19.

VAN WAEGENINGH, H. G., 1985. Protection of the Groundwater Quality in Porous Permeable Rocks. In: *Theoretical Background, Hydrogeology and Practice of Groundwater Protection Zones. Intern. Contrib. Hydrogeol.,* Ed. Heise, Vol. 6, p 111–21.

VELLNER, L., 1993. *Problems of Groundwater Quality Management in Estonia. Groundwater Quality Management.* IAHS Publication No. 220, pp. 449–60.

VRBA, J., 1985. Impact of Domestic and Industrial Wastes and Agricultural Activities on Groundwater Quality. *Memoirs XVIII IAH Congress,* Cambridge, Part 1, pp. 91–117.

——, 1994. Groundwater Protection Strategy, Policy and Management. *Proc. Congress Spain Group of IAH,* Madrid, Part II, pp. 363–83.

——; PEKNÝ, V., 1991. Groundwater Quality Monitoring; An Effective Method of Hydrogeological System Pollution Prevention. *Environmental Geology,* Vol. 17, pp. 9–13.

WEISENBURGER, D. D., 1991. Panel Consensus on Potential Health Consequences of Elevated Nitrate Levels. In: *Nitrate Contamination; Exposure, Consequence and Control.* NATO ASI Series G: Ecological Sciences, Springler-Verlag, Vol. 30, 511 pp.

WORLD HEALTH ORGANIZATION, 1999. *World Health Report.* Geneva.

ZWIRNMANN, K. H., 1981. *Analysis and Control of Non-Point Nitrate Pollution of Municipal Water Supply Sources*. IIASA, Luxenburg, Austria.

14 Hydrogeology of carbonate rocks

14.1 Introduction

Carbonate rocks as such are usually poorly permeable and have a low specific yield. Only when they are intensely fractured and particularly when the fractures are widened by solution (development of karst phenomena) they do become water bearing and water transmitting, often on a spectacular scale.

People have always been intrigued by carbonate rocks with karst phenomena, particularly caves (Fig. 14.1). In prehistoric times, early man chose the caves in limestone as living areas, and his explorations led him deep underground to sources of water. Even 2000 years ago, Greek and Roman philosophers speculated on the origin and nature of caves, underground rivers and springs The earliest hydrological concepts of the hydrological cycle, the source of water, its occurrence, and quality were related to the hydrology of karstic limestone.

How important are carbonate rocks to man? Approximately 12 percent of the earth's continental surface is underlain by carbonate rocks (Fig. 14.1) that have produced diverse topographical landforms by weathering under varied climatic conditions (Milanovich, 1981). The surface topography of some areas underlain by carbonate rocks is one of broad, rolling plains, but elsewhere it is characterised by steep-sided bluffs, canyons, sinks, and valleys. In some areas, carbonate rocks are overlain by fertile soils; in others areas they are bare.

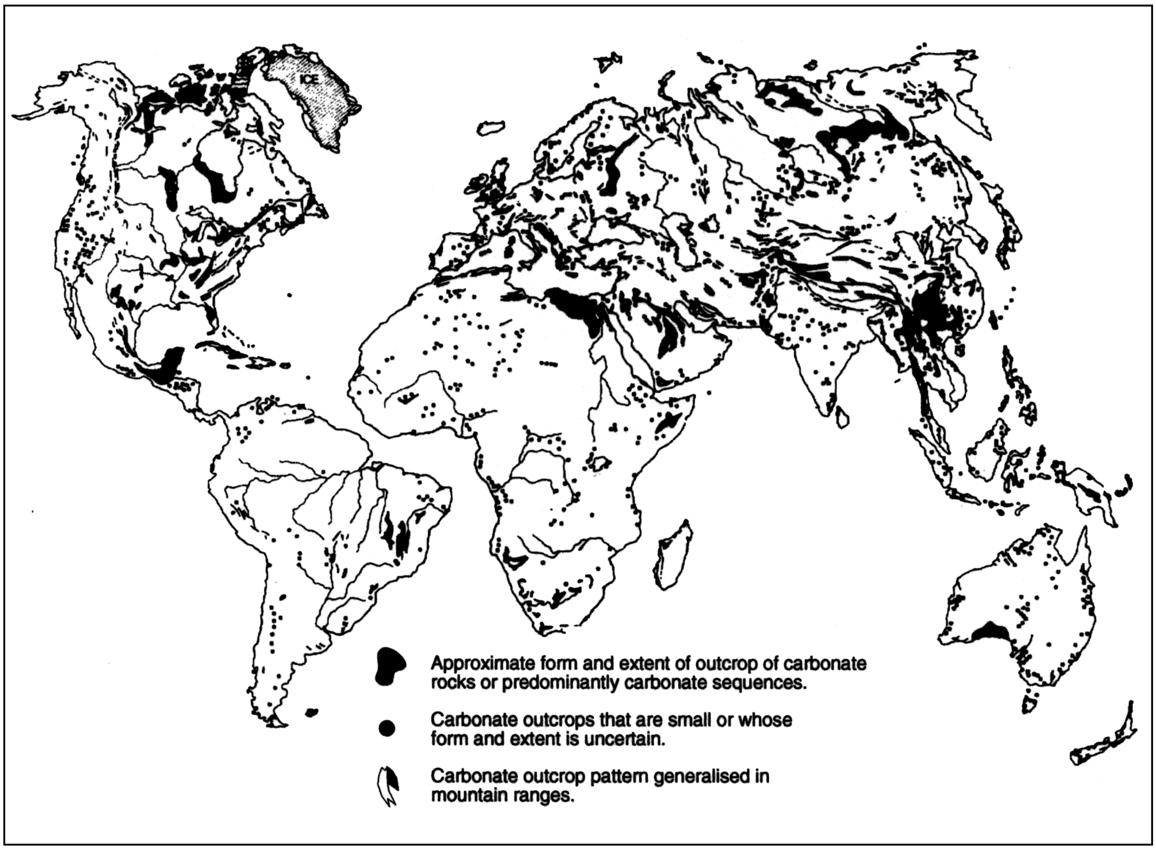
The impact of carbonate rocks is great on man and of substantial interest financially, e.g.:

(i) At the International karst meeting in Guilin, China, in October 1988 (Yuan Daoxian, 1988), the Director of the U.S. Geological Survey stated that karst aquifers are a major source of drinking water in the United States of America, providing 25 million cubic metres of water per day in 1985 (Peck et al., 1988).

(ii) In the midwest of the United States of America, a large area underlain by limestone is covered by a very productive, rich soil that is farmed to produce large quantities of food. This area is literally 'the breadbasket of a nation'.

(iii) John Newton (1986) reported that from 1950 to the early 1980s more than 6,500 sinkholes or related features appeared in the eastern United States. He further stated that the total cost of damage and associated protective measures resulting from these induced sinkholes is unknown, however, at five dam sites alone the repair costs were already in excess of US$ 140 million.

Figure 14.1 Location map of the major outcrops of carbonate rocks (After Ford and Williams, 1989)



(iv) A Panel on Land Subsidence of the U.S. National Research Council, in 1991, related that six states had individually sustained in excess of US$10 million from damage resulting from sinkholes, and an additional four states sustained from US$1 to 10 million in damage from the same cause. As a result, 'awareness programmes' for catastrophic subsidence areas, and insurance programmes applicable to sinkhole problems have been developed in the United States of America.

(v) Finally, carbonate rock or karst areas are dynamic and environmentally sensitive. The geological structure, the solubility of the rocks involved, and the climatic conditions determine to a large degree how rapidly changes can take place. Therefore, it is necessary to recognise the synergistic relation between circulation of water and solution of the rock. The solution leads to progressive lowering of water tables that will result in base level changes, progressive cave enlargement, and changes in karst topography that can all take place within a relatively brief period of time and bring about major environmental problems. One of the first studies relating geography, geology, mining, and resources and their development and impact on man and his environment was published as an atlas series for northern Alabama, United States of America (LaMoreaux, 1975).

Carbonate rocks are a source of abundant water supplies, minerals, and oil and gas. However, there are also many problems related to developing adequate water supplies, assuring proper drainage, providing stable foundation conditions, and preventing serious pollution problems. Because of the complexity of carbonate rocks, it is impossible to base the evolution of concepts related to the movement and occurrence of groundwater, methods of exploration and development of water, safe engineering practices in construction of all kinds, and adequate environmental safety precautions on one uniform set of rules.

The complexity of hydrogeological systems in carbonate rock terrain, requires that a thorough hydrogeological study be carried out to determine whether a specific site is, or can be rendered suitable, e.g. for construction, for highway location, for a water supply, or for a land disposal facility. Important components of hydrogeological studies are:

- field mapping of structural and stratigraphical units;
- interpretation of sequential aerial photographs;
- test drilling and geophysical analyses;
- fracture analyses;
- monitoring seasonal variation in water levels and in water quality;
- establishing the spatial variation of hydraulic characteristics of the aquifer and associated aquicludes, and the velocity and direction of movement of groundwater within the aquifers;
- determination of the controls for recharge, discharge, and local base level;
- evaluation of the effects of man's activities, such as pumping, dewatering and construction.

14.2 Factors determining the occurrence of groundwater in carbonate rock

The hydrogeology of carbonate rocks can only be understood by careful observation of the physical characteristics and distribution of the rocks. The fundamental geological approach to the exploration, use, and conservation of water involves considering:

- the composition of the rocks,
- the shape of their units,
- the variation of the composition of the unit throughout its extent,
- the sequence of units and how it varies,
- the deformation of the sequence and the influence of the deformation on the characteristics of the rocks,
- the units, the sequence, and the topographical and physiographical situation These characteristics are discussed in the following sections on structure, fracture systems, joints, hydrogeological features, porosity, permeability, groundwater flow, and hydrochemical character.

14.2.1 Structure

The structural and tectonic history of a region plays a significant role in determining the behaviour of carbonate aquifers. The range of variation among these many interdependent and independent factors is so wide that no two carbonate aquifers have identical characteristics. The influence of geological structure on aquifer characteristics is also variable, so that it is not possible to provide a set of simple rules governing the influence of joints, faults and folds on groundwater movement or availability. Thus in carbonate rock areas, it is imperative to develop a conceptual framework to unravel the structural controls of groundwater movement for a particular study area.

The tectonically induced position of a body of carbonate rocks determines potential recharge/discharge relationships. Folding, faulting, and fracturing of the rocks often determine the porosity and permeability of the aquifer, as well as specific directions of groundwater flow through the carbonate system. After a recharge/discharge system has been established, permeability of the carbonate rocks must increase with time if the transmitted groundwater is unsaturated in calcium and/or magnesium carbonate. Therefore, karst development is a progressive process although it's probably not cyclic in the sense described by Penck (1900) or Davis (1901). However, the characteristics of carbonate aquifers continue to change as long as an open groundwater system exists.

14.2.2 Fracture systems

In carbonate rocks that do not have a primary intergranular permeability, joints or fractures are essential for initiation of downward percolation of water (Stringfield et al., 1979). The lateral, as well as vertical routes along which groundwater flow is channelled prior to solution modification, may also be controlled by fracture patterns (Kiersch and Hughes, 1952). Research performed throughout the world has documented the critical importance of fractures in controlling groundwater movement in carbonate aquifers of Paleozoic and Mesozoic age. In fact, the relationship between fractures and permeability developed by solution in carbonate rocks is so well documented that one almost presumes the presence of fractures if solution cavities exist in Paleozoic or Mesozoic limestones. Bedding planes may also provide avenues for groundwater movement (Palmer, 1977), but even in such cases, movement of groundwater between bedding planes is often fracture controlled. Tertiary and Quaternary limestones frequently have permeability due to solution development of their primary textural porosity (Stringfield, 1966), but research by Vernon (1951) and Leve (1983) has demonstrated that in the Floridan aquifer groundwater movement is locally influenced and modified by the presence of faults.

If a favourable recharge/discharge relationship exists for a carbonate unit, then water can gain access to that unit and move through it. The groundwater movement will gradually enlarge the fractures through which it moves by solution, and permeability of the unit will increase with time.

Moore (1966) has postulated that pumping, oscillations from earth tides and distant earthquakes may move water through joints and partings too narrow to allow much movement of water under normal hydraulic gradients. Fractures in carbonate rocks commonly have an uneven distribution, consequently groundwater movement often develops secondary permeability in selective areas of closely-spaced fractures or along open fracture systems. Therefore, according to LeGrand et al. (1971), moderately large openings tend to enlarge by solution action while small openings not in the path of water flow generally enlarge only slightly. The net result is a typically uneven distribution of permeability in aquifers where permeability is controlled by fractures.

14.2.3 Joints

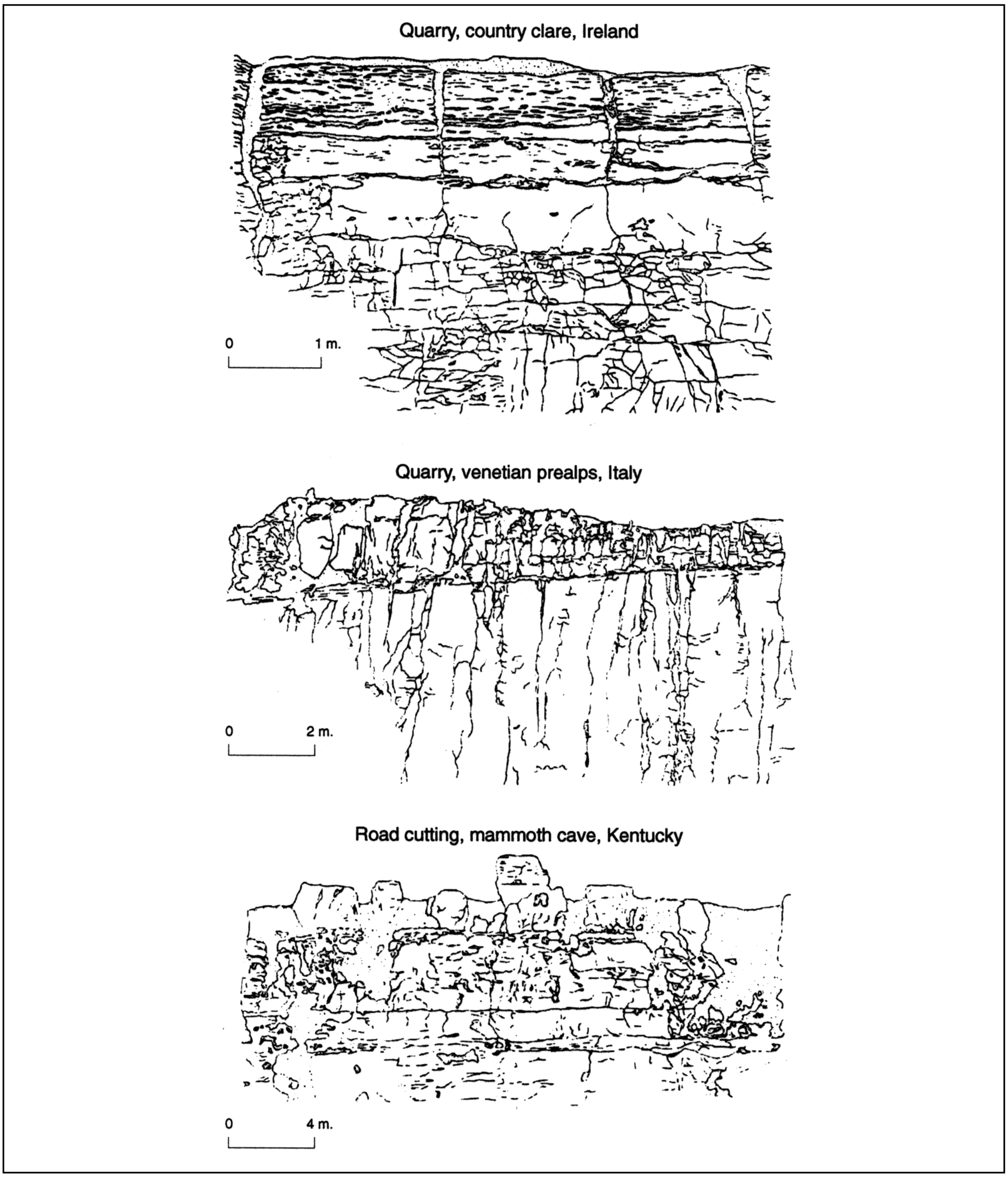
Joints are the most frequently occurring fractures found in carbonate aquifers. Typically, they are nearly perpendicular to bedding. However, depending on conditions of formation, joints may have any orientation with respect to bedding. For example, Grice (1968) found that joints near Grand Rapids, Manitoba, Canada, were essentially parallel to bedding planes.

Joints may exert primary control over the direction of groundwater flow prior to development of solution cavities in a carbonate aquifer. Therefore, non-vertical joints may exert a considerable lateral influence on the movement of groundwater. Therefore, use of rose diagrams or histograms to represent the orientation of joints may result in considerable error unless these data are carefully used to estimate groundwater flow directions.

If recharge/discharge relationships allow groundwater to move through an open carbonate aquifer, then joints can become enlarged by solution of the limestone. The result may lead to development of: (1) a very permeable and cavernous, unsaturated zone; (2) a zone of an exceptionally high permeability in valleys; and (3) a rapidly decreasing permeability with increasing depth below the water table, as postulated by LeGrand et al. (1976).

Figure 14.2 contains three examples of joint enlargement by solution. It also illustrates movement of water along bedding planes and an apparent rapid decrease in permeability with depth. Stearns (1977) has demonstrated that the openings in the middle Ordovician limestones of the Percy Priest Dam area in central Tennessee are a box-work of bedding planes and joints.

Figure 14.2 Vertical section through limestone outcrops showing diminution of fissure widths with depth (From Williams, 1983)



The porosity of the limestone is approximately 15 percent in the upper 3 m but decreases rapidly to about 1.5 percent at a depth of 10 m.

14.3 Hydrogeological features of carbonate rocks

The primary porosity of carbonate rocks is the result of open spaces in the rocks that have persisted throughout the period of deposition, diagenesis and lithification. Textural porosity and intergranular porosity are common terms used as synonyms for primary porosity. Secondary porosity is a term used to express the amount of open space in a rock that has been created by post-lithification processes such as fractures (joints, faults, parting) or solution cavities. To be able to solve problems in carbonate aquifers, the porosity characteristics of these aquifers must be thoroughly understood. One of the textbooks on this subject is *Guide to the hydrology of carbonate rocks*, published by UNESCO (LaMoreaux et al., 1984).

Hydrogeological parameters such as porosity, storage coefficient, permeability, and leakance must be determined in carbonate aquifers and related to the structure of the voids and fissures that predominate. To assess the water resources of an area the process of evaluation must use a variety of geological, chemical, engineering and mathematical disciplines

The principal characteristics of carbonate rocks, namely their heterogeneity and anisotropy, determine the porosity and permeability parameters at the spatial scale at which these parameters are studied. Results of laboratory and field tests represent only certain local values and cannot be generalised to the entire aquifer without considering the geological characteristics such as lithostratigraphy, structure, and geomorphology (Milanovich, 1981; White and White, 1989; LaMoreaux et al., 1989; Burger and Dubertret, 1975).

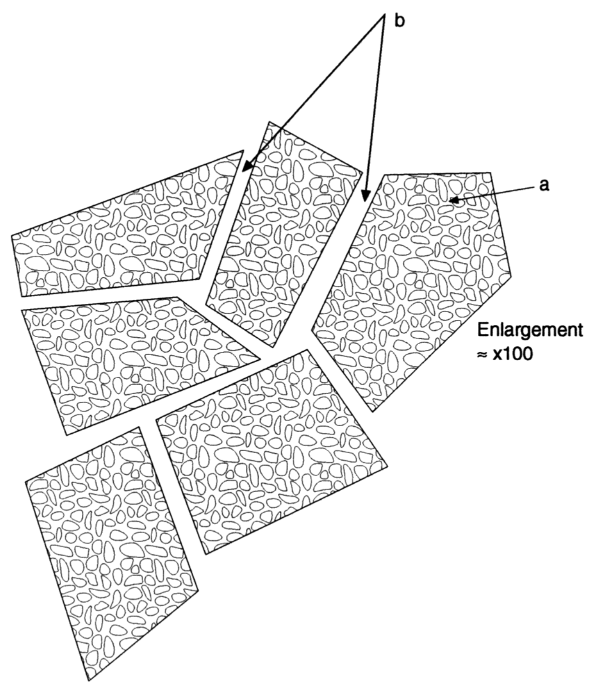
14.3.1 Porosity

Porosity is the property of a rock having voids or interstices. The types of porosity of unconfined or confined aquifers can be classified on the basis of the nature and proportion of voids, interstices, microfissures and channels in the rock and in terms of the amount of groundwater released gravitationally or retained. The characteristics of the voids, their shape, size, distribution, and volume, compared to the overall rock volume are the essential features necessary for defining the porosity types. On the basis of void types, porosity may be microscopic or macroscopic (Table 14.1 and Figure 14.3).

Table 14.1 Classification of voids and porosity

Reference scale	*Void types*		*Porosity types*
Microscopic	Pores or interstices	Intercrystalline	Intercrystalline porosity
	Pores or interstices	Intergranular	Interstitial porosity
	Microfissures	Joints Microfissures	Microfissure porosity
Macroscopic		Channels Cavities	Channel porosity

Figure 14.3 Porosity types in soluble carbonate rocks: (a) rock porosity; (b) microfissure porosity



14.3.2 Permeability

Permeability depends upon the void characteristics, mainly their shapes and opening sizes. There are several types: intrinsic permeability, fissure permeability, and channel permeability, but the hydrodynamic phenomena are the same. Nevertheless, it is convenient to distinguish between the rock or intrinsic permeability (microscopic) of the rock mass and the regional or formational permeability (macroscopic). The latter includes the added permeability supplied by joints, fissures, and channels. Rock's intrinsic permeability is very small compared to the regional or formational permeability.

Similar to the porosity, the permeability varies with time, and it usually increases with the increase of the size of openings of the fissures. Therefore, an initial permeability and a transient permeability resulting from the widening of the voids, may be recognised (White and White, 1989; Milanovich, 1981; LaMoreaux et al., 1984).

14.3.3 Groundwater flow

Considerable progress in understanding the hydrology of carbonate rocks has been made in recent years. However, some investigators still disagree on the nature of the occurrence and movement of water in carbonate rocks. Most investigators of groundwater in carbonate rocks identify a zone of saturation (phreatic water) with either a water table under non-artesian conditions or a piezometric surface under artesian conditions. A water table concept is useful where the permeability is so poor that the water table is discontinuous or so great that water moves through large artery-type openings. Special emphasis is placed on carbonate rocks as conduits for the transmission of water because only where they act as conduits are they in contact with moving water, and therefore, vulnerable to solution. A conduit containing water in a carbonate rock may be under either water table or artesian conditions; in both cases, three things are essential. It must have (1) an area of intake, and it must (2) transmit, and (3) discharge water. If any one of these three requirements is not met, the limestone body is hydrologically inert and cannot act as a conduit for water (LeGrand and Stringfield, 1966).

Theoretically, groundwater moves in arcuate paths following lines of flow that have their origin at the top of the zone of saturation in recharge areas (Swinnerton, 1942). The flow lines curve downward for some distance and then rise to an outlet or point of discharge. Diagrammatically, these lines can be represented by a family of curves, most closely spaced near the area of the outlet and becoming more widely spaced along the top of the zone of saturation with increasing distance from the outlet. The water may be expected to have this arcuate pattern of flow in uniformly permeable material where geological structure has only subordinate influence on its direction of movement. However, even in the initial stages of groundwater circulation in limestone and other carbonate rocks, lines of flow are modified. Circulation in the discharge area, having greater velocities, may result in an enlargement of the outlet and a consequent shallowing of the more arcuate paths. Solution openings along the more direct paths will become larger than those along the less direct paths and will permit progressively larger flows at the expense of other passageways.

14.4 Examples of groundwater flow systems in carbonate rocks

14.4.1 Karst hydrological systems

The functions of a karst hydrological system are controlled by stratigraphical features and geological structure. Its extent varies from 1 km^2 to thousands km^2. The input of the system may include direct inflow of precipitation through natural shafts, infiltration of precipitation through epikarst zone and soil, and allogenic water from non-soluble rock areas nearby, and condensation of water in caves In some cases, the cave condensation rate could be 0.5 l per m^3 cave volume per day, but depends on many local factors. The output of the system could be a spring outlet or an underground stream, underflow beneath a sediment plain, submarine spring, etc. A clear and quantitative understanding of the karst hydrological system must be the basis for studying groundwater in soluble carbonate rocks.

The prerequisites for identifying a karst hydrological system are the comprehensive mapping of:

- the stratigraphical relation between carbonate rock and insoluble rock;
- the pattern of geological structure;
- the relief and surface drainage system (river, lake, sea, glacial, etc.);
- the relationship with alluvium and other aquifers,

For instance, in Figure 14.4, pattern 1 is a tight folding of carbonate rock alternating with insoluble rock along deep gorges, giving a series of karst hydrological systems in a narrow belt with deep water circulation and small catchment area. Pattern 2 shows a gently folding hydrological system that enjoys a broader catchment area. The water table is shallow because it is perched on an impervious layer. Nevertheless, there is a deep gorge.

The basic approaches for studying the hydrology of carbonate rock areas are as follows:

- mapping of the geomorphology of the hydrologically important surface karst features, such as polje, doline, sinkhole, etc.;
- survey of caves and geophysical prospection for siting underground conduits;
- laboratory porosity analysis, applicable for young porous carbonate rocks;
- hydrological, hydrochemical and isotopical response of the system to storm (Fig. 14.5).

Figure 14.4 The relationship between stratigraphy, folding pattern, relief and karst hydrological system; 1: tight folding, 2: gentle folding

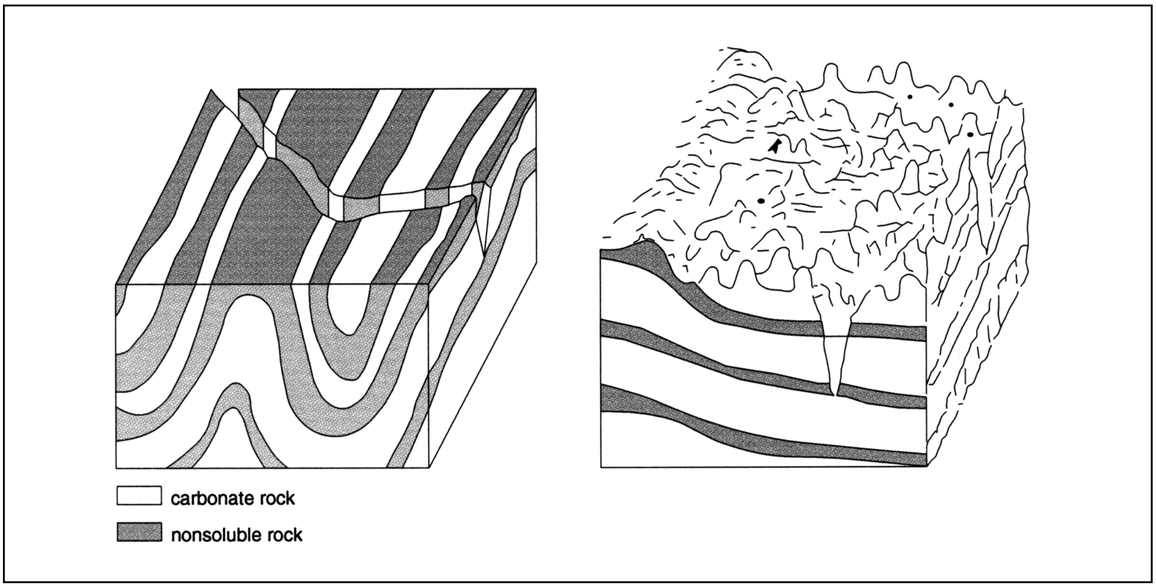
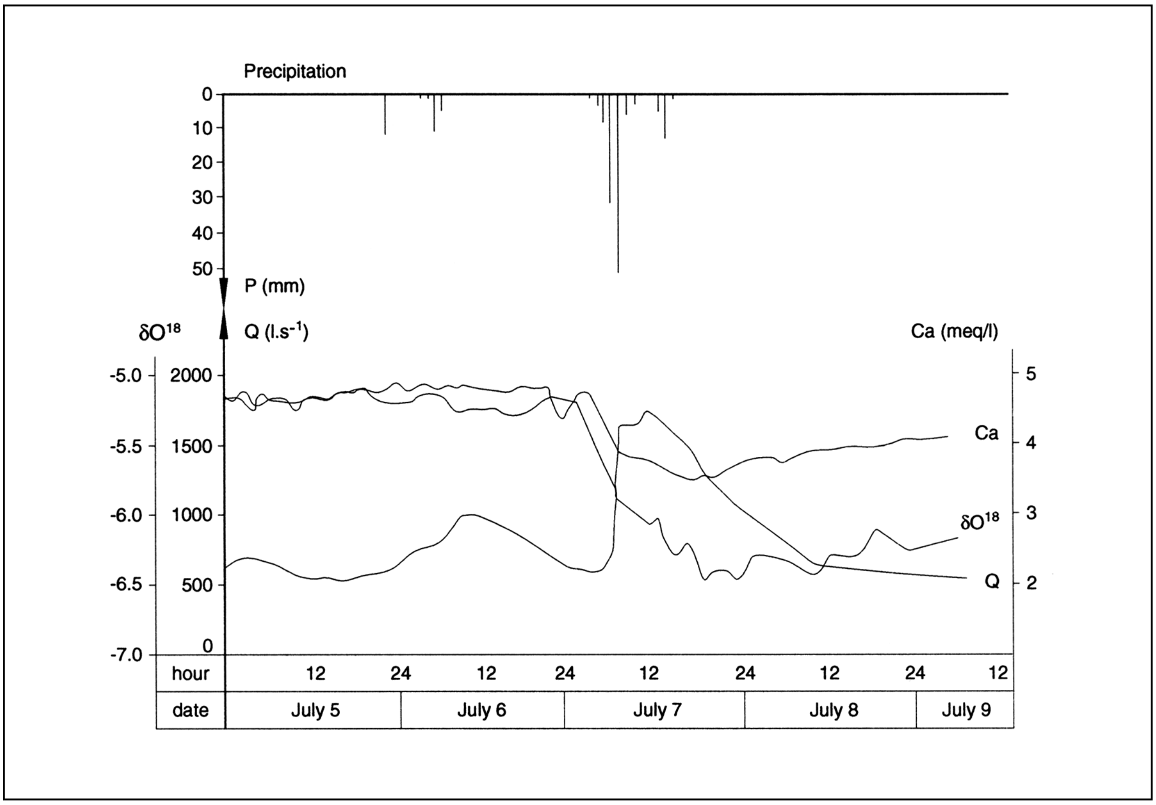


Figure 14.5 The hydrological, hydrochemical and isotopic response to a storm of S31 karst spring near Yaji, Guiin, China 1986 (Yuan Daoxian, 1988)



14.4.2 Characteristics of karst hydrological systems

Only a brief statement regarding the physical, hydrological, and geomorphological aspects of carbonate rocks can be provided here. What all karst terrain have in common is that they are heterogeneous, complicated, and dynamic. To summarise, a table of characteristics is provided as an aid to understanding any type of development in a carbonate rock area (Table 14.2).

The nature of underground circulation depends largely upon the geological structure and the relation of permeable rocks to groundwater discharge areas. Insofar as circulation is concerned, a carbonate formation or body at any one place possesses at least one of the following four types of hydrological zones (LeGrand and Stringfield, 1966):

Zone 1

Carbonates at or near the surface: The water table occurs in the karst rock. Water from precipitation moves vertically downward to the water table and then laterally toward a surface stream.

Zone 2

Carbonates buried beneath an impermeable bed, forming a homoclinal artesian system: under artesian pressure water from a higher area moves through the rock towards a lower discharge area.

Zone 3

Carbonates with no significant discharge facilities: either (a) a homoclinal artesian system so deeply buried beneath impermeable beds that almost no water can escape, or be (b) carbonates lying below the stream controlling the base level of erosion and perhaps so faulted or folded as to nearly preclude the discharge of water.

Zone 4

Denuded carbonates elevated above subjacent valleys and sufficiently impervious locally to preclude a continuous zone of saturation: no subsurface discharge of water occurs. This zone is absent in many areas.

Although both the occurrence and movement of groundwater in carbonate rocks are related to geological structures, the relationships are not direct and exclusive. We cannot rightfully place such geological structures as synclines, anticlines, monoclines, and faults into separate categories for an appraisal of their relation to the occurrence and movement of groundwater. In some areas, faults may serve as avenues for movement of water, and in others they may serve as barriers. An acceptable study technique requires the geological framework to be studied in relation to the superimposed hydrological continuum. Is the hydrological continuum real and significant for a particular setting? The answer may be negative if one of the following conditions applies: (1) insignificant recharge facilities, (2) insignificant discharge facilities, (3) insignificant permeability in the system, and (4) insignificant hydraulic head in the system.

Where carbonate bodies are thin and are compartmentalized by other rocks in complex structural settings, the hydrology is commonly similar to that of the enclosing rocks. Under these conditions some aspect of (a) discharge, (b) movement through the rock, or (c) recharge is likely to be so restricted that a carbonate rock does not act as a separate hydrological system. Such microcompartments, in which circulation of water is so slight as to restrict karst development, should be distinguished from macrocompartments in which the breadth of carbonates in a water-table circulation system may be tens of kilometres. Where carbonates occur as macrocompartments, many features of karst hydrology may develop; yet, the hydrology will be influenced by the hydrological and topographical features of the enclosing foreign beds.

Karstification occurs where circulation of water is not impeded. The circulation of water is not evenly distributed in carbonate rocks, therefore the solution-developed openings representing major permeability features of karst rocks are also unevenly distributed.

Compartments of groundwater in carbonate rocks may come about as a result of aquifers

Table 14.2 Important characteristics of carbonate rocks (Hughes et al., 1994)

Stratigraphy	*Geological Structure*	*Geomorphology*
(Regional and local)	(Regional and local)	(Regional and local)
Stratigraphical column	Nearly horizontal bedding	Relief slopes
Thickness of each carbonate unit	Tilted beds	Density of drainage network
Thickness non-carbonate interbeds	Homoclines	Characteristics of streams
Type of bedding	Monoclines	Drainage pattern
Thin	Folded beds	Dendritic
Medium	Anticlines	Trellis
Thick	Synclines	Rectangular
Purity of each carbonate unit	Monoclines	Other
(Limestone or dolomite)	Domes	Perennial
Pure	Basins	Intermittent
Sandy	Other	Terraces
Silty	Fractures	Springs and/or seeps
Clayey	Lineaments	Lakes and ponds
Siliceous	Locations	Floodplains and wetlands
Interbeds	Relationships with:	Karst features - active
	Geomorphic features	historic
Overburden	Karst features	Karst plains
(Soils and sub-soils)	Stratigraphy	Poljes
Distribution	Structural features	Dry valleys, blind valleys
Origin	Joint system	Sinking creeks
Transported	Joint sets	Depressions and general
Glacial	Orientation	subsidence
Alluvial	Spacing	Subsidence cones in
Colluvial	Continuity	overburden
Residual	Open	Sinkholes
Other	Closed	Roof-collapse
Characteristics and variability	Filled	Uvalas
Thickness	Faults	Caverns, caves and caveties
Physical properties	Orientation	Rise pits
Hydrological properties	Frequency	Swallow holes
	Continuity	Estavelles
Hydrology	Type	Karren
Surface water	Normal	Other
Discharge	Reverse	Paleo-Karst
Variability	Thrust	
Seasonal	Other	**Climate**
Gaining	Age of faults	Precipitation
Losing	Holocene	Seasonal
Groundwater	Pre-Holocene	Annual
Diffuse flow		Long-term
Conduit flow	**Activities of man**	Temperature
Fissure flow	Construction	Daily
Recharge	Excavation	Seasonal
Storage	Blasting	Annual
Discharge	Vibration	Long-term
Fluctuation of water levels	Loading	Evapotranspiration
Relation of surface water and	Fill	Vegetation
groundwater flow	Buildings	
	Changes in drainage	
	Dams and lakes	
	Withdrawal of groundwater	
	Wells	
	Dewatering	
	Irrigation	

being intercepted by impermeable or insoluble materials because of faulting or folding. Some compartments occur in broad carbonate aquifer systems that are close enough to the ground surface to be within the meteoric water circulation system. A good example of this type of groundwater occurrence is present in the gold mining area of South Africa near Pretoria (Enslin, 1967; Buttrick, 1992). In such a system, a drainage network with a dominant line of flow lies on the piezometric surface and depresses it. These drainage lines may be perennial surface streams into which water from the carbonates discharges, or, in the case of mature karst areas where no perennial surface drainage occurs, they may be lines of concentrated groundwater drainage that lie at the base of air-filled openings, depressing the water table and leading to points of groundwater discharge. These compartments are separated laterally by groundwater divides, which do not necessarily coincide with topographical divides on the land-surface. Vertically below they merge into the overall carbonate system.

Lack of uniform permeability may be attributed to a combination of the following conditions. Separate parts of carbonate formations may have had (1) different original porosity-permeability characteristics, (2) different opportunities for groundwater to circulate in them, (3) different degrees of access to water undersaturated in calcium carbonate and (4) different opportunities for erosion or, conversely, for preservation (Stringfield and LeGrand, 1966).

Karst hydrological systems can be divided into two types: (*i*) exposed karst and (*ii*) buried karst (carbonate rock aquifer buried under impervious formations), or covered karst (carbonate rock aquifer covered by loose sediments). Land surface topography has an ever present influence on the position of the water table. Combinations of topographical and permeable conditions control the positioning of the water table in important practical ways. The following general statement forms a basis for useful generalisations: 'A water table deep below land surface occurs where high relief and very permeable limestone are combined'. Local variations of considerable magnitude in permeability are common in carbonate terrain. They contribute to the disappearance of surface streams into underground courses and to their reappearance at other places.

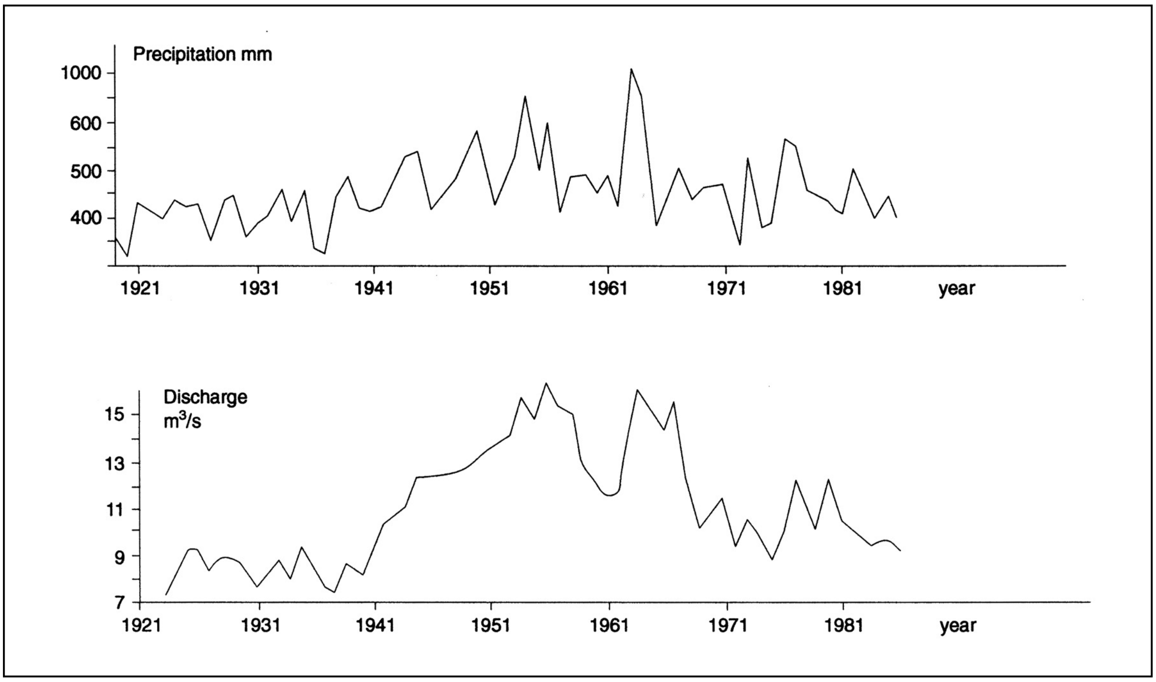
The theory of groundwater motion in small drainage basins, which has been described by Toth (1962, 1963), has application to broad carbonate regions on which local discharge areas are superimposed. Where these compartments occur it is wise to evaluate the most likely directions of flow of water. In many cases more than 90 percent of recharge is shunted out into one of the compartments, leaving little water to reach the lower and major carbonate system. The lack of opportunity for water to move out of the lower part of the carbonates allows this part to stay full of water and to repel most of the available recharge. A conceptual model that considers compartments of carbonate hydrology helps when evaluating the distribution of the flow of water and of the contaminants this water may contain.

Where an artesian circulation system occurs, base-level considerations may be less significant because the piezometric surface is above the aquifer. Zones of greater circulation and solution may develop, perhaps near the top of the confined carbonate aquifer, but these zones are not closely related to base level. Development of secondary permeability in an artesian carbonate system may have considerable overall importance but is slower and less dynamic than the development of secondary permeability in a water table carbonate aquifer. It should be noted, however, that some carbonates in the artesian system were at some earlier time in a water-table circulation system.

Under artesian conditions in inclined aquifers, water may move down dip where: (1) the aquifer is overlain by relatively impervious beds, (2) the relative positions of the recharge and discharge areas are favourable, and (3) the hydraulic gradient is in the same direction as the dip (Stringfield and LeGrand, 1966). Such conditions are not widespread because much water follows arcuate courses to concentrated discharge areas, and locally the direction of this water movement cannot coincide with the dip. Where the permeable zones connecting recharge and discharge areas are not homoclinal, as in folded aquifer systems typical of carbonates of the southern Appalachians (United States of America), water may move down dip on one limb of the structure and up dip to some degree on the other limb.

To obtain the most reasonable assessment of water resources in exposed karst, the time series of precipitation data should be as long as possible, as this is helpful not only for a more accurate assessment of the water resources, but also for a better understanding of the hydrological function of the karst system. For instance, the 60-year-long records of the biggest karst spring of North China, the Nyangziguan Spring, show 20-year cycles of dry and wet periods (Fig. 14.6) and a 3-year time lag of spring discharge records compared to the precipitation records.

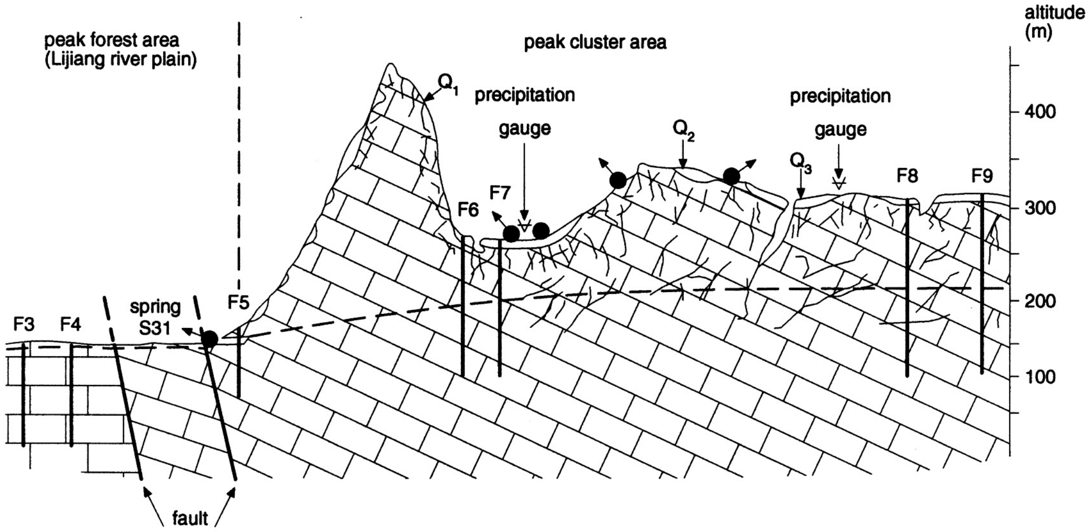
Figure 14.6 Correlation between precipitation and discharge of Nyanziguan karst spring, China, for the past 60 years (Zhang Fenngqi, 1992)



Because of the uneven distribution of rainfall in the mountainous region precipitation gauges should be installed in the studied catchment area and water table gauges in the epikarst zone, the soil water or vadose zone, and the general outlet of the system (Fig. 14.7). Given the quick response of water table and discharge to precipitation (Fig. 14.8), it is highly recommended to use automatic recorders for precipitation and water table gauges. For the observation of a deep and greatly fluctuating water table a pneumatic water table gauge should be used instead of a float gauge. If available, a satellite transfer multidata collecting platform powered by solar energy should be used.

Discharge of water from carbonate aquifers is commonly less diffused than that from other types of aquifers. Development of large solution openings through which there is preferential movement of water gives rise to large springs in many low places in carbonate terrain; the springs may be obscure or somewhat indistinct where they occur in river channels and in shallow seas. Discharge may be diffused where water passes from carbonate into sands or where relatively impermeable rocks underlie the aquifer above the low areas (Burdon and Papkis, 1963).

Figure 14.7 Components of water flow and monitoring facilities in Yaji karst hydrological system, Guilin, China. (Yuan Daoxian et al., 1990)



F; Borehole

S: Spring

— — Water table

Table 1

The hydrological features of output springs.

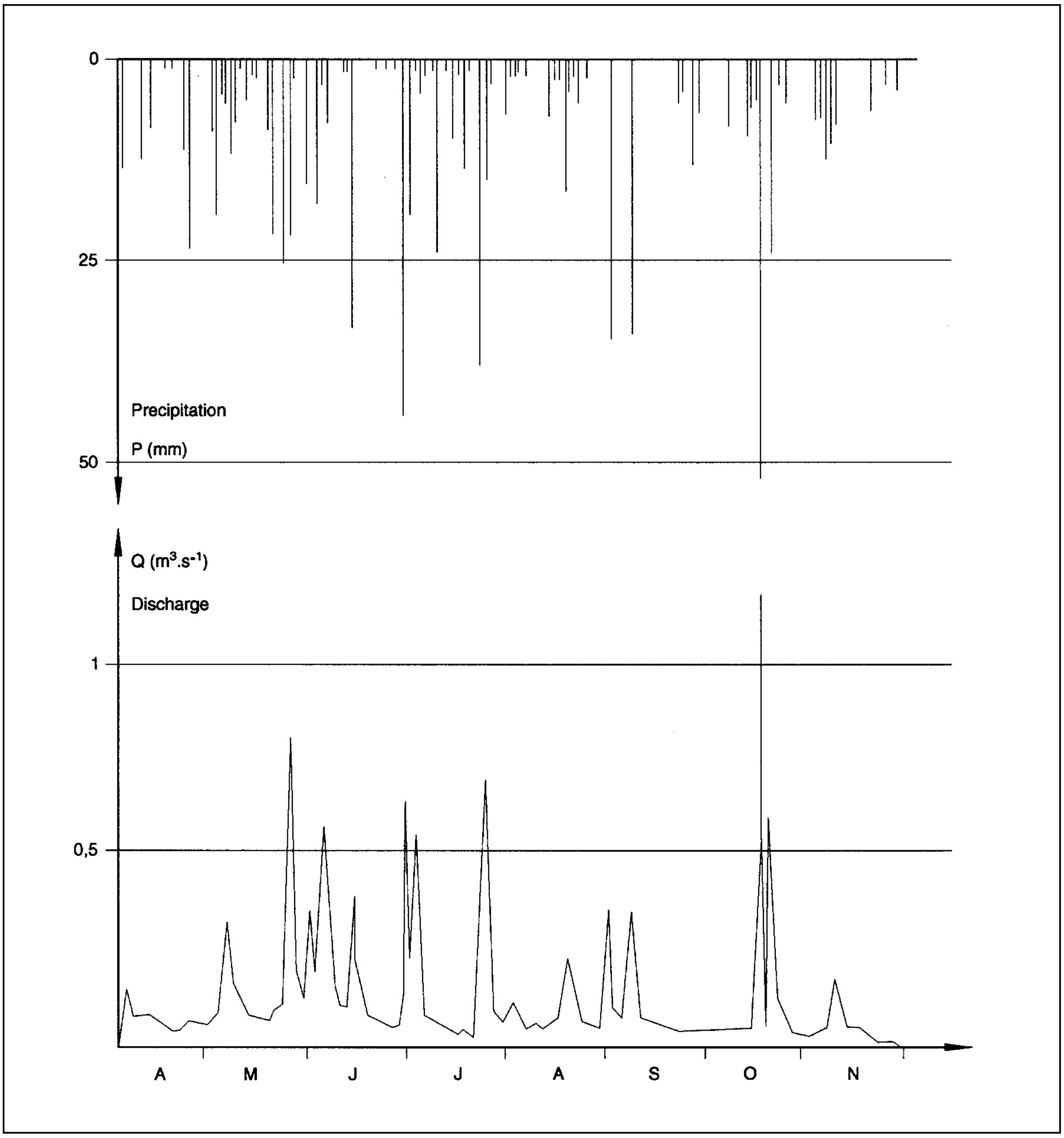
Spring	Altitude (m)	Maximum discharge ($l\ s^{-1}$)	Minimum discharge ($l\ s^{-1}$)
S31	153.4	7000	0.1
S29	153.3	470	0
S291	151.9	120	0
S32	160.3	30	0

Notes: Q_1 - slope runoff; Q_2 - epikarst; Q_3 - soil water.

Hydrological features of the springs

Spring	*Max. discharge (l/s)*	*Min. discharge (l/s)*
S31	7,000	0.1
S29	470	0
S29-1	120	0
S32	30	0

Figure 14.8 Hydrograph of spring S31 in Yaji karst system, Guilin, China, 1987 (Yuan Daoxian et al., 1990)



14.5 Hydrochemical character of carbonate rock aquifers

Hydrochemistry is an important part of water study for carbonate rocks. It can, on one hand, reveal the mechanism of karst formation, the intensity and direction of the process, i.e., whether it tends to dissolution or precipitation. The mixing corrosion on the coastal zone of Yucatan Peninsula, which was revealed through detailed hydrochemical studies, is a good example. On the other hand, the hydrochemical field in a carbonate rock aquifer, and the time variation of hydrochemical features at the output from the aquifer are important information from which the hydrological functions such as recharge, water storage and regulation, and the characteristics of the aquifer can be deduced. Accordingly, hydrochemical studies are very often used as a tool for hydrogeological exploration (Back et al., 1986).

14.5.1 Important hydrochemical features of carbonate rocks

The most essential hydrochemical characteristic of carbonate rocks is that it is a very active triphase, disequilibrium, open system because of the involvement of CO_2. The entering or outgassing of CO_2 will change not only the hydrochemistry of the water, but also the intensity and direction of karst processes. Consequently, the hydrochemistry is very sensitive to many environmental factors such as rainfall, temperature, photosynthesis, and other biogenic processes as sunshine, turbulence, and depth of water. The hydrochemical features are different even in different parts of the same water body. The traditional water/rock interaction approach based merely on laboratory chemical analysis may quite often lead to serious errors (see the special Issue of the Journal of Environmental Geology, JEG, 1995).

14.5.2. Environmental aspects and recommended references

The major environmental problems related to groundwater in carbonate rocks are drought, flood, deforestation, surface collapse and water pollution (Memon and Prohic, 1989). Some of these are caused by natural change of water regime, but most of them are induced by human activities, i.e., inappropriate management on groundwater in carbonate rocks (see the special issue of the Journal of Environmental Geology (JEG, 1993)). Different environmental problems are sometimes related to each other, when they occur in the same karst hydrological system. Accordingly, an overall understanding of the karst hydrological system is the basis for a proper management of groundwater and environment (Newton, 1976; Holzer, 1991).

Human endeavours involving karst terrain are confronted with a much wider range of conditions with poorly predictable surface and subsurface responses than occurs in most other terrains. Sensitive environmental problems prevail, many of which cannot be predicted and which have subtle and indirect cause-and-effect relations. Thus, certain legal aspects are very prominent when land is exchanged or dewatered and when the consequences of an action on one property affect neighbouring properties. It is not the purpose of this paper to describe the principles of karst hydrogeology that can be found in many geological texts on the subject (LaMoreaux and Newton, 1986; LaMoreaux et al., 1997).

A description of the structural setting is essential for understanding karst hydrology: (1) development (water and land use), (2) interactions, and (3) problems that may result in legal actions (LaMoreaux and Newton, 1986). Attention has been directed to these subjects in numerous publications, for example those by Meinzer (1923), Herak and Stringfield (1972), and Burger and Dubertret (1975). A summary of key principles described by LeGrand and LaMoreaux (1975), and specific methods for describing the geology, lithology, and structural setting in a karst area are discussed in great detail in the UNESCO's *Guide to the Hydrology of Carbonate Rocks* (LaMoreaux et al., 1984). Detailed descriptions of the important physical characteristics for siting a landfill in a karst terrain are given in *Landfills in Karst Terrains* (Hughes et al., 1994).

Various regional symposia and colloquia on karst have been organised world-wide over the past 20 years by IAH, IAHS, FAO and UNESCO within the International Hydrological Decade (IHD) and International Hydrological Programme (IHP). The IHD included a commission for the study of carbonate rocks in Mediterranean countries, and since 1970 there has been a permanent commission for karst hydrogeology within IAH. There have been over 150 field trips, meetings, symposia, and associated documents from this IAH activity, including *Karst Waters and Environmental Impacts* (UKAM, 1995). In southern France, one of the most detailed series of studies of spring flow, water management, and development for municipal use has been carried on by Professor Jacques V. Avias (1995) in his studies of the karstic spring Source du Lez. This spring supplies the city of Montpellier, France.

One of the most comprehensive series of papers on carbonate rocks appeared as a series of books on the symposium held under the leadership of Dr Barry Beck, first sponsored by the Sinkhole Institute of Florida and beginning in 1995, by the firm of P. E. LaMoreaux and Associates. These symposium papers, published by Balkema are: Beck, 1984, Beck and Wilson, 1987; Beck, 1989, 1993, 1995 and Beck and Stephenson, 1997. Since 1990, UNESCO and IUGS supported the 'International Geological Correlation programme' (IGCP) that organized three projects to deal with the karst hydrological and environmental problems of the world, incl. IGCP299 *Geology, climate, hydrology and karst formation* (1990–1994), IGCP379 *Karst processes and the carbon cycle* (1995–1999) and IGCP448 *World Correlation of karst ecosystems* (2000–2004). A series of books was published, including *Global Karst Correlation,* VSP, the Netherlands.

The reasons for the increased interest in the hydrogeology of carbonate rocks are the rapid increase in population, the application of new technologies, the more rapid development of natural resources, and the more extensive exploitation of hydrological systems. Karst occurs in many parts of the world where groundwater supplies represent the sole or most important natural resource to a population. Thus, understanding the behaviour of groundwater systems in a karst area directly affects the local or regional social and economic development.

14.6 References and additional reading

AVIAS, J. V., 1995. Active Management of Lez Karstic Spring (Hérault, France) 1957–1994. *Hydrogéologie,* No. 1, Éditions BRGM, Orléans, pp. 113–28.

BACK, W.; HANSHAW, B. B.; HERMAN, J. S.; VAN DRIEL, J. N., 1986. Differential Dissolution of a Pleistocene Reef in the Ground-Water Mixing Zone of Coastal Yucatan, Mexico. *Geology,* Vol. 14, pp. 137–140.

BAKALOWITCZ, M., 1979. Contribution de la géochimie des eaux à la connaissance de l'aquifère karstique et de la karstification. Thèse de doctorat, Lab. souterrain du C.N.R.S., Moulis, France.

BECK, B. F. (ed.), 1984. *Sinkholes: Their Geology, Engineering and Environmental Impact*. Proceedings of the First Multidisciplinary Conference on Sinkholes, Orlando, Florida, 15–17 October 1984. A. A. Balkema, Rotterdam, 429 pp.

—— (ed.), 1989. *Engineering and Environmental Impacts of Sinkholes and Karst.* Proceedings of the third multidisciplinary conference on sinkholes and the engineering and environmental impacts of karst, St. Petersburg Beach, Florida, 2–4 October 1989. A. A. Balkema, Rotterdam, 384 pp.

—— (ed.), 1993. *Applied Karst Geology*. Proceedings of the fourth multidisciplinary conference on sinkholes and the engineering and environmental impacts of karst, Panama City, Florida, 25–27 January 1993. A. A. Balkema, Rotterdam, 295 pp.

—— (ed.), 1995. *Karst Geohazards, Engineering and Environmental Problems in Karst Terrain.* Proceedings of the fifth multidisciplinary conference on sinkholes and the engineering and environmental impacts of karst, Gatlinburg, Tennessee, 2–5 April 1995. A. A. Balkema, Rotterdam, 581 pp.

BECK, B. F.; STEPHENSON, J. B., (eds.), 1997. *The Engineering Geology and Hydrogeology of Karst Terrain*. Proceedings of the sixth multidisciplinary conference on sinkholes and the engineering and environmental impacts of karst, Springfield, Missouri, 6–9, April 1997. A. A. Balkema, Rotterdam, 516 pp.

BECK, B. F.; WILSON, W. L., (eds.), 1987. *Karst Hydrogeology: Engineering and Environmental Applications*. Proceedings of the second multidisciplinary conference on sinkholes and the environmental impacts of karst, Orlando, Florida, 9–11 February 1987. A. A. Balkema, Rotterdam, 467 pp.

BURDON, D. J.; PAPKIS, N., 1963. *Handbook of Karst Hydrogeology, with Special Reference to the Carbonate Aquifers of the Mediterranean Region.* Inst. Geol. and Subsurface Research, Athens, United Nations Special Fund, mimeo, 276 pp.

BURGER, A.; DUBERTRET, L., (eds.), 1975. *Hydrology of Karstic Terrain.* Int. Union Geol. Sci., Ser. B, No. 3, 190 pp.

BUTTRICK, D. B., 1992. *Characterisation and Appropriate Development of Sites on Dolomite.* In partial fulfilment of the requirements for the degree of Doctor of Philosophy in the Faculty of Science, University of Pretoria, November 1992, 182 pp.

CHEN XIAOPING ET AL., 1998. *South China Karst I.* Zalozil (ZAL, ZBA, ZRC), Republic of Slovenia, 247 pp.

DAVIS, W., 1901. An Excursion in Bosnia, Herzegovina and Dalmatia. *Bull. Geograph. Soc.,* Vol. 3, No. 2, pp. 47–50.

DREW, D.; HOETZL, H., (eds.) 1999. *Karst Hydrogeology and Human Activities: Impacts, Consequences and Implication* A. A. Balkema, Rotterdam, 322 pp.

ENSLIN, J. F., 1967. Dolomitic Water Supplies in the Transvaal Republic of South Africa. *Mem. IAH Conference, Istanbul, Sept. 1967,* p. 78.

FORD, D.; WILLIAMS, P., 1989. *Karst Geomorphology and Hydrology*. Unwin Hyman Ltd., London, 601 pp.

GRICE, R. H., 1968. Hydrogeology of the Jointed Dolomites, Grand Rapids Hydroelectric Power Station, Manitoba, Canada. In: *Engineering Geology Case Histories: No. 6*, Geol. Soc. of Am., pp. 33–48.

GULTEKIN GUNAY; JOHNSON, A. I., (eds.), 1997. *Karst Waters and Environmental Impacts*. Proc. 5th international symposium and field seminar on karst water and environmental impacts, Antalya, Turkey, 10–20 September 1995. A. A. Balkema, Rotterdam/Brookfield, 525 pp.

HERAK, M.; STRINGFIELD, V. T., (eds.), 1972. *Karst: Important Karst Regions of the Northern Hemisphere*. Elsevier, Amsterdam, 526 pp.

HOLZER, T. L. (ed.), 1991. *Mitigating Losses from Land Subsidence in the United States, Panel on Land Subsidence*. U.S. Geological Survey, Menlo Park, California, National Academy Press, Washington, D.C., 58 pp.

HUGHES, T. H.; MEMON, B. A.; LAMOREAUX, P. E., 1994. Landfills in Karst Terrains. *Bull. Assoc. Eng. Geol.,* Vol. XXXI, pp. 203–8.

JAMES, N. P.; CHOQUETTE, P. W., (eds.), 1987. *Paleokarst*. Springer-Verlag, New York, NY, 416 pp.

JEG, 1993. Environmental Change in Karst Areas. *Journal of Environmental Geology*, special issue, Vol. 21, No. 3, pp. 107–82.

JEG, 1995. Geochemical Environments in Karst. *Journal of Environmental Geology,* Vol. 25, No. 1, 69 pp.

KIERSCH, G. A.; HUGHES, P. W., 1952. Structural Localisation of Groundwater in Limestone – Big Bend District, Texas-Mexico. *Econ. Geology,* Vol. 47, pp. 794–806.

KLIMCHOUK, A. B.; FORD, D. C.; PALMER, A. N.; DRYBRODT, W., (eds.) 1999. *Speleogenesis; Evolution of Karst Aquifers*. National Speleological Society, Inc., Huntsville, Alabama, 527 pp.

LAMOREAUX, P. E., 1975. *Environmental Geology and Hydrology, Huntsville and Madison County, Alabama.* Geological Survey of Alabama, Atlas Series 8, University, Alabama, 119 pp.

LAMOREAUX, P. E.; HUGHES, T. H.; MEMON, B. A.; LINEBACK, N., 1989. Hydrogeologic Assessment – Figeh Spring, Damascus. *Environmental Geology and Water Sciences*, Vol. 13, No. 2, pp. 73–127.

LAMOREAUX, P. E.; NEWTON, J. G., 1986. Catastrophic Subsidence: An Environmental Hazard, Shelby County, Alabama. *Environmental Geology and Water Sciences,* Springer-Verlag, Heidelberg, Vol. 8, pp. 25–40.

LAMOREAUX, P. E.; POWELL, W. J.; LEGRAND, H. E., 1997. Environmental and Legal Aspects of Karst Areas. *Environmental Geology,* Vol. 29, pp. 23–36.

LaMoreaux, P. E.; Wilson, B. M.; Memon,B. A., (eds.), 1984. *Guide to the Hydrology of Carbonate Rocks, Studies and Reports in Hydrology*. UNESCO, Paris, 345 pp.

LeGrand, H. E.; LaMoreaux, P. E., 1975. Hydrogeology and Hydrology of Karst. In: A. Burger, L. Dubertret, (eds.). *Hydrogeology of Karst Terrains*. Intern. Assoc. Hydrogeol., Paris, pp. 9–19.

LeGrand, H. E.; Stringfield, V. T., 1966. Development of Permeability and Storage in the Tertiary Limestones of the Southeastern States. *Intern. Assoc. Sci. Hydrol.,* Vol. 9, No. 4, pp. 61–73.

——, 1971. Development and Distribution of Permeability in Carbonate Aquifers. *Water Resources Research,* Vol. 7, No. 5, pp. 1284–94.

LeGrand, H. E.; Stringfield, V. T.; LaMoreaux, P. E., 1976. Hydrologic Features of United States Karst Regions. In: J. W. Tolson, F. L. Doyle, (eds.), K*arst Hydrology.* AIH Memoirs, Vol. XII, pp. 31–46.

Leve, G. W., 1983. Relation of Concealed Faults to Water Quality and the Formation of Solution Features in the Floridan Aquifer, Northeast Florida, United States of America. In: W. Back, P. E. LaMoreaux, (guest eds.), V. T. Stringfield Symposium: Processes in Karst Hydrology. *Journal of Hydrology,* Vol. 61, pp. 251–64.

Meinzer, O. E., 1923. *Outline of Groundwater Hydrology with Definitions*. U.S.G.S. Water Supply Paper 494, 71 pp.

Memon, B. A.; Prohic, E., 1989. Movement of Contaminants in Karstified Carbonate Rocks. *Environmental Geology and Water Sciences,* Vol. 13, pp. 3–13.

Milanovich, P., 1981. *Karst Hydrogeology*. Water Resources Publications, Fort Collins, Co., 434 pp.

Moore, G. C., 1966. Introduction to Limestone Hydrology. *Nat. Speleol. Soc. Bull.,* Vol. 28, pp. 109–17.

National Research Council, 1991. *Mitigating Losses from Land Subsidence in the United States*. NRC, Panel on Land Subsidence. T. L. Holzer and committee members (eds.). National Academy Press, Washington, D.C., 58 pp.

Newton, J. G., 1976. *Early Detection and Correction of Sinkhole Problems in Alabama, with a Preliminary Evaluation of Remote Sensing Applications*. HPR Report No. 76, Research Project 930–070, prepared by United States Geological Survey in co-operation with Alabama Highway Department and United States Department of Transportation, January 1976, 83 pp.

Newton, J. G., 1986. *Development of Sinkholes Resulting from Man's Activities in the Eastern United States*. U.S. Geological Survey, Denver, CO, 54 pp.

Palmer, A. N., 1977. Influence of Geologic Structure on Groundwater Flow and Cave Development in Mammoth Cave National Park, United States of America. In: J. S. Tolson, F. R. Doyle, (eds.), *Karst Hydrogeology*. AIH Memoirs, Vol. XII, pp. 405–30.

Palmer, A. N.; Palmer, M. V.; Sasowsky, I. D., (eds.), 1999. *Karst Modelling.* Proceedings Symposium held February 24–27, 1999, Charlottesville, Virginia. Karst Water Institute special publication No. 5, 265 pp.

Palock, H.; Back, W.; Herman, J. S., (eds.), 1992. Hydrogeology of Selected Karst Regions. *IAH Contributions to Hydrogeology,* Vol. 13. Verlag Heinz Heise, Germany, 494 pp.

Peck, D. L.; Troester, J. W.; Moore, J. E., 1988. Karst Hydrogeology in the United States of America. In: *Proc. of the 21st Congress of the International Association of Hydrogeologists, Karst Hydrogeology and Karst Environment Protection, Guilin, China.* U.S. Geological Survey, Open-File Report 88-476, 19 pp.

Penck, A., 1900. Geomorphologische Studien aus der Hercegovina. Z. Deut. Osterreich. *Alpenver,* Vol. 31, pp. 25–41.

Sauro, U.; Bondesan, A.; Meneghel, M., (eds.), 1991. *Proceedings of the International Conference on*

Environmental Changes in Karst Areas (Italy, 15–27 September 1991). Quaderni del Dipartimento di Geografia 13, Italy, 414 pp.

SMITH, D. I.; ATKINSON, T. C., 1977. Underground Flow in Cavernous Limestones, with Special Reference to the Malham Area. *Field studies,* Vol. 4, pp. 597–616.

STEARNS, R. G., 1977. Porosity and Hydrology of Jointed Middle Ordovician Limestones in the J. Percy Friest Dam Area of Central Tennessee. In: J. S. Tolson and F. L. Doyle, (eds.), *Karst Hydrology.* IAH. Memoirs, Vol. XII, pp. 431–2.

STRINGFIELD, V. T., 1966. *Artesian Water in Tertiary Limestone in the Southeastern States.* United States Geological Survey Professional Paper 517, Washington, D.C., 226 pp.

STRINGFIELD, V. T.; LEGRAND, H. E., 1966. Hydrology of Limestone Terraines in the Coastal Plain of the Southeastern United States. Geological Society of America, Special paper, Vol. 93, 45 pp.

STRINGFIELD, V. T.; RAPP, J. R.; ANDERS, R. B., 1979. Effects of Karst and Geologic Structure on the Circulation of Water and Permeability in Carbonate Aquifers. *Journ. of Hydrol.,* Vol. 43, pp. 313–32.

SWEETING, M. M., 1995. *Karst in China; its Geomorphology and Environment*. Springer-Verlag, Berlin, 265 pp.

SWINNERTON, A. C., 1942. Hydrology of Limestone Terranes. In: O. E. Meinzer (ed.), *Physics of the Earth, pt. 9: Hydrology.* Doves Publications, Inc., New York, pp. 656–77.

TOTH, J., 1962. A Theory of Groundwater Motion in Small Drainage Basins in Central Alberta, Canada. *Journ. Geophys. Research,* Vol. 67, pp. 4375–87.

TOTH, J., 1963. A Theoretical Analysis of Groundwater Flow in Small Drainage Basins. *Journ. Geophys. Research,* Vol. 68, pp. 4795–812.

UKAM, 1995. *Karst Waters and Environmental Impacts.* Proceedings of the Antalya symposium, September 10–20, 1995. UKAM, Haceteppe University, Ankara.

VERNON, R. O., 1951. Geology of Citrus and Levy Counties, Florida. *Florida Geological Survey, Bulletin 33,* pp. 256.

WHITE, W. B.; WHITE, E. L., (eds.), 1989. *Karst Hydrology, Concepts from the Mammoth Cave area,* Van Nostrand Reinhold, New York, 346 pp.

WILLIAMS, P. W., 1983. The role of the Subcutaneous Zone in Karst Hydrology. In: W. Back, P. E. LaMoreaux, (eds.), V. T. Stringfield Symposium – Processes in Karst Hydrology. *Journal of Hydrology,* Vol. 16, pp. 45–67.

YUAN DAOXIAN, (chief ed.), 1988. *Karst Hydrogeology and Karst Environment Protection*. Proceedings of the IAH 21st congress, 10–15 October 1988, Guilin. Geol. Publish. House, Beijing, 2 parts, 1261 pp.

YUAN DAOXIAN (ed.), 1991. *Karst of China*. Geological Publishing House, Beijing, 224 pp.

YUAN DAOXIAN; DROGUE, C. ET AL., 1990. Hydrology of the Karst Aquifer at the Experimental Site of Guilin in South China. *Journ. of Hydrol.,* Vol. 115, pp. 285–96.

YUAN DAOXIAN; LIU ZAIHUA (eds.), 1998. *Global Karst Correlation*. Final report of IGCP299: Geology, Climate, Hydrology and Karst Formation (1990–1994). Science Press, Beijing-New York; VSP, Utrecht, 308 pp.

ZHANG FENNGQI; LI BOTAO, 1992. *Artificial Adjustment and Recharge of Karst Springs, and Prospect of Water Resources in Northern China*. Proceedings of the Symposium on karst and karst waters in north China, October 7–11, 1991. Jinan, Shandong; Guangxi Normal University Press, pp. 88–96.

15 Hydrogeology of hard rocks

15.1 Introduction

For the purpose of this chapter, hard rock aquifers are aquifers in non-carbonate, fractured rock like crystalline basement complex and metamorphic rocks. Extensive, ancient volcanic terrains are also included as a special case in this discussion.[1] For example, in western India, hundreds of nearly horizontal, basaltic lava flows form a thick pile that is known as the Deccan trap, and covers around 500,000 km^2. This sequence was not tectonically disturbed after consolidation and due to the non-frothy nature of the lava a hand specimen does not show any primary porosity. Due to the great thickness of the lava flows, it is not possible to drill through them to obtain groundwater from the underlying strata. Hydrogeologically, the Deccan trap, therefore, acts as a fractured basement complex.

The most significant features of hard rock aquifers are as follows:

(i) A topographical basin or sub-basin generally coincides with the groundwater basin, the groundwater resources tend to be concentrated towards the central valley portion, closer to the main stream and its tributaries.

(ii) The depth of groundwater occurrence, in useful quantities, is usually limited to a hundred metres or so.

(iii) The aquifer parameters like Storativity (S) and Transmissivity (T) often show erratic variations within small distances. The annual fluctuation in the value of transmissivity is considerable, due to the change in saturated thickness of the aquifer from the wet season to the dry season. When different formulae are applied to pump test data from one bore well, a wide range of S and T values may be obtained. The applicability of mathematical modelling is limited to only a few simpler situations.

(iv) The saturated portion of the mantle of weathered rock or alluvium or laterite overlying the hard fractured rock often makes a significant contribution to the yield obtained from a dug well or bore well.

(v) Only a modest quantity of groundwater, in the range of 1.0 to 100 m^3 per day, is available at one spot. Drawdown in a pumping dug well or bore well is often almost equal to the saturated thickness of the aquifer.

(vi) For sustainable groundwater development, it is advisable to take a sub-basin as a unit and to plan for conjunctive use of surface water and groundwater. It is also advisable to adopt soil and water conservation techniques for augmenting recharge

1. For a detailed discussion of the hydrogeology of volcanic rocks see Chapter 16.

to groundwater body, so as to minimise the adverse effects of increased pumpage due to groundwater development. In many sub-basins underlain by fractured rock aquifers, groundwater has a residence time of a few years only and the system is very sensitive to droughts.

In hard rock aquifer areas, groundwater development has always played a secondary role compared to that in the areas having high-yielding unconsolidated or semi-consolidated sediments and carbonate rocks. This is because of the relatively poor groundwater resources in hard rocks, the low specific capacity of wells, erratic variations and discontinuities in the aquifer properties, and difficulties in exploration and assessment of the resource.

It should, however, be remembered that for the millions of farmers in developing countries, with small farms in fractured basement or basaltic terrain, whatever small supply is available from these poor aquifers is their only hope for upgrading their standard of living by growing irrigated crops or by buffeting their rain-fed crops against the vagaries of rainfall. It is also their only source for drinking water for the family and cattle. In many developing countries, hard rock hydrogeologists have an important role to play. In the developed countries, interest in hard rock hydrogeology, other than in relation to drinking water supplies to small communities, has increased recently because of the prospects of using these low permeability rocks for the storage of hazardous nuclear and chemical wastes. The study of groundwater flow through faults, fissures and fractures is also of interest to scientists studying the migration of contaminants and to engineers engaged in tunnelling through hard rocks for water supply and highway construction projects.

Hard rock hydrogeologists are therefore divided into two main groups: those interested in obtaining groundwater for domestic, irrigation or industrial use by exploring fractured and permeable zones in a relatively less permeable matrix of hard rock and those interested in locating impermeable or very low permeability zones for storage of hazardous waste. Ironically, for the first group even the most permeable zones are often not good enough to yield an adequate water supply, while for the second group even the least permeable zones are often not good enough for safe storage of hazardous waste over a prolonged period.

15.2 Occurrence of groundwater

Groundwater under phreatic condition occurs in the mantle of weathered rock, alluvium and laterite overlying the hard rock, while within the fissures, fractures, cracks, joints and lava flow junctions within the hard rock, groundwater is mostly in a semi-confined state. Compared to the volume of water stored under semi-confined conditions within the body of the hard rock, the storage in the overlying phreatic aquifer is often much greater. In such cases, the network of fissures and fractures serves as a permeable conduit feeding this water to the well.

The recharge to groundwater takes place during the rainy season through direct infiltration into the soft mantle overlying the hard rock and also into the exposed portions of the network of fissures and fractures. In India and other Asian countries, the ratio of recharge to rainfall in hard rock terrain is usually between 3 and 15%, depending upon the amount and nature of precipitation, the nature and thickness of topsoil and weathered zone, and the topographical features of the sub-basin. Groundwater flow rarely occurs across the topographical water divides; for planning the development of groundwater resources each basin or sub-basin can be treated as a separate hydrogeological unit. After the rainy season, the fully recharged hard rock aquifer gradually loses its storage, mainly due to pumpage and effluent drainage by streams and rivers. The dry season flow of the streams is thus supported by groundwater outflow. The flow of groundwater is from the peripheral portions of a sub-basin to the central valley portion, thereby causing dewatering of portions closer to the topographical groundwater divides. In many cases, dug wells and bore wells yielding a perennial supply of groundwater can only be located in the central valley portion,

The average residence time in a sub-basin of about 100 km^2 is up to five years. The annual recharge is thus a sizeable part of the total resource of an aquifer and the whole system is very sensitive to the availability of recharge during the rainy season. Two drought years in succession can pose a serious problem. Under such conditions the low permeability of hard rock aquifers is a redeeming feature, because it makes small quantities of water available, at least for drinking purposes, in the dug wells or bore wells in the central valley portion of a sub-basin. If the hard rocks have a very high permeability, the groundwater body will move quickly towards the main river basin, thereby leaving sub-basins high and dry. The low permeability in the range of 0.05 to 1.0 m/d thus helps in retarding the outflow and regulating the availability of water in individual wells. More farmers are able to dig or drill their wells and irrigate small plots of land without causing harmful mutual interference.

15.3 Groundwater development

Development of a natural resource like groundwater towards its optimum utilisation will benefit mankind. In the highly populated but economically backward areas in hard rock terrain, many governments in the developing countries have taken up schemes to encourage small farmers to dig/drill wells for irrigation. This is especially true for the semi-arid regions where surface water resources are meagre. For example, in peninsular India, hard rocks such as granite, gneiss, schist, quartzite and basalts (Deccan traps) occupy about 1.15 million km^2 area, out of which about 40% is in the semi-arid zone, receiving less than 750 mm rainfall per year. Over 3.5 million dug wells and bore wells are being used in the semi-arid region for irrigating small farm plots and providing domestic water supply.

Development of groundwater resources for irrigation and domestic use is thus a key factor in the economy of vast stretches of semi-arid, hard rock areas. The basic need of millions of farmers in such areas is to obtain an assured irrigation supply for at least one crop per year and to have a protected, perennial drinking water supply within a reasonable walking distance. The hard-rock hydrogeologists in many developing countries have to meet this challenge to impart social and economic stability to the rural population, which would otherwise migrate to the neighbouring cities. Exploration and assessment of groundwater, finding suitable sites for locating dug wells and bore wells and planning for long term sustainability of the wells, are the main tasks under these circumstances.

For promoting large-scale irrigation development from groundwater resources, institutional finance has to be made available to the farmers at concessional rates of interest, for digging or drilling wells. Before selecting an area covering several sub-basins, for financing irrigation wells, the primary requirement is the assessment of the available groundwater. In the absence of reliable data on rainfall, evapotranspiration, flood discharge of a stream and pumpage from existing wells in the sub-basin drained by the stream, a fair estimate of groundwater resource available for new development can be made by assessing the dry season flow and underflow of the stream. This base flow is supported by the effluent drainage from groundwater in the sub-basin and a part of it can be tapped by the proposed new wells. Once the technical feasibility is ascertained, the financing institutions or banks need to estimate the economic viability of an average individual well, based on the average yield of the well in different cropping seasons, supporting a suitable cropping pattern. The total incremental income from such a cropping pattern should enable the farmer to repay the loan with interest, over an average period of 7 to 9 years. Some banks also make provision for insurance of failed wells.

Groundwater development in a sub-basin results in increased pumpage due to the new wells. This lowers the water table and hence reduces the effluent drainage from the sub-basin. Such development in several sub-basins draining into a main river basin reduces the surface flow and the underflow of the river, thereby affecting some of the surface water schemes dependent on the river flow. In order to minimise such an interference, it is advisable to apply water conservation and recharge augmentation techniques in the sub-basins, simultaneously with the groundwater utilisation programmes.

15.4 Types of wells

It is common to use dug wells and dug-cum-bored wells to obtain shallow, phreatic groundwater in low permeability hard rock aquifers. A dug well offers more area than a tubewell for inflow of groundwater into it and also provides sizeable storage for accumulating water. The volume of rock excavated while constructing a dug well is often much more than the REV (Representative Elemental Volume) of the fractured rock aquifer and there are reasonable chances of tapping water-bearing vertical and horizontal fractures and fissures during the excavation, especially if the site for the well is carefully selected. Otherwise, these fractures may be dry or may bear water for only a few months of the year.

A typical dug well has a diameter of 3 to 8 metres and depth of 5 to 15 metres. It usually penetrates through the weathered mantle and goes a few metres into the underlying fractured rock. A retaining wall of stone or brick masonry is required to support the upper portion of a dug well excavated in soil, sub-soil and highly weathered rock overlying hard rock. The masonry wall is provided with weep holes, especially near its base to allow inflow of phreatic water into the well. In areas where hard rock is covered by laterite or a weathered zone rich in lime nodules, a retaining wall may not be necessary to support the excavation. Some of the deeper dug wells are dug through the weathered mantle and the fissured rock, into the massive hard rock below, so as to provide adequate storage for water flowing into the well during a period of 8 to 10 hours. This allows the farmer to conveniently pump the stored water, once in the morning and once in the evening, by using a low-cost centrifugal pump set. It is customary to provide two or three platforms at different levels inside the well, so that the pump set can be installed according to the water levels in different seasons. The foot valve of the pump is kept in a small pit excavated in the well bottom, so as to facilitate emptying of the well. On shallow wells, animal-drawn devices or human-powered water wheels can also be employed to lift water for irrigation of small plots.

Horizontal bores can be drilled radially outward from a dug well, at various levels below the water table, so as to increase well yield. Horizontal bores of 25 mm to 75 mm diameter and of lengths up to 10 to 15 metres are drilled in laterite, alluvial cover and weathered rock overlying the fractured hard rock, manually or with a drilling machine. In very soft strata, such bores have to be supported by inserting slotted casing pipes. Such bores increase the effective diameter of the dug well, without a proportional increase in the cost. Horizontal boring in fractured hard rock is done to tap more water from the neighbouring vertical fissures and fractures. Such bores usually extend radially outward from the wet patches or mouths of springs issuing into the dug well from the fractured strata.

Vertical bores of 100 mm to 150 mm diameter and depths up to 50 to 60 metres are drilled in the bottom of the dug wells to tap semi-confined water in the deeper horizontal or sub-horizontal fractures, joints or flow junctions. Water from such bores rises over the bottom of the dug well and accumulates there if the bottom of the dug well is at least 5 metres below the water table. A centrifugal pump installed on the dug well can then pump this water out. If water from the vertical bore does not rise into the well, a separate pump is required. A submersible pump can be installed if the yield from the bore is above 4 to 5 m^3 per hour, and the water can be directly taken for irrigation, rather than stored in the dug well. Submersible pumps being expensive, 'low water level' guards are installed to stop the pump operating when the water level approaches the top of the pump. This prevents dry running of the pump. The pump can be restarted by using a timer switch on the control panel or by using a high water level sensor in the bore. If the yield is small, a Jet or Ejecto pump is used to pump out water from the bore. The water is stored in the dug well until a sizable quantity has accumulated in 8 to 10 hours and then pumped out in 2 to 3 hours by the centrifugal pump installed on the dug well. In this case, a casing pipe of about 3 to 4 metres length is taken and half of it is inserted into the bore, with the remaining half sticking out over the well bottom. Cement concrete is put around this pipe so that the water accumulating in the dug well does not flow back into the bore.

Figures 15.1(a) and (b) illustrate dug wells with horizontal and vertical bores respectively, in horizontally and vertically fractured strata. In the first figure, the horizontal bores in the dug well are successful because they intersect a vertical fissure, however the vertical bore has failed. In the second figure the vertical bore in the dug well is successful but horizontal bores have failed. Figure 15.1(c) shows two wells on either sides of a stream in metamorphic strata of alternating bands of low permeability schist and more permeable fractured quartzite. Well A, located on the down dip side from the stream, gives some water from the phreatic aquifer in weathered rock and also has a successful vertical bore in its bottom. Well B is a relative failure.

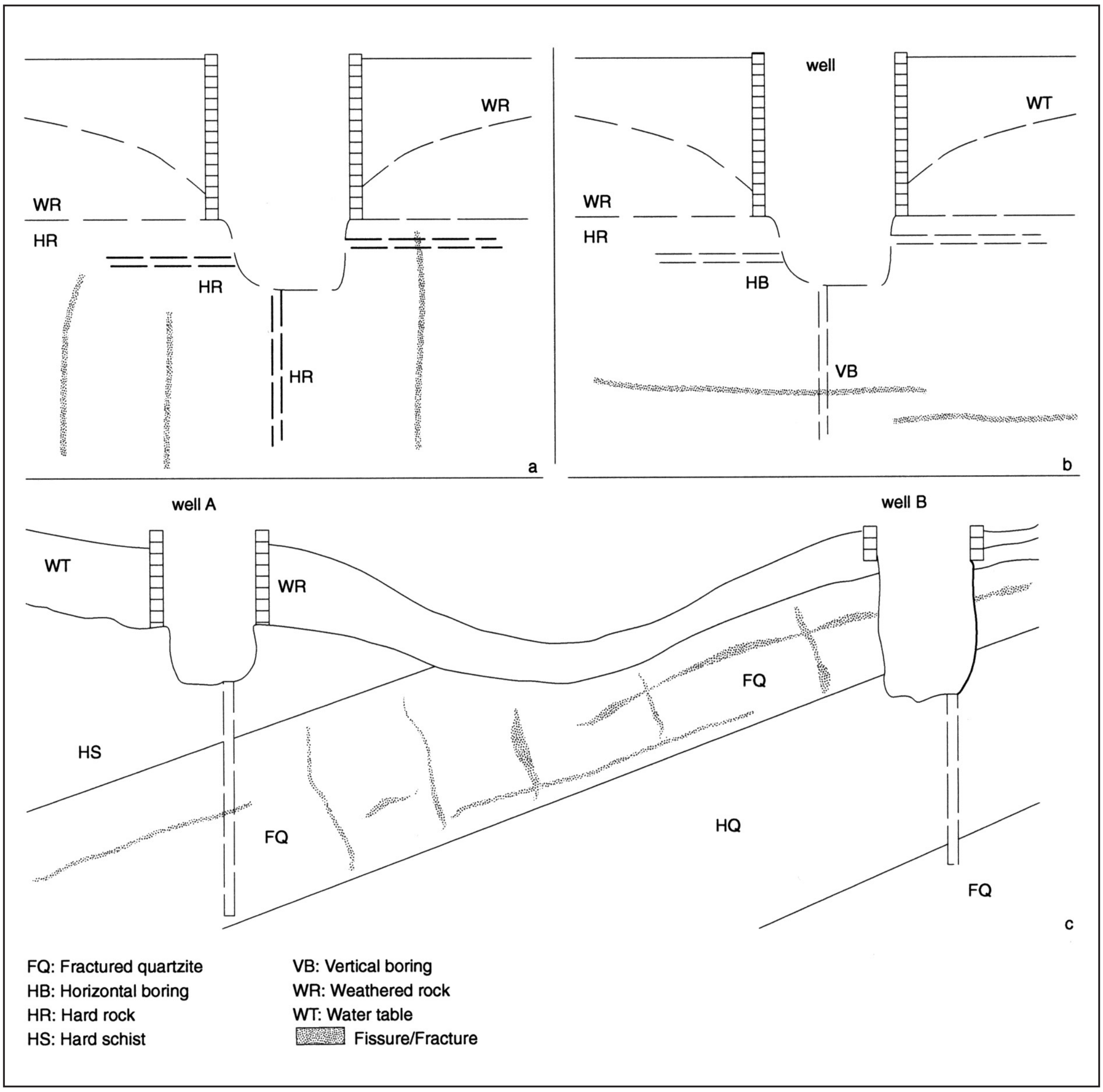
It will be clear from the figures that in areas where vertical fractures, fissures and joints occur in the vicinity of a dug well, horizontal bores will be more successful, while in areas where horizontal or sub-horizontal joints, bedding planes, cleavages, flow junctions or fractures occur below the bottom of dug wells, vertical bores will be more successful.

The cost of a dug well of about 6 m diameter and of 15 m depth in India and other Asian countries like Nepal, Pakistan and Sri Lanka, would be around US$3,000. In African countries like Kenya or Zimbabwe, the cost would be almost double due to scarcity of trained labourers. In the soft strata near the ground surface, the excavation is cheaper but a masonry wall has to be built to support the strata. In the underlying hard fractured rock, the excavation is expensive because dynamite blasting is often necessary. However, this portion of the dug well does not need support. When blasting in the fractured rock for well excavation below the water table level, it is advisable to blast only one or two holes at a time, instead of simultaneously blasting all the 40 or 50 holes drilled in the bottom of the excavation. If all the holes are simultaneously blasted, the fracture network often receives a heavy shock and the yield of water into the dug well may be reduced. The cost of drilling horizontal bores is around US$2 to 3 per metre; in India vertical bores cost around US$4 to 5 per metre.

Other types of dug wells include large diameter (10 to 20 metres) shallow dug wells of 4 to 5 metres depth, excavated in coastal areas for skimming fresh water floating over saline water and rectangular 'trench' wells of 1 to 2 metres width and 10 to 50 metres length excavated at right angles to the direction of groundwater flow. Such a 'trench' well dug across the bed of a stream or river, at a carefully selected location, taps a large quantity of the underflow, even after the stream dries up. It is customary to back-fill the excavation, creating an inverted filter, with boulders in the bottom, gravel in the middle and coarse sand in the upper portion. The porous and permeable trench thus created across the river bed is connected to a small, circular dug well or 'jack well', on the river or stream bank.

The construction of a dug well or a dug-cum-bore well is expensive and time consuming. Down-the-hole-hammer rigs have provided an alternative that enables a surface bore well of 150 mm diameter and up to 100 metres depth to be drilled in one day. The cost in India is around US$600. The upper portion of the bore well in soft mantle and weathered rock has to be supported with steel or PVC casing pipe. Unless this casing pipe has slots near its lower end to allow inflow of water from the phreatic aquifer, the bore well yields all of its supply from the network of fractures and fissures in the hard rock. However, this network is hydraulically

Figure 15.1 The well in (a) is mainly filled by flow from the horizontal bores, in (b) the well is mainly filled by flow from the vertical bore and (c) shows two wells of which well A is successful and B is a relative failure



connected to the phreatic aquifer, and pumpage from the bore well results in elastic compression of the network plus dewatering of the phreatic aquifer.

Experience in Afro-Asian countries indicates that most of the bore wells in fractured rock terrain meet their supply within the first 50 metres. In some cases the supply increases by drilling deeper, up to 100 metres. Very few bores, less than one percent in India, tap water-bearing fissures or fractures below 100 metres depth. However, some of these deep bores do come into the category of 'high-yielding bores'.

In peninsular India, basement complex and basalt (Deccan trap) cover an area of around 1.15 million km^2. Most of the bore wells drilled here fall in the category of low-yielding bores, yielding up to 2 m^3 per hour. In basaltic terrain there is some correlation between shallow water table and good yield of the bore wells but such a correlation is not observed in granitic terrain.

Revitalisation of low-yielding bores is a topic of great interest to hydrogeologists and farmers. If the low yield is due to some fissures becoming clogged with rock dust during the drilling operation, the yield gradually improves after using the bore for a few years. Immediately after the rainy season, when the water table is high, the bore well is pumped to create a large drawdown. The clogged fissures get flushed when water rushes into the bore under a greater head difference than normally encountered. In some cases, working a rubber rimmed plunger up and down like a piston, below the water table in the bore well, also gives beneficial results. Blasting up to 2 kg of dynamite charge inside a low-yielding bore well is the last resort. The charge may be shrouded in coarse sand. The bore well is filled with water up to the brim and the top of casing pipe is plugged with a cotton ball. Blasting helps in some cases where the fissures, that have opened up or are newly formed by blasting connect the bore well to more permeable network of fissures in the vicinity. In some other cases the bore well collapses and is rendered useless. In the remaining cases, blasting has no effect on the yield.

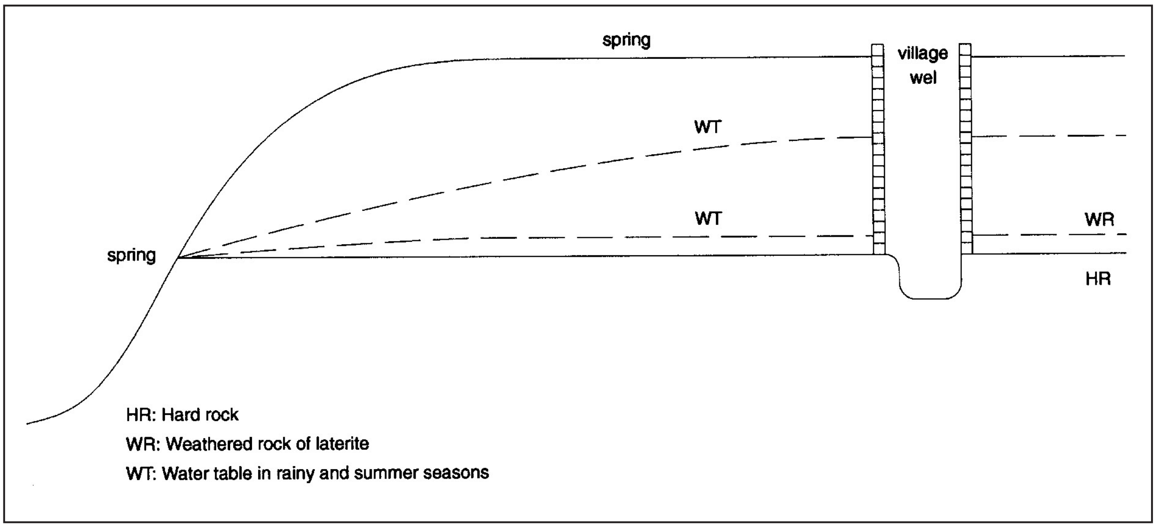
Hydrofracturing of low-yielding bore wells is a relatively new technique and requires expensive equipment for injecting large volumes of water under high pressure into the bore, so as to 'jack up' the existing low permeability fractures met within the bore well. Some of these fractures become extended and connect with the network of more permeable fractures if there is such a network in the vicinity. Hydrofracturing is commonly used in United States of America, Australia and South Africa. In South Africa, flow rates of injection water in a successful operation are more than 15 l/s at a pressure of 80 bars. (Less et al., 1993). It has been observed that hydrofracturing is more successful at sites selected on the basis of lineament mapping and geophysical exploration. At such sites the initial low yield is due to the unfortunate fact that the bore well has missed the main fracture network by a few metres, hence during hydrofracturing its connection to the main network is established.

15.5 Drinking water supply

Villages and communities situated away from any surface water source depend on locally available groundwater to meet their drinking water needs. Economically it is not possible for them to bring water in pipeline from distant sources. In remote hilly areas in semi-arid hard rock terrain, any small supplies of groundwater available from dug wells or bore wells are of vital importance for survival, especially in summer months, when yields as low as 1 m^3/d are considered as very useful. Figure 15.2 shows the situation of a village located not far from a steep incised plateau. The water table in the fractured and weathered rock overlying the hard rock is high after the rainy season and the villagers get their drinking water supply from the spring at the cliff or from the dug well. Towards summer, the water table gradually declines and the spring dries or its yield is reduced to a wet patch, just supporting some local vegetation at the mouth of the spring. The thickness of the saturated aquifer is negligible at the spring, but away from the cliff face the saturated thickness gradually increases up to a metre or so and the dug well is able to yield a small quantity of drinking water during summer. Due to the small hydraulic gradient and the low permeability, the groundwater body does not get completely drained and assumes a 'quasi-static stage' towards the summer season. In some villages, only the spring water is used as long as the spring runs, and the dug wells are brought into use only after the spring dries at the beginning of the summer. The cone of depression thus starts developing around the dug well in summer, not earlier.

In some villages in hard rock terrain, bore wells do not yield perennial water supply, due to the absence of deeper fissures and fractures. Dug wells of 6 to 8 metres depth also go dry at the start of the summer, due to the small thickness of the weathered zone, rugged terrain and limited rainfall. It is expensive to provide drinking water to such villages in the summer season

Figure 15.2 Effect of natural drainage on a village well



by tankers, especially in hilly terrain. In Maharashtra state in western India, experiments in creating an artificial aquifer by blasting around the existing dug well, have been successfully carried out. In these experiments about 80 to 120 bore holes of 100 mm diameter and 6 to 8 metres depth were drilled around the dug well up to about 60 metres distance and blasted simultaneously, in order to create a jacket of fractured rock aquifer around the dug well. Such a dug well could then provide a small quantity of drinking water, even in summer.

However, water from dug wells easily becomes polluted due to percolation from cattle sheds, septic tanks, manure pits, garbage dumps and fertilisers used on farms. Dug wells located a few hundred metres away and on the up-gradient side of a village have a better quality of water. Biological contamination and high nitrate values appear in some cases. Villages accessible to drilling rigs are therefore increasingly being supplied with drinking water from bore wells. The commonly used norm in Asian countries is: 'one bore well of 150 mm diameter and up to 60 to 100 metres depth, fitted with a hand pump, for providing water supply to a population of 250 or less'. Such a bore well is often the 'least cost solution' for providing protected drinking water supply to rural communities, as the cost works out at only around US$0.48 per person per year, even after allowing for interest, depreciation and maintenance at 16% on the initial capital cost of about US$ 750 (Limaye, 1995). The quality of water is usually good if the bore well has a casing pipe up to about 6 metres depth and if the surroundings are kept clean. In some granitic areas the problem of fluorosis is present but small-scale water treatment plants have now been developed.

The experience from several Afro-Asian countries, in programmes undertaken by Government agencies and by NGOs and voluntary organisations, for providing drinking water bore wells in villages, indicates that the following factors are of crucial importance for successful implementation:

(i) selection of a drilling site should be done in the presence of village leaders, after careful hydrogeological and, if necessary, geophysical resistivity survey;

(ii) participation of the local rural community should be encouraged at all stages of the programme, such as providing labourers during the preliminary survey, preparing an access road for the drilling rig, providing food for the drilling crew, assisting in the construction of a concrete slab around the bore well and looking after proper operation and maintenance of the pump.

(iii) a facility which is available 'totally free' often gets neglected. The villagers should therefore, be required to contribute an amount between 5% to 20% of the cost of bore well and pump. If this is not possible for poor communities, they should at least volunteer their labour;

(vi) the village council should collect a monthly fee per family for maintenance of the pumps;

(v) women must be involved in the management, through proper representation in the Village council. Also, arrangements must be made to train some of the young men and women in the village in hand-pump maintenance.

15.6 Exploration

Exploration for locating sites for well digging and drilling in hard rock terrain is vitally important for successful completion of irrigation or drinking water supply projects, because hard rock aquifers are not extensive and their properties vary in short distances. Basic exploration is done by collecting topographical and geological maps, aereal photographs and satellite images, if available, and by conducting a hydrogeological survey during which the following data and information are collected:

(i) Inventory of existing wells. Their depth, diameter and yield and the type of strata encountered. Elevation of the water table. Area irrigated by each well. Type of pump and pumping schedule. Seasonal fluctuation of the water table.

(ii) Rainfall and drainage patterns.

(iii) Lithological units encountered in wells. Strike and dip in sedimentary strata. The thickness of soft overburden and its relation to topography. The extent of fissuring in hard rock and its relation to topography.

(iv) The sandy or rocky nature of the stream or river bed. Whether the stream is seasonal or perennial. The prospects of attracting influent seepage from the stream to a pumping well on the bank.

(v) Shifting and meandering of river. Erosional or depositional features on the river bank. Evidence, if any, of rejuvenation.

(vi) Locations and discharge of natural springs, if any, in the area.

(vii) Locations of surface water reservoirs, if any, in the area. Possibility of receiving recharge during the dry season from surface water reservoirs and/or the irrigation canals leading off from the reservoir. The possibility of obtaining recharge from deep percolation below the root zone in the irrigated area.

(viii) The occurrence of dykes, pegmatite veins, etc. in the area and their nature as groundwater conduits or barriers. Whether there are any good wells upstream from dykes. The direction of any preferred fracture orientation in the area as observed from rock exposures and strata encountered in dug wells.

(ix) Correlation, if any, between the lineaments observed in air photos or satellite images and the locations of successful wells in the area or patches of dense natural vegetation in an otherwise sparsely vegetated landscape.

(x) Variations, if any, in the quality of groundwater along its general flow direction.

(xi) Whether there are any erratically successful or erratically failed wells, which do not fit into the conceptual model of groundwater occurrence in the area. Such wells indicate discontinuity and lateral variation in the aquifer.

Such observations and information are useful in delineating promising zones for groundwater development in a sub-basin. Geophysical resistivity or electromagnetic surveys can then be carried out in these zones for selection of suitable well sites. Given that there are lateral variations in the strata, Wenner profiling is more useful than Wenner or Schlumberger sounding. Profiling is carried out with electrode spacings between 20 to 50 metres, to locate a fractured, low resistivity zone in the hard rock covered by a soft mantle. To find out the fracture orientation, azimuthal resistivity survey can be carried out over the low resistivity zone. In such a survey, resistivity readings are taken around one central spot, with the same electrode configuration but in different directions.

Even within a fractured zone, the intensity of fracturing, interconnections, apertures, infilling matter and recharge from the overlying phreatic aquifer determine the quantity of water available in a dug well or bore well. Tension fracture zones usually have a higher storage capacity, while shear fracture zones could be tight or permeable, depending on their stage of development (UNESCO, 1984).

15.7 Recharge augmentation

With the increase in pumpage due to new wells in a sub-basin, the water table falls and the effluent drainage from the sub-basin is reduced. In the hard rock areas, where both the total storage of groundwater and the average residence time are small, the system is much more sensitive to variations in pumpage and recharge than a similar system in alluvial or carbonate aquifer areas. As mentioned earlier, it is advisable to start soil and water conservation and recharge augmentation activities concomitantly with groundwater development schemes. Some of these activities, such as hill slope trenching, contour bunding, afforestation, gully plugging, are useful in increasing the infiltration to a groundwater body during the rainy season. But the geometrical factors of the sub-basin and the thickness and storativity of weathered rock and fractured rock set a limit to the recharge that can be accepted in the rainy season.

Many developing countries have a well defined rainy season in the year which is followed by a prolonged dry period. In such a climate, a sizable portion of the recharge received in the rainy season may leave the sub-basins by way of effluent stream flow and groundwater outflow, within the first few months of the dry season. Water scarcity may thus occur towards the later months of the dry period, which in the monsoon climate is the summer season. It is therefore necessary to undertake activities which would retard the groundwater outflow and which would lead to recharge to the groundwater body during the earlier months of the dry season.

Construction of underground impermeable bunds (low dikes) across stream beds is a useful technique in semi-arid regions for retarding the groundwater outflow. Construction of a percolation tank by putting an earthen bund with side waste weirs across a stream is also very useful, because during the dry season the water stored behind the bund during the rainy season gradually percolates to recharge the groundwater body. A typical percolation tank has a bund of 8 to 10 metres height and a catchment area of about 20 to 50 km^2. Ideally, the water stored in the tank should percolate within first 3 to 4 months of the dry season so that the shallow water body is not exposed to excessive evaporation rates in summer months. In western India, thousands of percolation tanks have been constructed in semi- arid regions to augment recharge after the rainy season is over. In this drought-prone area, construction of a percolation tank is also a relief measure preferred by the government authorities, because it provides employment to about 2,000 people for 3 to 4 months in a drought year. The percolation tank becomes operative from the next year's rainy season.

The storage efficiency of a percolation tank is the ratio of the volume of water stored in the tank at the end of rainy season to the volume of runoff water available from the catchment. The percolation efficiency is the ratio of the volume of water percolated to the volume of water stored. The overall efficiency is the product of the above two efficiencies and is around 40 to 70% for percolation tanks constructed at technically suitable locations (Limaye and Limaye, 1985).

Occurrence of exposed hard rock in the tank bed impedes percolation. Similarly, silting in the tank bed over the years reduces both storage and percolation efficiencies. Regular desilting is therefore necessary. A percolation tank and a few smaller stream bunds and underground bunds is an ideal combination for augmenting recharge in a sub-basin.

An additional benefit of percolation tanks in coastal areas is to improve the quality of water in drinking water wells excavated on the downstream side of the tank. In granitic areas, where a high fluoride content in groundwater is a problem, dug wells near the percolation tanks show much less fluoride and are preferred for obtaining drinking water supply.

Recently, a novel experiment in recharge augmentation was implemented in semi-arid, basaltic terrain in western India, by a voluntary organisation. In this experiment, surface water flowing in effluent streams during the early part of the dry season was lifted by the farmers, using their pump sets and was delivered into several dug wells in each stream basin. About 100,000 dug wells were thus charged with water that would have left the area as surface flow in a few weeks after the rains. The residence time was prolonged to more than a few months. The beneficial effects of this experiment were felt by the farmers during the summer season. The wells which used to dry up in summer started to yield small supplies of drinking water for the family and the cattle. Wells which used to give a small supply in summer started supporting irrigation of summer vegetables on small plots of land.

15.8 Sustainability and pumpage control

Sustainable development is achieved when the quality and quantity of water available from the wells remains unaffected over many years. In hard rock areas, groundwater or the resource itself is modest in quantity, erratic in occurrence and sensitive to changes in pumpage and recharge. It is therefore not easy to ensure sustainability of all the wells over decades because new wells are constructed each year and the pumpage in a sub-basin increases year by year. Some of the sub-basins become over-developed and the yields from the wells decline due to mutual interference and general depletion leading to a permanent decline of the water table. Under such conditions the drinking water supply wells need to be protected first. This can be achieved either by preventing the construction of new wells within a specified distance from a drinking water supply well or by giving the government authorities a right to acquire any private well in the vicinity, after paying due compensation, to provide water supply to the village.

However, digging and drilling of new wells cannot be stopped in a sub-basin to protect the yields of the existing, private irrigation wells, because this would amount to of a scarce resource being captured and others being permanently denied an equitable share of the resource. Recharge augmentation efforts should receive priority in over-developed sub-basins, in co-operation with the farmers, their cooperative societies and financing institutions. It is also desirable to incorporate voluntary agencies and NGOs in these efforts and through them to advise the farmers to reduce wastage of water. If these efforts are not enough, pumpage control may have to be imposed.

Pumpage control is a negative way of management but when it has to be imposed, it should only be through mutual monitoring by farmers or by local council at the village level. The concept of sustainability in such a case may consider a period of only about 7 to 9 years in which a farmer usually recovers his investment in constructing the well. The owners of high-yielding wells should be the first to cut down on their pumpage and pump only an equitable share, so that other farmers may also dig/drill new wells. This is more likely to be successful through persuasion and social pressure rather than through any rigid legislation. In complex, anisotropic and discontinuous hard rock aquifers, any rigid legislation is technically unsound and may merely lead to endless and futile court battles.

15.9 References and additional reading

LESS, C.; ANDERSEN, N., 1994. Hydrofracture; State of the Art in South Africa. *Applied Hydrogeology,* Vol. 2, pp. 59–63.

LIMAYE, S. D., 1995. *Groundwater Management in Semi-Arid, Hard Rock Terrain in India, (Keynote Address).* Proc. 2nd South Asia Geological Congress, Geological Society of Sri Lanka, Colombo.

——, 2000. *Some Aspects of Sustainable Groundwater Development*. Abstract, Special Symposium B-7, Proc. 31st International Geological Congress, Rio de Janeiro, Brazil.

LIMAYE, S. D.; LIMAYE, D. G., 1985. Groundwater Development in Deccan Traps of Western India. In: *Proc. International Workshop on Rural Hydrogeology and Hydraulics in Fissured Basement Zones.* Dept. of Earth Sciences, University of Roorkee.

UNESCO, 1984. *Groundwater in Hard Rocks*. Studies and Reports in Hydrogeology 33, UNESCO, Paris.

16 Hydrogeology of volcanic rocks

16.1 Introduction

Volcanic rocks constitute thick and widespread formations in many areas of the world such as Eastern Siberia, the Ural mountains, the Deccan in India, the Paraná basin in eastern South America, the Permian basin of Siberia or several areas in central western United States. In large regions of such areas, as well as in many archipelagos and islands, such as the Canaries, Azores, Iceland, Hawaii and Réunion, volcanic rocks are the dominant or only significant geological formation around and most of the available groundwater resources are held in them. In other cases, volcanics are simply one of the local geological formations, below, interlayered or topping other geological formations, as in Sicily, or in many cases in Mexico, Central America and the Andes. Their hydrogeological role as aquifers, aquitards or even aquicludes depend on the relative water-bearing and hydraulic properties and location of the volcanic rocks with respect to the other formations.

Various books deal with volcanic rocks (e.g. MacDougall, 1988; Middlemost, 1985; Fisher and Schminke, 1985; Araña and López-Ruiz, 1974; Araña and Ortíz, 1984; Francis, 1993) and how they are related to global tectonics (Smellie, 1994). More general descriptions can be found in standard geology books (e.g. Duff, 1993; Aubouin et al., 1981).

Specialised literature exists on hydrogeology of volcanic rocks is scarce, although the topic is explicitly considered in some widely used textbooks. Chapters on volcanic rock hydrogeology can be found in Davis and De Wiest (1966) and Custodio and Llamas (1983). Details on coastal and island volcanic aquifers can be found in Falkland and Custodio (1991). A specialised report (Custodio, 1978) largely summarises and enlarges the contents of the UNESCO sponsored International Symposium on Volcanic Rocks and Islands Hydrogeology, which took place in 1974 in Arrecife de Lanzarote (published by CEDEX, 1987), and comments on the experience up to that time in the Canary Islands, including the UNDP-Government of Spain project on groundwater resources in the Canaries.

16.2 Volcanic rocks and formations

Volcanic rocks cover a wide range of chemical and mineralogical compositions. Generally, volcanic rocks are basic, with a low silica content (basalt and basalt-like rocks), some are rich in plagioclase (tholeiites) and others may be rich in olivine (alkaline basalts). A high silica content is found in acidic rocks, such as rhyolites, and andesites and phonolites have an intermediate silica content.

Molten rock, called magma, includes variable quantities of dissolved gases (mainly CO_2) and water vapour. Magmas may be fully melted matter or a mix of melt and crystals. Basic magmas are hot (about 1100°C), present a relatively low viscosity and typically have a low dissolved gas content. Acidic magmas, which are not as hot (900°C), are much more viscous and often have a higher dissolved gas content. Magmas may vary from molten earth mantle material, more or less chemically differentiated due to partial crystallisation and separation of the denser constituents before being ejected, to melted earth crust that has been carried down to great depths in plate subduction zones. The latter are acidic, and the former are dominantly basic. Variable differentiation and mixing of rising deep magmas with melted crust produce a wide range of magma and volcanic rock compositions, with conspicuous changes in space and along time.

The deep melt is lighter than the surrounding rocks and slowly ascends by buoyancy. Most often it accumulates in large magma chambers, some of them relatively shallow (a few km deep), from which the magma flows out through tension cracks (fissures). Generally only a fraction of the magma contained in the chamber flows out, the rest remains in place. After cooling, it may leave a coarse-grained rock mass or it may be intruded through enlarged fissures into the surrounding rock as dykes, sills, or cone sheets. The surrounding rock, which may or may not be of volcanic origin, may be deeply altered (metamorphosed) by heat and/or hot gases and water, and sometimes intensively intruded by dykes. There are alteration products that correspond to a wide range of underground conditions, from hot to cold ones.

The magma may escape from the chamber through fissures but the flow often concentrates in short sections when approaching the land surface and pours out at discrete points as volcanoes. It seems that the large outflows that formed the large volcanic plateaus, such as the Deccan in India or the Paraná in eastern South America, were intense fissural emissions, not devoid of point outflows. Important recent volcanic episodes show volcanoes along lines representing the main regional fissures, as in the seventeenth and eighteenth century major effusions of Lanzarote, Canary islands. In other cases, cylindrical volcanic vents can be seen after erosion has exposed them, as near Stuttgart, Germany. But interpretations differ from researcher to researcher. Magmas poured out in rift zones and into transform faults of sliding crustal plates are dominantly basaltic, whilst those poured out in plate subduction areas are dominantly of the acidic type. Basaltic magmas form the extensive plateaus of Deccan and Paraná, and others in Siberia, South Africa and Central Europe, as well as the large oceanic ridges, whose top emerge as islands such as Iceland, Azores, Ascension, Hawaii or Tahiti. Other dominantly basaltic islands such as the Canaries and Madeira seem to correspond with transform faults. Acidic magmas form large masses along the western part of North, Central and South America, and around the Mediterranean sea, and also form long island arcs such as the Kuriles, Nippon and Philippine islands, and dominate in Java and Sumatra (Indonesia). The upwelling of mantle magma plumes, sustained by buoyancy, persists through geological times and leaves a trace on top of the plates. The age of the volcanic deposits increases as one moves away from the present point of eruption.

Submarine volcanism is more extensive than continental volcanism, but is of little hydrogeological interest since the formations are mostly impermeable due to intense alteration. They appear on continental land masses as relatively old, compact materials after having been tectonically uplifted, such as some important volcanic deposits dating from the Permian and the Keuper.

The build up of a volcanic oceanic island is rather complex (Carracedo et al., 1997). First a submarine edifice rises from the ocean bottom and comprises extruded and intruded materials. Later on, when the submarine edifice attains the sea surface eruptions are explosive and produce loose, fragmented materials which are easily transported downslope by shallow water erosion. If volcanism persists or the materials are rised, finally an island is formed, with an erosive surface and the possible outcrop of the resistant inner core of intrusive bodies and densely dyke injected volcanics. Afterwards subareal volcanics pile up and form the island. Vertical movements may be relatively important in islands due to regional tectonics (Watts et al., 1997) and the bulging due to magma chamber emplacement. Thus, relatively young submarine volcanic rocks and intrusive bodies may crop out, as in some of the Canary islands 'volcanic complex' formations. The sea floor on which these islands lie is about 3,000 to 4,000 m below sea level; Miocene submarine volcanics may be found 1,000 m above present sea level, while, due to these important vertical tectonic movements subaerial volcanic rocks have been found at a depth of 2,000 m below sea level.

The magma that pours out through the volcanic vents suffers a loss of pressure that allows dissolved gases and water vapour to exsolve. The molten, more or less degassed rock forms the lava, which spreads away from the outflow area. Gases may exsolve quietly or violently in the form of bursts, in which part or much of the magma is fragmented and thrown into the air, forming pyroclasts (tephra) varying in size from coarse blocks to fine dust (ashes or cinder). Pyroclasts may be deposited as cooled fragments which remain loose, or as hot fragments which may be welded to some extent. The block, gravel and sand size pyroclasts (lapilli or 'picon') pile up around the outflow area, forming a volcano, which is a symmetrical or deformed cone, depending on wind strength and land slope. The volcano may include lavas and intruded material. It is not rare for lavas to outflow at a lower altitude instead of from the crater-like summit of the volcano. Large gas outbursts and explosions produce large quantities of ash which the wind may transport far away. Volcanism in the Andes has supplied large quantities of volcanic dust that is now included in the sediments that cover large areas in Argentina and Uruguay, and plays a significant role in groundwater chemistry. The finest particles produced in large explosions may reach the stratosphere and extend over a large region and even the entire earth. The change in atmospheric dust content and the gases released during large volcanic events may effect the earth's climate for some time (Robock, 2000). The sudden outpouring of gas-rich magmas, especially of the acidic and intermediate type, in the response to a rapid loss of pressure may produce a fluidised layer of dust and fragments suspended in the very hot gas which is able to move easily down slope until it settles down and cools off. This forms the ash flow (tuff) deposits or ignimbrites that differ from the ash fall formations that are formed in a cool environment. After cooling, ignimbrites look like lava, especially when alteration and weathering obscure their texture. Large explosive events may also produce extensive and thick formations of volcanic conglomerates. The so called 'Roque Nublo' formation once covered most of the island of Gran Canaria and filled previous relief.

In the following sections subaerial (continental and island) volcanism will be considered, focussing on the relatively young formations of Quaternary, Tertiary and in some cases Mesozoic age, because of their hydrogeological importance.

From the hydrogeological point of view, the most important volcanic formations are lavas, pyroclastics, ignimbrites and dykes. Dykes are more numerous deeper and closer to the magma chambers. When considering lava and ignimbrite flows, it is important to distinguish between the dense, slowly cooling inner part and the highly vacuolar and brecciated top and bottom (scoria) of a flow. Some lava flows are almost devoid of top scoria (pahoehoe type) but others are formed almost entirely by scoria and blocks (aa type). This depends on viscosity, cooling speed and gas content.

Fluid lavas spread laterally, forming discrete lava flows on sloping ground, and when the land is flat, collect into large lava fields. In this way hundreds of square metres of lava sheets, commonly 1 to 10 m thick, accumulate. In Lanzarote (Canary islands) in the early seventeenth century, a 6-year-long series of eruptions covered more than 200 km^2 and deposited up to 300 m of volcanics in some areas. On the same island subaerial volcanics may be up to 2,500 m thick. Viscous lava tends to pile up around the outflow point. In some cases only a plug of semi-solid hot rock has extruded, accompanied by sudden outbursts of ash flows. In the Paraná basin, major basaltic outflows covered thousands of km^2 of land with very extensive pyroclast-free lava sheets that are in some cases up to 1,000 m thick.

Near the volcanoes the ratio of lavas to pyroclasts is highly variable, from high in some basaltic shield volcanoes, such as those in Hawaii, to very low in explosive eruptions, mainly of the acidic type, but also of the basic eruptions when groundwater increased the gas content of upflowing magma (phreatovolcanoes). Explosions not only form large quantities of pyroclasts but also take broken parts (xenoliths) of deep volcanic or non volcanic formations to the surface.

Volcanoes commonly have steep sides, close to the geotechnical slope equilibrium. Thus erosion, earthquakes, uplifting due to deep magma emplacement and volcanic explosions may easily start large landslides that leave deep scars (Elworth et al., 1999). This is well known in Hawaii and in the Canaries. The pressure and friction of the sliding mass may form a layer of finely crushed and altered material, sometimes rich in bentonite, called 'mortalon' in the Canaries. Such layers are poorly permeable and difficult to drill, and may be regional hydro-geological boundaries, although they are still poorly known. Large mudflows (lahars) can form when the hot gases and pyroclasts cause the fast melting of snow on top of high volcanoes (Nevado de Ruiz, Colombia; Mount Saint Helene, NW United States of America) or when the explosions are accompanied by intense rain storms and lake water displacement, as in Java. The sudden loss of weight after a major landslide may reactivate volcanism from still hot deep magma chambers, as happened on some of the Canary islands (Tenerife, Hierro).

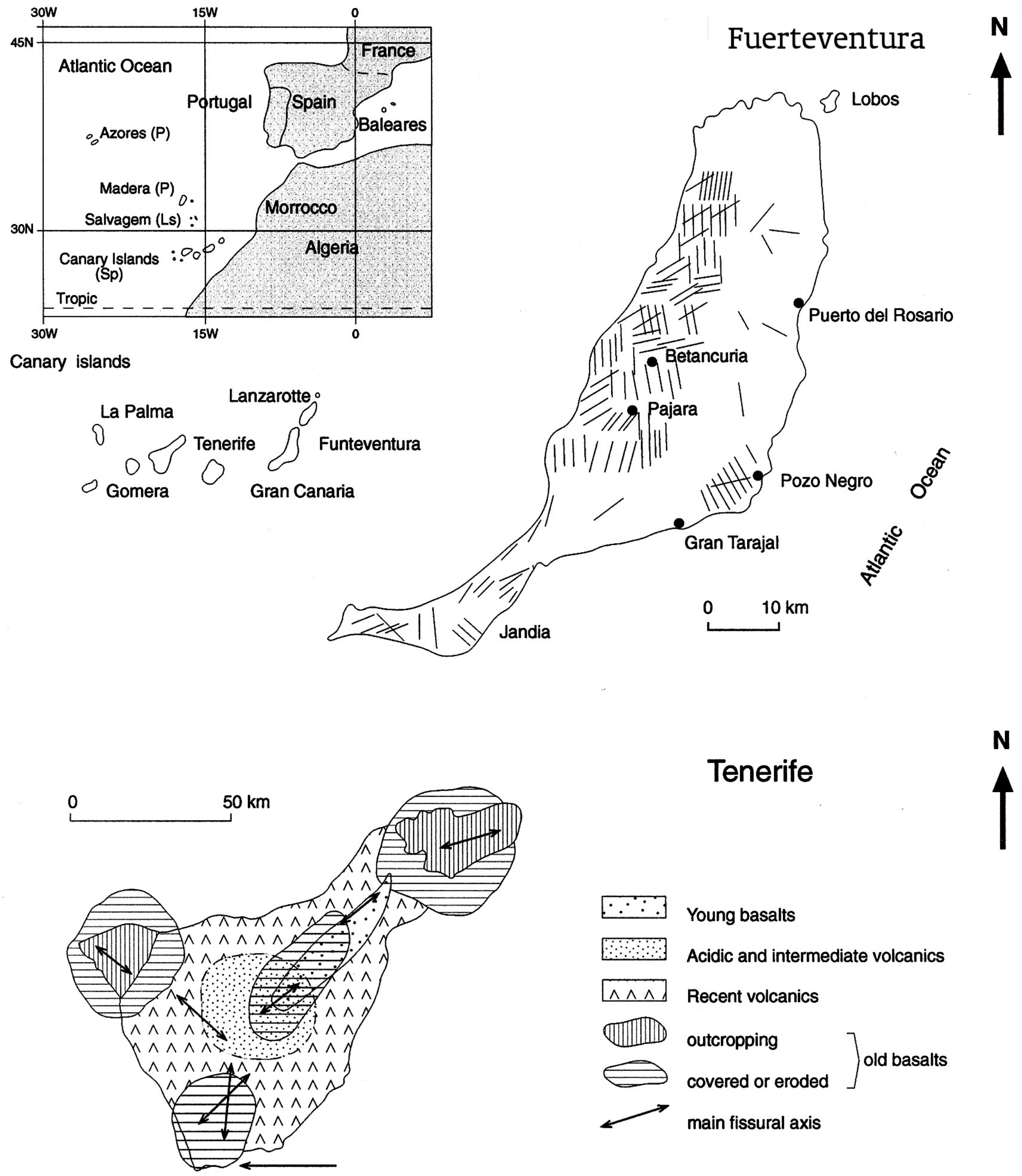
Dykes often present one or two main orientations. They tend to be vertical but actually their thickness and slope accommodate the host rock conditions. The horizontal length of the dykes varies from a few tens of metres to several km and the width varies from some cm to a few metres. Close intrusive bodies and around long-lasting active calderas swarms of dykes are common, and their volume may be larger there that of case rock. Some magma may be intruded horizontally, interlayered between existing formations, forming a sill or horizontal dyke, a few to some tens of metres thick and up to many km^2 of surface area.

Over a long time (up to several million years) the slow cooling of magma chambers up to many km^3 in volume, releases large quantities of hot gases, mostly CO_2 and water vapour, which play an important role in groundwater chemistry and in modifying rock properties by thermal metamorphism.

There is an important distinction between volcanic formations close to the eruption points, where pyroclasts as well as other features that will be described below are found, and volcanic formations far from these eruption points. In the former case coarse pyroclasts and other features linked to the eruption area predominate. In the latter case the volcanic deposits are mostly represented by the accumulation of gently sloping flows with occasional sediments and weathered layers sandwiched in between. The most important of these formations are basalts, such as those in the Paraná and in the Deccan. These formations are also called 'traps', derived from the Scandinavian word for stairs, referring to the appearance of these horizontal layers after erosion.

Volcanism is a constructive (positive) agent, but erosion is a negative agent that tends to destroy what has been formed. Volcanic formations are easily erodible. Steep slopes around eruption areas are prone to fast destruction by gravity and running water, as well as by sea abrasion in coastal areas. This is especially conspicuous on islands, in which building episodes (eruptions) alternate with destructive (erosive) ones during calm stages. The oldest volcanic land forms are more erosion-resistant than the more recent ones, because of the greater dyke density and higher degree of consolidation (Fig. 16.1). The valleys and depressions of old landscapes carved into relatively poorly permeable volcanics may later be fully or partly covered by new outpourings and fallout (Fig. 16.2). Thus rock hydraulic properties change in both vertical and horizontal directions. Figure 16.2 shows a large variety of structures that produce contrasting

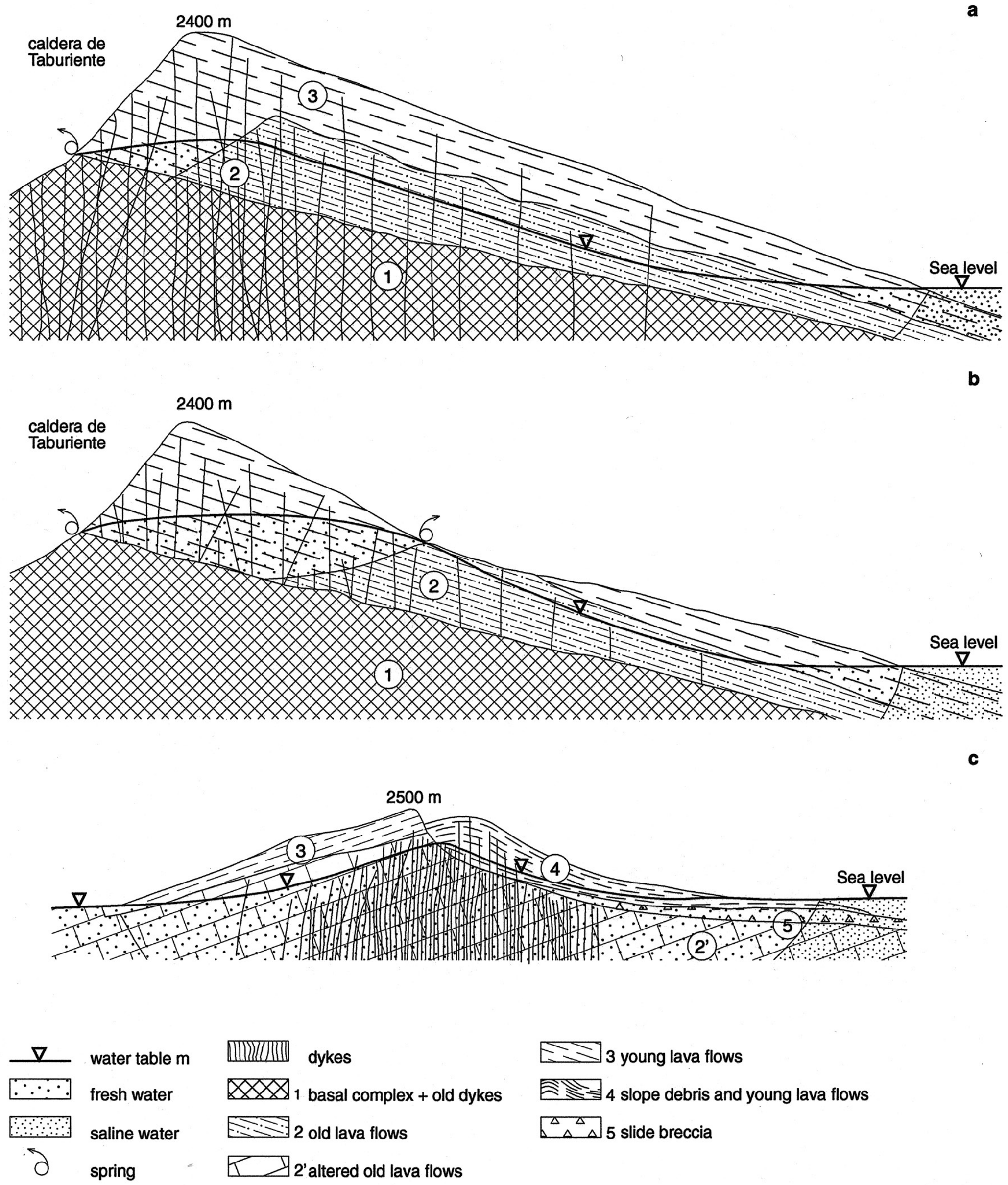
Figure 16.1 Structure of volcanic formations affecting groundwater as exemplified by the Canary Islands, Spain (Custodio, 1978)



The upper figure represents Fuerteventura which is a high, deeply dissected massif, with poor permeability, densely intruded by dykes (Araña and Carracedo, 1979). It resists erosion due to the massiveness of the old, mostly submarine volcanic complex. The remaining part of the island is formed by permeable, recent volcanics extending seaward and filling the spaces between the dykes.

The lower figure represents Tenerife island in which the remnants of old (Miocene) deeply dissected volcanic structures outcrop at the ends of three main fissured areas. At the confluence point young volcanic deposits have covered part of the old formations and spread towards the sea, filling the intermediate depressions. A similar structure can be found in Mauritius (Jawaheer et Proag, 1978) in which a deeply eroded central volcano with a caldera was later filled by younger volcanics; the present island is a plateau of recent lavas from which the old volcanics outcrop forming radial ridges.

Figure 16.2 Simplified cross section through island volcanic formations



A and B represent two cross sections from La Palma island (Canary Islands, Spain, from the Taburiente caldera in the centre, towards the north (modified after the studies for the La Palma island water plan). Resting on very poorly permeable, densely dyke intruded submarine and subaerial old volcanics (1), there is a basaltic cover (2) flowing from the flanks of the of the Taburiente caldera, which is less permeable than the younger basaltic flows (3) that fully cover it or leave only small windows. The water table is inside formation (2) except near the top, where it is in formation (3). In case A, a small part of the recharge is discharged through springs inside the caldera, the remainder going underground towards the coast, if not tapped by wells or galleries. In case B, a large part of the recharge outflows through large midslope springs. The water table slope decreases towards the coast due to radial flow pattern, and when it is in formation (3). C shows the effect of large landslides (4) which produce a low permeable breccia (5), which in Tenerife (Canary Islands) is locally called 'mortalón' (modified after the studies for the Tenerife island water plan). The water table attains high altitudes in the low permeable, dyke-intruded and altered old materials ('core'), and groundwater mostly flows through the thin overlying layer of younger volcanics.

hydraulic properties, including the interbedding of sediments and palaeosoils. They have to be taken into account when dealing with groundwater occurrence and flow. Uplifting allows erosion to exhume deep structures and even the underlying magmatic chambers and submarine volcanic formations.

Large violent eruptions produce a fast reduction of pressure in the magma chamber and may cause the roof of volcanic or sedimentary formations to collapse, producing a caldera bounded by ring faults and crossed by radial faults. Afterwards sediments and volcanics from subsequent activity tend to fill up the caldera. Figure 16.3 shows two examples.

Volcanism is associated with land elevation disturbances related to the emplacement of magma (uplifting) and to the sinking after the eruptions, as well as to the general tectonics of the area. Furthermore there are intrusions of dykes, landslides, explosions, cover collapse after sudden outbursts and to relief internal pressure, differential compaction of collapsible porous materials due to overburden weight, stress relief due to erosion and retraction due to cooling. All this produces fissures, joints, cracks and discontinuities.

16.3 Hydrogeological properties of volcanic formations

Volcanic formations are highly variable and so are their hydrogeological properties, which cover the full range of possible values found in natural formations. The scale of observation is important. Heterogeneity is great at small-scale (say up to metres) but is smoothed at large-scale (hundreds of metres) in large and extended formations.

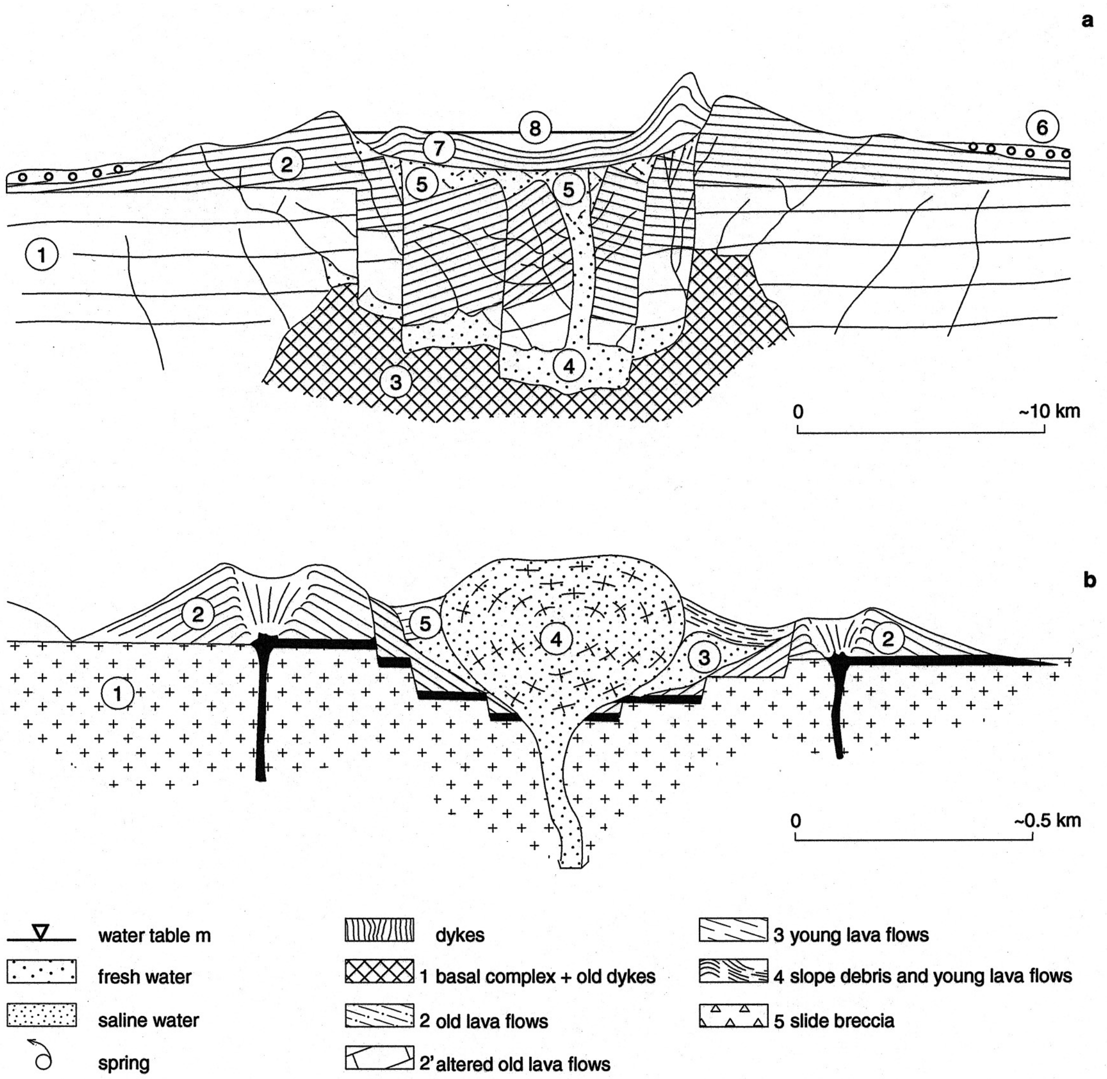
Total porosity of volcanic formations is generally high, due to voids created by exsolved gases and to the frequent scoriaceous and brecciated parts, as well as to their often clastic nature. Values above 40 percent are not rare in lapilli, pumice and ash deposits. But compact lavas and dykes have less than 5 percent total porosity. In some rocks most of this porosity is due to closed, isolated voids, through which water cannot readily flow unless the rock becomes weathered. The same can be said for closed-end pores. The result is that the porosity available for groundwater flow is often much less than total porosity, especially in fresh (young) volcanics. Specific yield (drainable porosity) is small for fine-grained pyroclastic formations, ashes and weathered rocks. See Table 16.1 for some indicative values.

Permeability varies considerably, from almost zero (less than 0.0001 m/day) to more than 1,000 m/day. Massive lavas and welded ash flows (ignimbrites) present very low permeability, while fresh, highly brecciated lavas and loose lapilli are very permeable. Table 16.2 shows some representative values, but they have to be taken as indicative, since large variations can be expected from site to site and inside a given formation; sometimes secondary permeability is dominant.

The distribution of permeability and porosity is related to the characteristics of the rocks. This is important when explaining the hydraulic properties and when choosing sites for drilling wells to exploit groundwater resources.

Lava flows generally present a massive core with scoria and breccia layers at top and bottom that are intrinsically more porous and permeable. Some lava flows are almost devoid of top layers while others are dominantly brecciated from top to bottom. Some lavas are highly vacuolar, which produces high porosity and often a large fraction of scoria and breccia, while others are compact. A lava flow may lie directly on top of another lava flow that outpoured shortly before, or rest on older materials covered by a weathered zone, even soil. In the latter case the heat from the new lava flow thermally alters (bakes) the underlying materials, which acquire a reddish colour (rubefaction). If a more or less developed soil existed, an often reddish, brick-like layer of variable thickness is formed, commonly known as red layer or 'almagre'. Sometimes it may be traced for many km when exposed in cliffs that cut across the formation.

Figure 16.3 Schematic internal structure of a volcanic massif conditioning the permeability distribution



A. Caldera after the collapse of the magmatic chamber, following a violent episode in which differentiated magma is outpoured. 1. Pre-volcanic formations (old volcanics or other formations); 2. volcanic formations from the magma chamber, generally of basic composition; 3. magma chamber containing scarcely differentiated magma from a deeper chamber; 4. differentiated (more acidic) magma in the top of the chamber, responsible for the violent volcanic episodes whose products; 5. partly fill the collapsed caldera; 6. extra-caldera volcanics thrown out during the violent episodes; they may include lahars (mudflows produced by water) and debris from large landslides; 7. filling of the caldera with erosion material and products of reactivated volcanism; 8. final stages of filling, which may include water borne sediments from the sides. The figure may represent, with due changes, the Cañadas del Teide (Tenerife, Spain), the Crater Lake (Oregon, United States of America), or Mount Etna (Sicily, Italy), which lies on top of folded and thrusted pre-volcanic formations.

B. System of various volcanoes, inspired by the Gran Sarcouy (Auvergne, France), but also representative of the Gomera necks (Canary Islands, Spain). 1. basement, which may or may not be volcanic; 2. volcanic landforms dating from before the violent episode; 3. new volcanics ejected when the area collapsed, thus forming a caldera; 4. ejection of viscous, more acidic poorly permeable lavas that fill the caldera and produce a tight plug; 5. post-eruption sediments or volcanics.

Table 16.1 Common porosity values of volcanic formations. The ranges are only indicative. Values are derived from a wide literature review and the experience in the Canaries

Material	*Total porosity (%)*	*Drainable porosity (%)*	*Comments*
Basalt flows	0.8 – 20	0.1 – 8	Dense to highly vacuolar
Basaltic interflows	20 – 50	5 – 15	Breccia at the top and bottom of a lava flow
Basaltic formations	5 – 40	2 – 8	Increases with content of scoria and pyroclasts
Basalt sheets (traps)	4 – 10	<1 – 2	Several flows, no pyroclasts, moderately old
'Core' basalts	2 – 5	< 0.1 – 4	Thermally altered and dyke-intruded
Conglomerates	2 – 25	1 – 4	Mostly basaltic
Loose pyroclasts	25 – 50	5 – 10	Fresh lapilli and blocks
Ash fall	25 – 40	1 – 5	Relatively fresh
Phonolites	2 – 4	1 – 6	Dense flows
Phonolitic ignimbrites	20 – 60	0.5 – 8	Dense to poorly welded tuff
Pumices	50 – 85	< 0.1 – 1	Non-connected pores, unfractured
Obsidians	~ 0.5	< 0.1	Not a formation
Rhyolites	0.1 – 30	0.5 – 5	Dense to vacuolar
Rhyolitic ignimbrites	15 – 70	0.5 – 10	Dense to poorly welted tuff
Rhyolitic interflows	30 – 70	1 – 5	Breccia at the bottom of a lava flow
Lahars (mudflows)	20 – 35	< 0.5	Heterometric, variably consolidated
Volcanic soils	40 – 60	<1 – 5	Variable

Note: Total porosity and drainable porosity decrease with age.

Its continuity depends on the shape of the previous land surface and soil cover, and the extent of the baking lava flow. There are often changes of thickness and permeability, and discontinuities, and it may be crossed by joints and dykes.

The inner part of ignimbrites (ash flows) tends to be compact and dense, like lavas, while the outer parts become more porous and loosely consolidated towards the top and bottom, often grading into breccias. Here some primary permeability may be found while the remaining rock is almost impervious.

Both lavas and ignimbrites develop joints due to shrinking during cooling. The joints run from top to bottom forming vertical faces in a hexagonal prism-like distribution, spaced ten to hundreds of centimetres. There are usually also some horizontal joints. Generally joints are deformed (curved joints, irregular patterns). These joints are primary tight.

When lavas, hot pyroclasts and volcanic tuff engulf tree trunks from pre-eruption forests, what remains after the wood has been charred are tube-like cylindrical moulds.

In gently sloping land, when a lava flow stops moving because its surface has consolidated, but the inner part is still fluid, sudden outbursts of the molten inner part may happen after internal pressure increases. This forms a downflow extension of the lava flow, leaving behind an elongated cavity below the hardened crust. Most often the roof collapses and the cavity becomes filled with rubble and, later on, by materials from new lava flows, but sometimes a more or less

Table 16.2 Common ranges of permeability for water at normal temperatures (10 to 25°C) and values of transmissivity. The ranges are only indicative. Values are derived from a wide literature review and experience in the Canary Islands

Material	*Perme-ability (m/day)*	*Reported trans-missivity values (m^2/day)*	*Comments*
Basalt flows	10^{-5} – 10	2 – 100	Dense, unfractured to young, highly brecciated
Basaltic formations	0.01 – 20	2 – 100	Several flows with pyroclasts, aa to pahoehoe types
'Core' basalts	0.001 – 0.05	0.1 – 10	Thermally altered, dyke intrusion
Basaltic traps	0.001 – 10	1 – 50	Several flows, no pyroclasts, moderately old
Recent basaltic formations	1 – 10^3	100 – 10^5	Unaltered
Conglomerates	0.01 – 0.5	3 – 50	Mostly basaltic
Loose pyroclasts	0.1 – 50	10 – 500	Young
Ash fall	0.01 – 0.1	0.5 – 5	Relatively fresh
Andesites	0.1 – 1	2 – 200	
Phonolites	0.1 – 20	20 – 1500	Effect of major fissures
Phonolitic ignimbrites	10^{-6} – 0.01	0.1 – 10	Welded to fractured
Trachy-syenites	0.01 – 0.1	1 – 5	
Rhyolites	0.01 – 0.1	0.1 – 10	
Rhyolitic ignimbrites	10^{-4} – 0.01	0.02– 0.4	
Alluvium and terraces	1 – 10	2 – 200	Poorly sorted, derived from volcanics
Lahars (mudflows)	0.01 – 10	1 – 100	

Note: Permeability decreases with increasing age and degree of thermal alteration. Old volcanics are generally poorly pervious, but after exhumation they develop fractures. Reported transmissivity values may not represent the whole formation but the part penetrated by the well.

long, complex cave remains for a while. All these are short-lived features in recent lava flows, easily collapsible or fillable, and generally above the saturated zone. In areas close to sea level or near large lakes, or when recharge is very intense, saturated local permeability may be very high. In Lanzarote island (the Canaries) advantage has been taken of these circumstances in lava flows extending beyond the former coast line; wells have been drilled to get large flows of sea water free from sand and algae for feeding desalination plants.

Young pyroclasts are often loose and very permeable, but in other cases they form more or less cemented volcanic agglomerates and tuff, very porous and prone to collapse under weight. Permeability may vary from great to almost impervious. Other volcanic structures like dykes, sills and necks, have low porosity and very small permeability.

It is very important to note that what has been said so far refers to primary volcanic rock characteristics, e.g. the characteristics of the rock when formed. Subsequently there may be important changes that are able to transform these properties. The two main causes of change are internal adjustments and rock alteration, which includes the 'ageing' effect. These secondary properties may dominate the hydraulic behaviour of the volcanic formation.

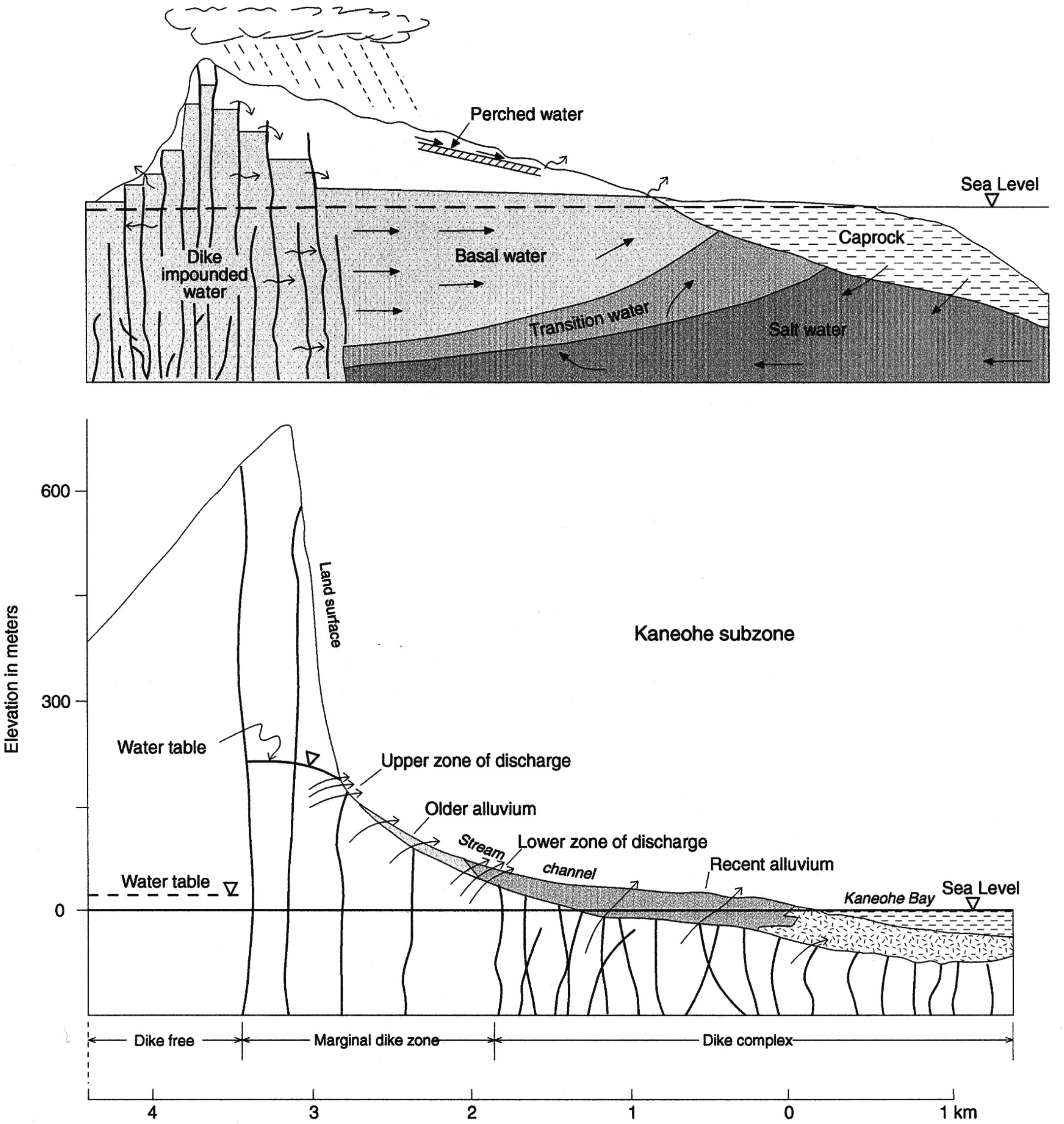
The designation 'internal adjustments' includes a series of poorly studied circumstances which generally produce an increase in the large-scale permeability, with little change in total porosity. They include small differential movements due to compaction of collapsible formations under increasing load and after alteration, the bulging or shrinking in response to deep magma emplacement or emptying and pressure relief. They are also the result of the tectonic response to deep faults, large landslides, the stress produced by dyke intrusion, the stress relief after erosion and so on. The result is that joints are subjected to small relative displacements which widen them, and new joints are created. This makes massive lavas and ignimbrite flows permeable, the more frequent the dislocations, the more permeable they become. Fissure widths of some mm are not uncommon, especially near volcanic centres (they are rarer further away) and in some cases they grow centimetres wide. It seems that subvertical fissures dominate. Fissures remain open unless thermal and low temperature weathering line and fill them with minerals. The effect of these fissures is less important in breccias, loose pyroclasts and similar formations, but dominant in harder rocks. The role of dykes deserves special attention. In Hawaii, because of their long linear extent, dykes are often considered impervious walls against groundwater flowing through the more pervious volcanics in between. This is true when the dyke is parallel to the ridge, not fissured due to internal adjustments – generally it is stiff and fragile – and the country rock is permeable, as is the case in Hawaii, where dykes penetrate through relatively fresh volcanics in the very high rainfall highlands (Fig. 16.4). Circumstances may change dramatically with depth or in older volcanics, since then the country rock may have a lower macroscopic permeability than the fractured dyke and, what is more important, the contact between the case rock and the dyke may develop into a fissured zone through which groundwater may flow readily.

The preferential orientation of dykes makes the permeability anisotropic. When perpendicular to the ridge they may facilitate the drainage. But it must be clear that each dyke behaves differently from the others according to the complex local circumstances, and consequently an individual behaviour cannot be generalised, as the mean characteristics may differ greatly from what is found in a particular well or water gallery. This can be seen in the numerous, long water galleries in the Canary Islands, where the most permeable, recent volcanics are unsaturated due to the relatively low rainfall.

Volcanic formations are easily altered, both at the surface and at depth. Their glassy nature, great specific surface and geochemistry explain this. In wet environments thick soils and weathered rock layers may develop under moderate erosion rates. But more important are internal changes. Around magma chambers hot fluids and gases (mainly CO_2) metamorphose the surrounding rocks into a poorly permeable and dense mass; frequently criss-crossed by a dense network of dykes near the magma chamber. The action extends outwards with progressively decreasing intensity, but still with large geochemical changes produced by CO_2-rich, slow-flowing groundwater. The formation of low temperature silicates may often produce a dramatic reduction of permeability. Fissures tend to be filled with new minerals. This is why around the volcanic centres not only are dykes frequent, but the rock forms a core with a very low permeability. The slow release of CO_2 from the magma chamber and from originally closed pores of the volcanic rock is a long lasting process.

The above processes explain why old volcanics are much less permeable than young ones. This is the 'ageing effect'. There is no absolute scale of ageing effects, since the result depends on a series of geochemical factors, including groundwater, but also on the compactness of the rock. Formations such as breccias, lapilli, ash fall deposits and loosely-cemented ash flows are easily altered to a low permeability mass in only a hundred thousand to a few million years, whilst massive lavas and dense ignimbrites are much more resistant to changes at low temperature, conserving some of their primary and secondary properties for a much longer time, up to some tens of million of years. Thus, Pleistocene volcanic tuff and Pliocene block and scoria lava flows

Figure 16.4 Two modes of interpreting groundwater in the volcanic island of Oahu (Hawaii Islands, United States of America)

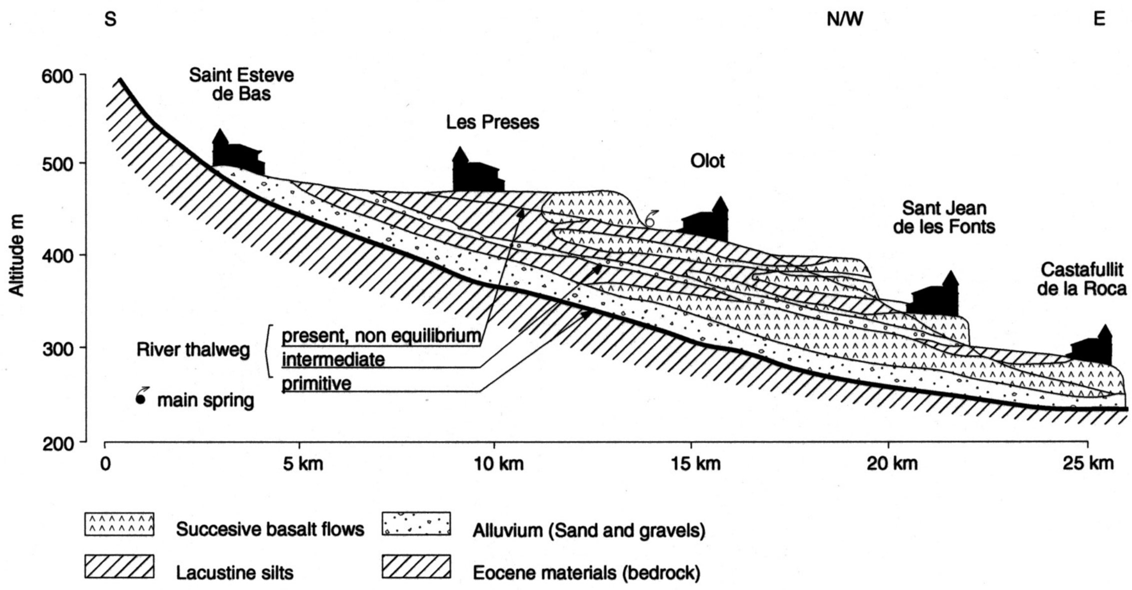


The upper one (from Hunt et al., 1988) shows the assumed compartmental behaviour of the water table in the densely dyke (dike) intruded area, while away from that area a continuous, gently sloping water table is depicted; this basal aquifer does not extend below the dyke impounded area. There are unshown springs in the dyke-impounded water area and all recharge is assumed to be transferred to the basal area, where discharge is along the near coast, either directly or through the cap rock, on top of the marine salt water wedge. Only occasional perched springs are depicted.

The lower figure (from Takasaki and Mink, 1985) shows a more continuous water table and explains in more detail the cap rock of sediments along the coast and offshore, although the salt water wedge is not shown. The sudden water table fall to the left represents the basal water at the other side of the island, without details on how it is connected to the dyke-intruded zone. These two figures do not represent the same cross section of the island. In the first one the dyke-intruded area is far from the coastal area shown at the right hand side of the figure, and in the second one the cap rock covering it is depicted.

(a-a type) may become quite a poorly permeable mass whilst Cretaceous massive lavas may conserve a relative good permeability in distorted joints and near the top and bottom of the flows (pahoe-hoe flows). Volcanics several million years old are generally quite impervious, and generally behave as aquitards if not as aquicludes. The effect is even more conspicuous when they occur with other sediments such as sandstones, as happens in the Jurassic-Cretaceous Paraná basin (Serra Geral basalts juxtaposed to the Botucatù sandstones), or in the Quaternary patchy volcanic formations in Olot (Catalonia, Spain) in which confined fluvial gravels constitute the main aquifer, and altered ash fall and lapilli act as aquitards (Fig. 16.5); the lava flows are only considered good local aquifers when no gravels are present.

Figure 16.5 Cross section along the Fluviá river valley, Catalonia, Spain (modified from Fayas and Doménech, 1974)



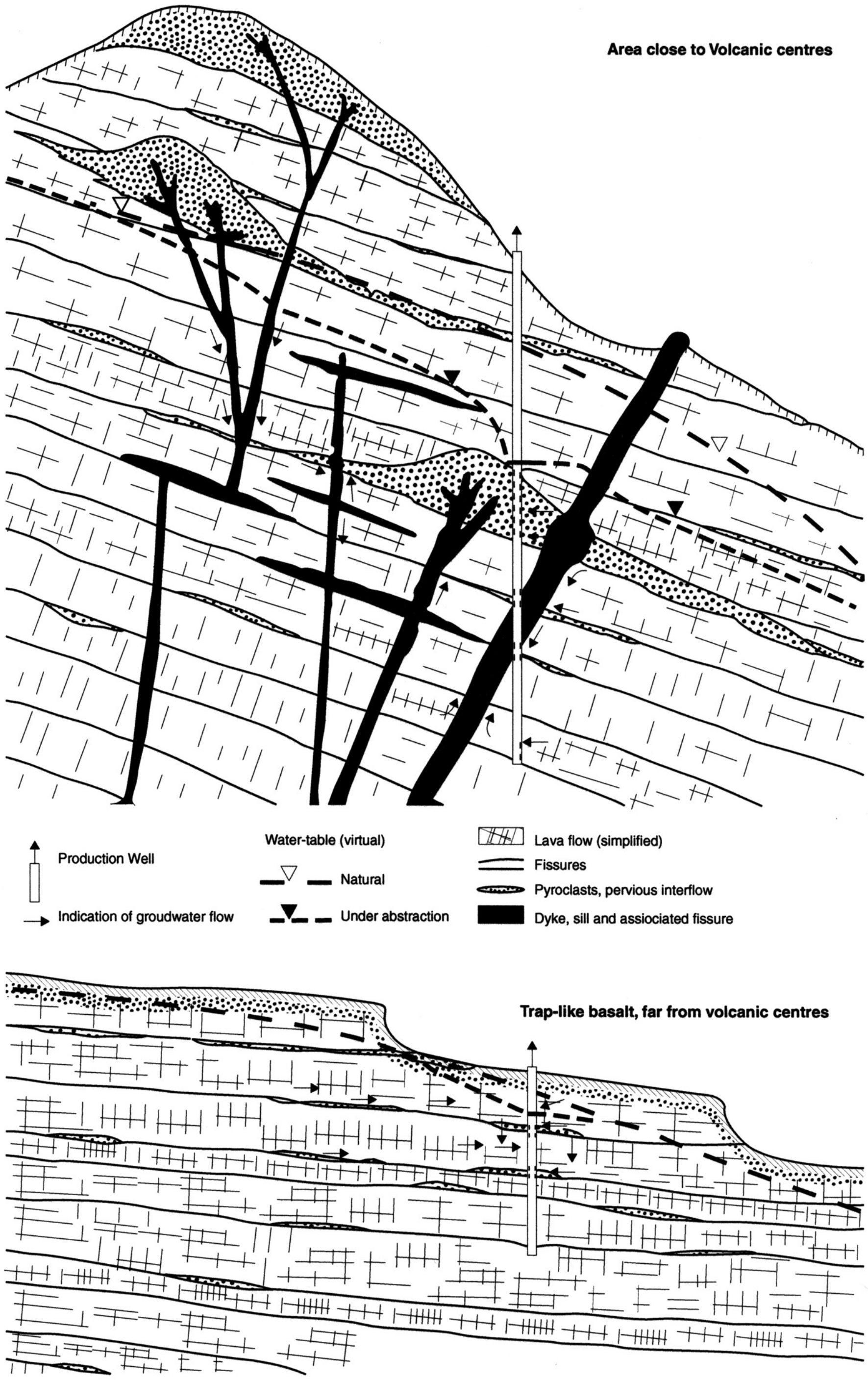
It is shown how successive lava flows from the Pleistocene volcanoes of the Olot area have dammed the valley until the river was able to breach the barrier. The lavas and pyroclasts confine very permeable river terrace gravels. Behind the dam a lake formed, which progressively filled up with clays and silts, which are much less permeable than the volcanics. Large springs appeared when the present river valley cut through the saturated permeable, mainly volcanic, formations. The deep confined river terraces are the more permeable formations and are exploited by means of deep wells. Buried alluvium and terrace deposits may play an important role in volcanic islands as well, as in Réunion (Join et Coudray, 1992, 1993), or in Telde, eastern Gran Canaria island (Cabrera et al., 1992).

16.4 Groundwater flow in volcanic formations

A volcanic formation can be considered as a combination of porous layers and blocks criss-crossed by fissures. When weathering is intense, thick altered zones (regolith) are formed if erosion is not too intense and removes them.

Recharge water originates as soil moisture and tends to move vertically downwards towards the water table. The actual detailed movement depends on the combination of layers, blocks and fissures. In the unsaturated medium most of the flow is probably through the layers

Figure 16.6 Schematic, highly idealised cross sections through volcanic formations



The upper one corresponds to an area close to the eruption centres (inspired by what is seen in cliffs in Gomera and Hierro, Canary Islands, Spain), and the lower one to a piling up of basaltic flows far from the eruption centres (inspired by descriptions of the Deccan traps, India, or of the Serra Geral basalts, Brazil). It shows the 'virtual' water table (it can be determined only from permeable features, correcting for possible vertical groundwater head gradients) and how it changes due to the effect of water abstraction from a well. The arrows show the local groundwater flow (from Custodio, 1989).

and blocks, and not through the fissures, except when recharge is intense, during and shortly after heavy rains, when surface water infiltrates or below perched saturated layers, both near the surface or in depth. The formation of perched saturated zones is not rare when recharge rate is high, and some very poorly permeable layers are found, such as extensive red beds. In sloping areas, seeps and springs appear when such a layer intersects the land surface. The outflow is generally only a fraction of local recharge, the rest following the more or less vertical path toward the water table. Part of the flow of the perched springs and seeps may be locally evaporated or transpired by the vegetation growing at the site. The remaining flow often reinfiltrates downstream if permeable formations are found, but in other situations ravines with perennial flow can be found.

The thickness of the unsaturated zone may vary from shallow (high rainfall, flat areas, poorly permeable materials, or a combination of them) to very deep, sometimes several hundred metres (low rainfall, sloping areas, highly permeable materials).

Water flow in the saturated zone tends to follow preferential paths such as little- altered brecciated zones and open joints, but most groundwater is stored in the bulk of the volcanic formations as slowly flowing water (Fig. 16.6). As a whole, the volcanic formations approach a double-porosity/double-permeability system, but less clear-cut than in other hard rocks such as granite or limestone. The model of storage in blocks and flow in fissures is a better description than that of a granular-like aquifer, but the actual behaviour is somewhere between the two. The behaviour changes from place to place according to the dominant type of volcanic formation, such as lava flows, pyroclasts, ignimbrites or altered volcanics with intruded dykes. Groundwater flow in the Deccan basalt traps is mostly through vertical fissures and horizontal interflow layers, and storage is in the blocks and these interflow layers (Kulkarni et al., 2000).

The actual groundwater flow system in lavas and ignimbrites far from eruptive centres is best described by the fissure and block model.

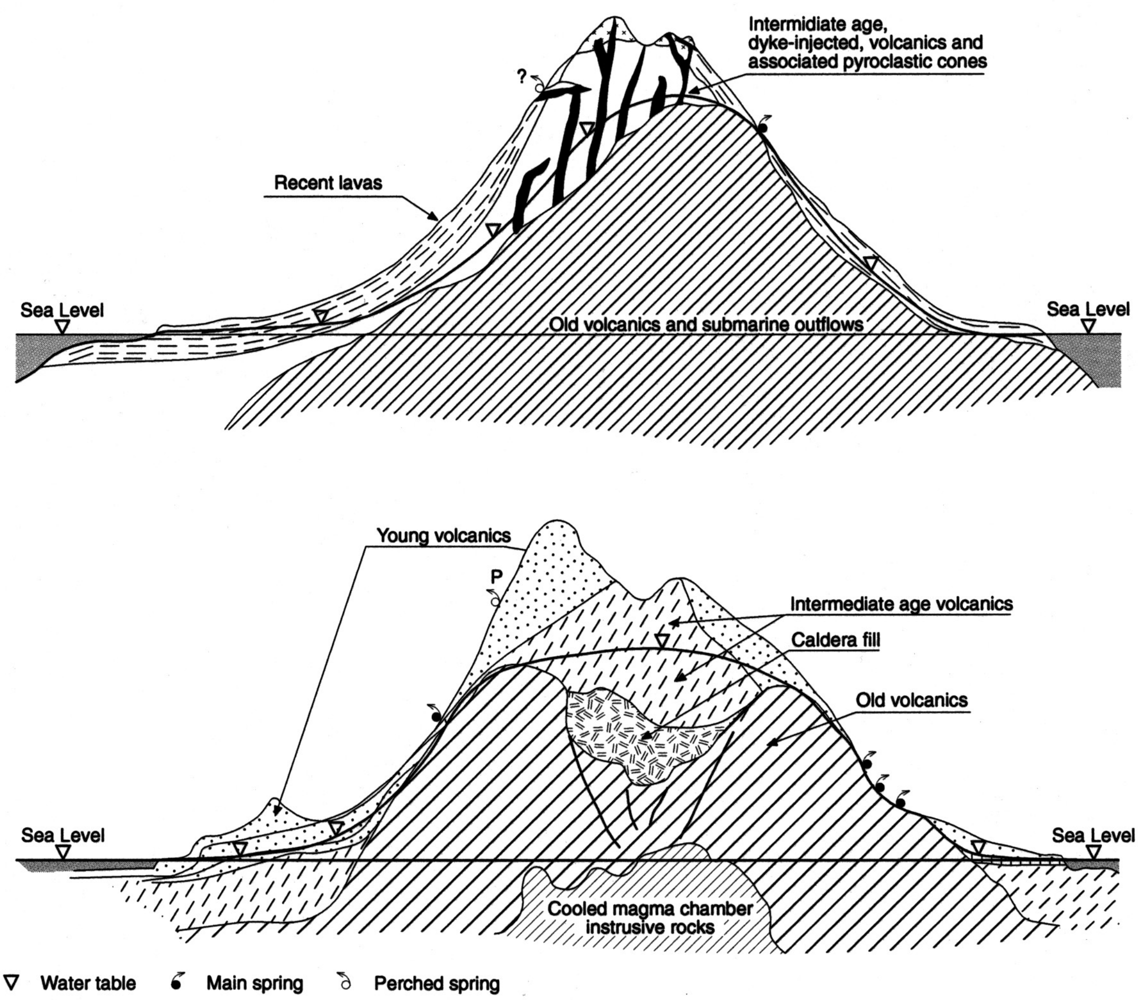
Large lava fields may confine older aquifers e.g. in the Paraná basin and the Cretaceous trap basalts (Serra Geral, Alto Paraná and Arapei formations). The latter are up to several hundred metres thick; there are poor aquifers in the 50 to 100 m upper fractured layer below the clayey soils and the rest is an aquitard or even an aquiclude that confines the underlying Jurassic Guaraní sandstone aquifer, a main regional aquifer that is exploited by deep wells, locally free flowing and producing warm water, that extends over Brasil, Uruguay, Argentina and Paraguay.

The groundwater flow description becomes more uncertain when approaching the eruptive centres as well as in volcanic islands. In these islands young volcanics may form a relatively permeable cover on top of a poorly permeable, densely dyke-intruded core in which groundwater flows slowly. The water table may attain high elevations, up to the gradational transition to the cover of young volcanics. In some of the Canaries the water table may be more than 1,000 m above mean sea level, with slopes greater than 0.1, which indicates the existence of a poorly permeable core of older, more altered volcanics. Permanent springs appear in areas where the young volcanic cover thins out and is no longer able to transmit all the groundwater flowing inside it (Fig. 16.7).

The existence of such low permeability core of old, even submarine volcanics, or physically and chemically highly metamorphosed ones is not always clear since it may not outcrop or it is difficult to recognise due to slope debris and dense vegetation cover, as in humid tropical volcanic islands. However, geophysical surveys and thermal and hydrological characteristics may show its existence, as in La Fournaise, southern Réunion island (Violette et al., 1997; Robineau et al., 1997; Courteaud, 1996), and also in Hierro, Canary Islands, after the recent drilling and excavation of a deep, long water gallery.

When erosion cuts through the core, numerous very small springs may form, many of them just sustaining vegetation. The landscape is very different from areas covered by young

Figure 16.7 Schematic cross section through volcanic islands of the high type



The figure shows the formation of a water-saturated high dome due to the presence of a poorly permeable core. This corresponds to the situation of Gran Canaria and Gomera (Canary Islands) and Madeira, and to some extent can be applied to Tenerife and La Palma (Canary Islands). The core may not exist above sea level in recent islands and then this mounding is not produced.

volcanics in which the vegetation is reduced to what the local soil cover and rainfall permits, except around the sparse, relatively large springs. Some of these springs form permanent streams, although water may be partly hidden in the torrential bed deposits. When the young cover is thick enough and when there are no large permanent springs the entire recharge is discharged into the valley bottom or into the sea.

Groundwater flow may be modified by major features, such as the sedimentary formations (marls and limestones) sustaining Mount Etna, Sicily (Italy); the terrestrial and marine deposits ('cap rock') confining permeable basalts in some coastal areas in Oahu, Hawaii, or acting as a main drain in eastern Gran Canaria; the river terraces and lake deposits in Olot, Catalonia, Spain; the sandstone formations below the trap basalts in the Paraná basin, etc. Inside volcanic formations densely dyke-intruded zones may act as anisotropic ridges, with open fissures parallel to the dominant dyke orientation and very low regional permeability at right angles to this. Major landslides may produce fine-grained breccias at the slipping surface, which may be a major internal discontinuity, as in the case of the 'mortalón' in Tenerife (Canary Islands).

Major discharges (springs, 'nacientes') tend to concentrate in breccias and enlarged joints. Poorly consolidated local features surrounded by more competent formations tend to result in caves being formed in the cliffs, since loose material is easily removed, producing what have sometimes been erroneously called karst-like features. These features do not behave like karst, nor are they produced the same way. The same holds for springs outflowing from tree log moulds and volcanic caves in high rainfall, high water table areas. These are occasional features which allow the concentration of otherwise diffuse groundwater discharge, but lack the organisation of karst. The same can be said of the deep vertical holes sometimes found near young volcanoes. They are vertical moulds made from thermally hardened rock or vertical lava pipes after the magma inside was suddenly expelled by a burst of gas. They are in no way like a karst sinkhole.

Groundwater flow in volcanic formations has been successfully described at regional scale by means of anisotropic, continuous media equations. Adjusted parameters are not true storage and permeability values (or transmissivity in two-dimensional horizontal situations) but do not differ greatly from values obtained by means of pumping tests. Double continuum models combining a high permeability, low storativity continuum with a high storativity, low permeability continuum linked by means of a block fissure water exchange function seem a better approach to groundwater flow simulation. Up to now these models have seldom been applied to volcanic formations except for studying poorly permeable formations considered as possible repositories for nuclear waste. Many of them correspond to Yucca Mountain, in western United States of America (Hinds et al., 1999; Sonnenthal and Bodvarsson, 1999), where advanced tests method have been devised (Huang et al., 1999).

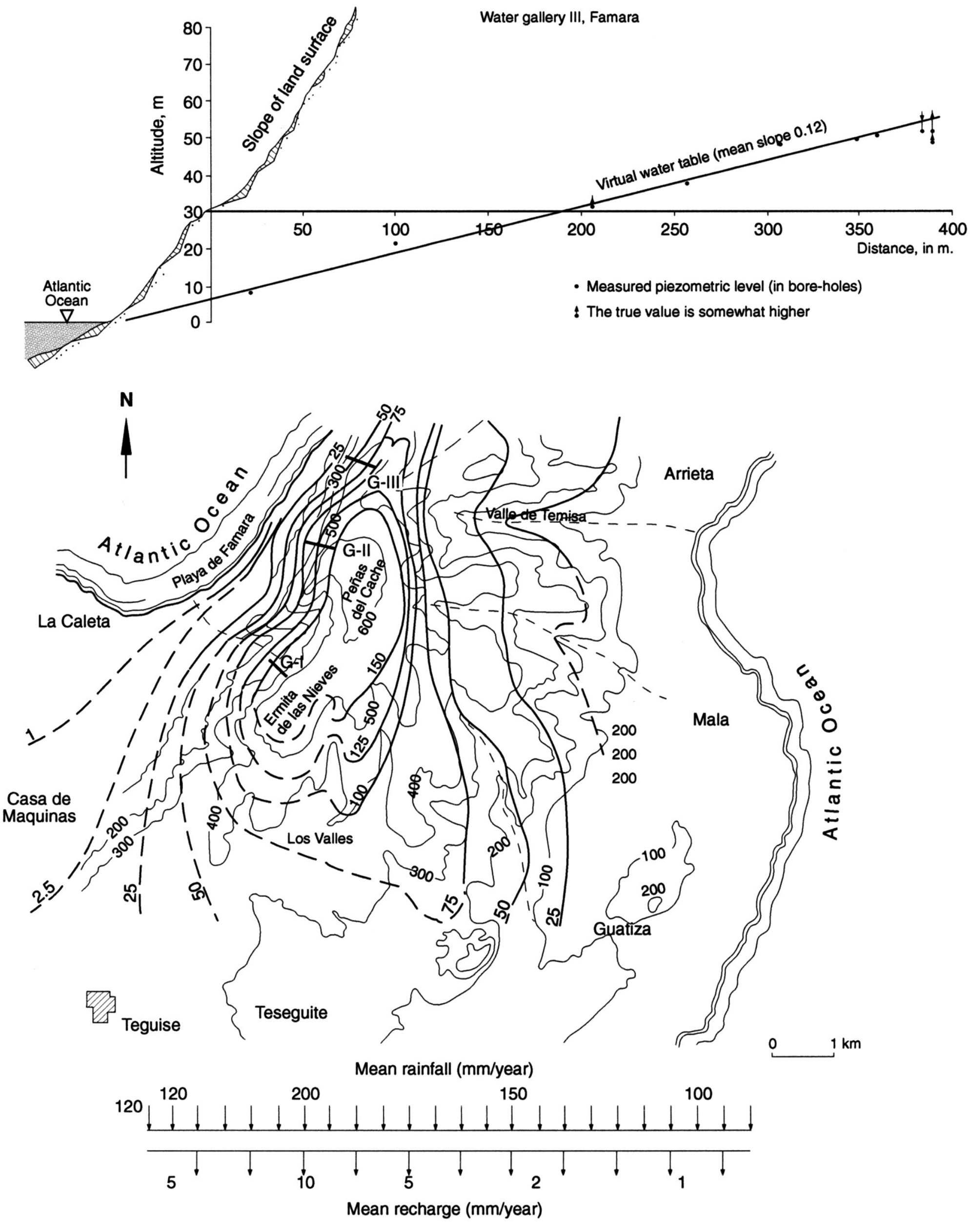
The validity of Darcy's law to describe macroscopic flow in volcanic formations is a recurrent topic. Detractors often rely on speculation and poorly interpreted local behaviours. Experience in the Canaries and elsewhere show that macroscopic behaviour can be linearly described by groundwater heads and permeability-like coefficients. Measurements obtained from several piezometric boreholes in water galleries in Lanzarote (Fig. 16.8) and Gran Canaria show that the water head line along the gallery is macroscopically continuous and has a constant slope, in spite of the different lava flows, pyroclastic layers, dykes and red beds appearing along the test tunnel.

Pumping tests to determine hydraulic properties are generally interpreted in the same way as the tests carried out in granular materials. There is little experience of tests carried out in well bounded conditions. In such cases the common models (Theis, Hantush) are able to reasonably reproduce drawdown-distance-time behaviour, yielding transmissivity and storage coefficient values which are averaged properties of a relatively large volume of rock. It seems that the consideration of anisotropic effects (Papadopulos-Cooper model) improves the fit, but multi-piezometer tests are rare. In other cases the Neuman model of delayed response of the water table seems to improve the fit to real data. Often only drawdown at the pumping well is measured. Where the wells are large diameter ones – sometimes with enlargements, galleries and horizontal drains – that partially penetrate a variable formation aquifer, well storage may dominate the test. In such circumstances it is difficult to obtain aquifer parameters, and the good fits sometimes obtained may owe more to the fitting effort than to the interpretation model.

16.5 Hydrogeochemistry and mass transport in volcanic formations

Volcanic formations do not usually contain significant quantities of water-soluble material, such as alkaline halides or sulphates, because these compounds are easily altered (hydrolysed) if

Figure 16.8 Experimental water gallery in the Famara massif, Northern Lanzarote, Canary Islands, Spain



The massif consists of more than 700 m of late Tertiary sheet lava flows, with some extensive red layers, buried cinder cones and a moderately dense network of dykes, oriented approximately N–S. Several vertical and horizontal bores were drilled along the gallery, and fitted with pressure meters in the section below the original water table. The heads that were recorded correspond to the points in the upper figure. These points lie on a nearly perfect straight line. The small deviation can be explained by vertical head gradients created by the water flowing into the gallery. Although the gallery is relatively short, it penetrates several dykes and cinder cones and affects several lava flows and red layers.

The lower figure shows the Famara massif water table contours in 1970–72 (Custodio, 1978, 1989). They show the mound-forming effect inside the poorly permeable volcanics of the massif, even for the small recharge shown below. The water table flattens in the more recent, even sub-historical lava flows of the eastern side. The massif has some very small, perched springs (seeps), supported by the red layers when these are locally thick and continuous. This water has the same chemical characteristics as water from the saturated part of the gallery and shows similar aridity effects on its isotope composition.

water is pH buffered. Dissolved CO_2 plays this buffering role. Low temperature alteration (weathering) frees alkaline and earth-alkaline metals in the rock as soluble ions, together with some silica, and bicarbonate ions. The rock minerals are transformed into clay minerals which include oxides of other metals such as iron. Most iron in fresh volcanic rocks is generally in the reduced state. In oxygen-free acidic water it dissolves as Fe^{++}, but usually it is oxidised by oxygen dissolved in the water and precipitated as Fe_3 oxides and oxyhydroxides. Aluminium is practically immobile and remains in place as part of clay minerals, or as aluminium hydroxide (gibbsite) if there is intense leaching.

Most of the Cl^- and SO_4^{2-} in groundwater comes from rainfall and dry atmospheric deposition (fallout). But some volcanic rocks, especially the submarine ones, may contain enough sulphide to contribute significant quantities of SO_4^{2-} upon oxidation.

In some cases there is a lithological origin of Cl^- and SO_4^{2-}, as an Fuerteventura island (Canary island), probably related with submarine volcanic formations (Herrera and Custodio, 2000) and sometimes the joint consideration of Cl^- and Br^- is an useful tool to know the origin of salinity and the possible contribution of volcanic activity (Davis et al., 1998; Custodio and Herrera, 2000).

Though their overlying soils contain oxidisable organic matter, volcanic formations do not contain significant quantities of oxygen-consuming substances other than reduced iron and manganese. The groundwater is rarely under strong reducing conditions and most often it is under oxidant conditions. Buried organic matter from soils and vegetation covered by lava flows may create strong reducing conditions, even producing methane, but they are rare and local.

The cationic composition of groundwater is related to the type of volcanic rock (Fig. 16.9). Considering the ratio of alkaline to earth-alkaline ions, the lower values correspond to basic (low silica) volcanic rocks such as basalts, and the high values to acidic (high silica) rocks such as rhyolites. The results are more realistic if the contribution of atmospheric cations is first subtracted. This can be done by using the Cl^- content as a guide.

Groundwater salinity in volcanic rocks is the result of atmospheric deposition of salt , climatic conditions and rock weathering. There are no major differences compared with other easily weatherable rocks under similar circumstances (Custodio et al., 1997). Since chloride is not significantly contributed by the rock in areas with no significant human influence (agriculture, urbanisation, etc.) the simplified long-term balance (without considering storage terms) is:

$$P \cdot C_P = R \cdot C_R + S \cdot C_S$$

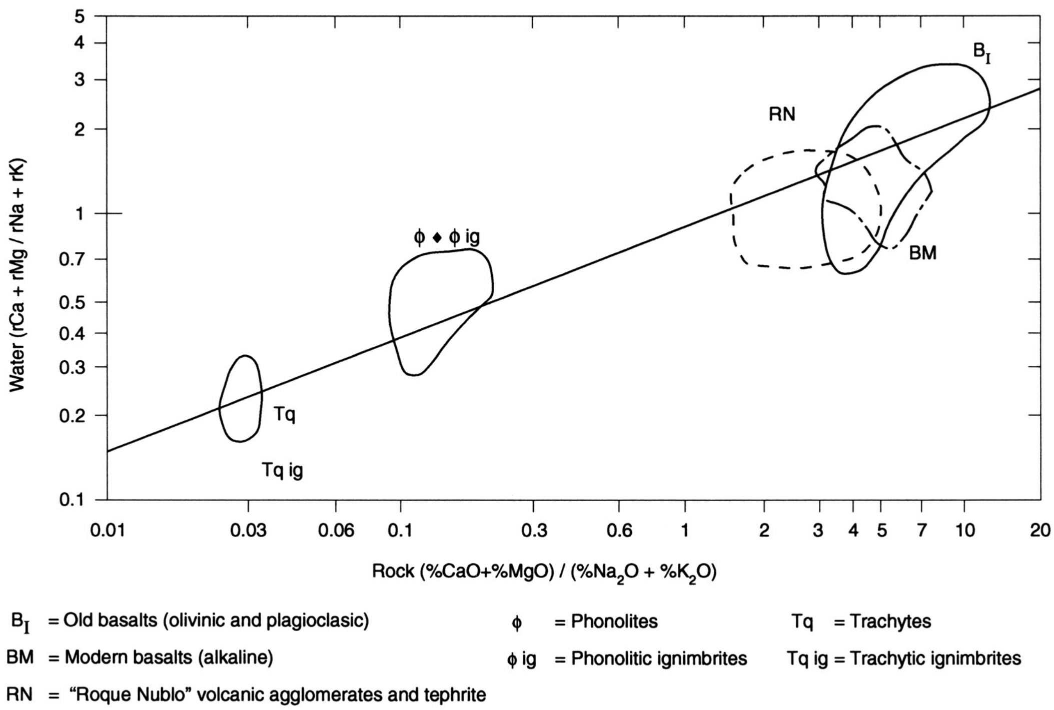
in which:

C	=	mean concentration of the conservative solute of interest in the water balance component shown by the subscript	*[M L⁻³]*
P	=	mean local rainfall = ET + R + S	*[L T⁻¹]*
ET	=	mean actual evapotranspiration	
R	=	mean recharge to the aquifer	
S	=	surface runoff	

C_P includes the chloride contributed by atmospheric fallout and may be obtained from an open, rainfall collector protected against evaporation. If D is total chloride contribution per unit land surface area and unit time *[M L^{-2} T^{-1}]*, C_P = D/P. Typically D may vary from 2–5 g m^{-2} a^{-1} near the coast to 0.2–1 g m^{-2} a^{-1} in continental situations. Then, since no other sources are present and chloride does not precipitate, except in very extreme climatic conditions:

$$C_R = (P/R)\, C_P - (S/R)\, C_S$$

Figure 16.9 Relationship between rock and water composition



The figure shows the relationship between mobile cation rock composition and water composition and major monovalent and divalent cation composition of groundwater from different volcanic rock types (lower figure), as deduced from studies in Gran Canaria (CEDEX, 1987; Custodio, 1989). The more basic (closer to basalt) the rock is, the higher is the ratio of earth-alkaline to alkaline ions. The anion content is to some extent influenced by arid climatic conditions and by transformation of volcanic CO_2 into bicarbonate ion. The r represents concentration in meq/L.

In permeable soils the second term of the right-hand side of the equation is small, and then:

$$C_R \approx (P/R)\, C_P$$

(P/R) is the climatic-pedological concentration factor, which means that when converted into recharge rainfall salinity increases moderately in humid climates, but greatly in dry climates. Rainfall salinity depends on distance from the sea along the dominant wind direction (Custodio et al., 1997). The same is true for Br^- and also for SO_4^{2-} if the contribution by rock sulphide oxidation is negligible and there is no net sulphate precipitation in the upper soil. Near populated areas and far from the coast SO_4^{2-} increases relative to Cl^- due to anthropogenic sources and recycling.

For other ions the balance has to be changed into:

$$P \cdot C_P + W = R \cdot C_R + S \cdot C_S + F$$

in which:

W = contribution from weathering *[M L^{-2} T^{-1}]*

F = separation by net precipitation, ion exchange and/or incorporation into weathering products *[M L^{-2} T^{-1}]*

W is limited by oxygen availability (which is transported with recharge water and slowly diffused from the atmosphere) and especially by carbon dioxide supply. In very simplified terms, each molecule of CO_2 contributed is transformed into $HCO_3 + H^+$ and each H^+ is exchanged (in mass balance terms) by an atom equivalent quantity of cations. The actual behaviour is somewhat more complex, depending on the final pH and redox reactions.

CO_2 availability in the soil is limited by the organic matter oxidation rate. It is low in arid and cold climates and relatively high in warm, humid conditions with abundant vegetation. In this case weathering (rock factor) dominates groundwater chemistry. The resulting water contains the rock dominant cation and bicarbonate, is rich in dissolved silica and is relatively diluted due to the high recharge. In dry climates the climatic-pedological factor may remain dominant and in areas under marine influence recharge water generally is of the sodium chloride type, but local circumstances play an important role.

As in other rock formations, the chemical composition of recharge water is related to the soil rock which is being weathered, with little change afterwards, except for groundwater paths with very long residence time, but in volcanic formations volcanic CO_2 has to be taken into account. The CO_2 is slowly released from the cooling magma chambers and diffuses through the volcanic rock mass, even if this is water saturated, towards the external atmosphere. This includes CO_2 in the closed pores of volcanic rocks that is released after weathering or increasing fracturing due to overburden pressure and tectonics.

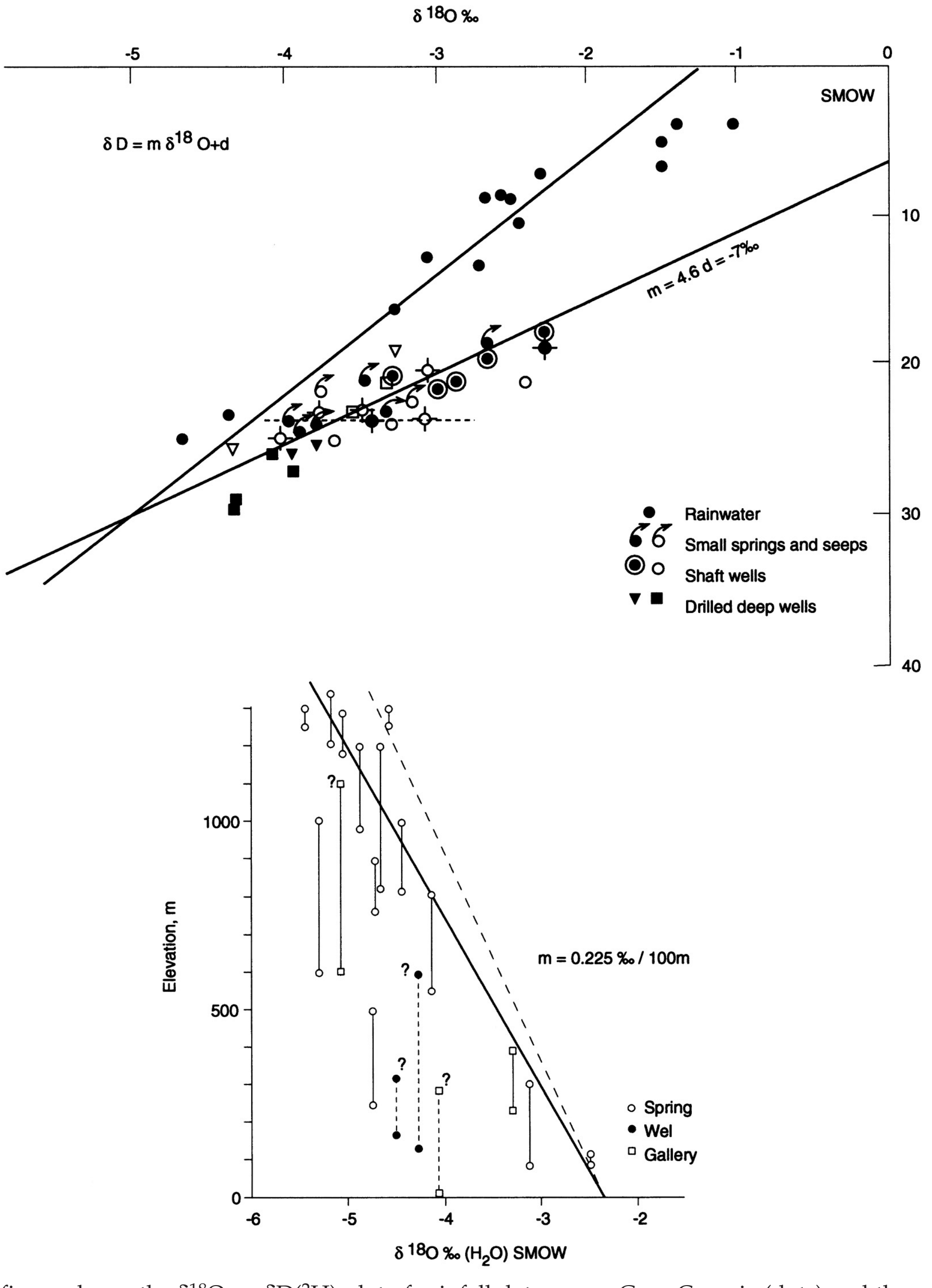
CO_2 release from volcanic effluent is intense at the initial cooling stages and decreases with time. It may produce bicarbonate-rich water if enough contact time with the rock is allowed. In the areas affected by outgasing, chloride-poor but very bicarbonate-rich (up to 2 g/l) water may form, especially in the volcanic cores where the poor permeability ensures that there is enough contact time. Acidic and intermediate volcanics and alkaline basalts may thus contain water rich in sodium bicarbonate. Less alkaline water, rich in magnesium and calcium bicarbonate and generally with a high pH, is found in plagioclase basalts. Calcium carbonate may precipitate, filling voids and cracks.

Released CO_2 from active or recent volcanism, coming from degassing magma, generally diffuses slowly through the rocks and groundwater. In some case it may accumulate in deep lakes (Delmelle and Bernard, 1994) and even create episodically sudden releases. It may also accumulate in confined aquifers that when tapped by a borehole may burst out during some time. A spectacular case has attracted the attention of newspapers during the summer of year 2000 in the now extinct Quaternary volcanic area of Campo de Calatrava (Ciudad Real, Central Spain).

Chemical denudation rate depend on recharge by rainfall and dissolution capacity, mostly due to CO_2 (Louvat and Alègre, 1997). Actual conditions vary from very slow in arid areas (rainfall may be less than 0.1 m/a, as in some parts of the Canary and Cape Verde islands), up to more than 10 m/a in some tropical, humid areas, as in some places in Central America, Réunion, Tahiti and Hawaii.

The stable isotope composition of groundwater is not affected by volcanic rocks except under hot rock conditions or intense gas flow. In a hot environment the ^{18}O in the water is readily exchanged with that in the rock, and the water becomes heavier (less negative $\delta^{18}O$), but the δ^2H (deuterium) value does not change. The exchange is less marked than with carbonates. If enough CO_2 bubbles through the water the exchange makes the water lighter in $\delta^{18}O$, but the δ^2H is unchanged if there is no evaporation. Evaporation makes the water heavier, both in $\delta^{18}O$ and δ^2H, with a $\delta^{18}O-\delta^2H$ slope less than 8, depending on temperature. The use of ^{18}O and ^2H to study and characterise groundwater flow does not present special difficulties. Classical interpretation rules and models can be applied in most situations. They are very useful tools in areas with large topographical variations, as in many volcanic islands and formations (Fig. 16.10). The same can be said for radioactive ^3H (tritium), but in this case its transient

Figure 16.10 Water isotopes applied to volcanic areas



The figure shows the $\delta^{18}O$ *vs.* $\delta D(^2H)$ plot of rainfall data across Gran Canaria (dots) and the results of sampling small seeps, shaft wells and deep bored wells in the southeastern part of the island (Amurga phonolitic massif), which is an arid area. There is some recharge from occasional storm water which follows the World Mean Meteoric line (m = 8; d = +10 ‰), but part of it evaporates from the bare rock and the soil surface due to the sparse vegetation cover. The lower figure shows the $\delta^{18}O$ *vs.* altitude relationship for Gomera (Canary islands). Most water samples come from springs, from near sea level up to 1,300 m. The isotopic slope is about 0.225 ‰/100 m. The lower point of each plot is the sampling altitude and the upper one the mean altitude of the assumed recharge area. Some springs and galleries receive water from a higher altitude, which means significant lateral transfer of water through the more permeable upper volcanic formations resting on older volcanics.

concentration introduces additional mass transport problems. Waters of different origin and variable residence time can be identified (Falkland and Custodio, 1991; Pennisi et al., 2000).

Since volcanic rocks are devoid of carbonate minerals – except for some secondary deposits in veins and pores of old volcanics – the dissolved inorganic carbon (DIC) comes exclusively from CO_2. In areas not affected by volcanic carbon dioxide, DIC isotopic composition is derived directly from CO_2 released by decaying vegetation. Thus, if vegetation is characterised by $\delta^{13}C_v$, groundwater has a composition between $\delta^{13}C_v$ (closed system to CO_2) and $\delta^{13}C_v + \varepsilon$ (open system to CO_2 when dissolved inorganic carbon is dominantly HCO_3^-; pH between 6.8 and 7.8). ε is the temperature-dependent fractionation factor between CO_2 gas and dissolved HCO_3^-. Thus, for Calving-cycle plants and temperatures around 18°C, the dissolved inorganic carbon $\delta^{13}C$ varies between –25‰ and –17‰ relative to the PDB standard. Carbonate precipitation in arid areas may change these results.

The same is true when the dominant DIC source is CO_2 from magma degassing, but in that case $\delta^{13}C_v$ is close to −8‰. For a system that does not lose CO_2 (closed system to CO_2) $\delta^{13}C_D \cong -8‰$ but when groundwater is almost stagnant and CO_2 diffuses through it toward outer regions the behaviour of the CO_2 is close to an open system and $\delta^{13}C_D \cong 0‰$ at about 18°C, when HCO_3^- dominates.

A series of intermediate situations between the two origins of CO_2 is possible. This makes it difficult to interpret radioactive ^{14}C content to obtain groundwater age. The carbon hydrogeochemical behaviour in the volcanic system under study needs to be considered carefully. If volcanic rock circumstances are not duly taken into account, classical correction procedures, mostly devised for carbonate rich formations, may fail to give reasonable answers. Water devoid of significant quantities of radiocarbon are not necessarily old but may merely reflect a dominant volcanic CO_2 contribution. This pose additional problems to interpretation (Rose and Davisson, 1996). Correlation between $\delta^{13}C$ and total dissolved carbon and between $\delta^{13}C$ and $\delta^{14}C$ content may help in obtaining correct interpretations.

From the few applications so far it is clear that mass transport of solutes through volcanic formations cannot be accurately described by the common simple continuum mass transport equation. A double continuum approach seems to improve results, but there is little experience of using this method. The solute is diffusely exchanged between the porous blocks and the low storativity fissures. Chemical or temperature changes are dispersed and delayed, with sharp fronts and long tails. Thus, conservative solutes such as tritium are affected and groundwater turnover time calculation by means of environmental tritium is more complex than in granular aquifers under similar geometrical circumstances.

Case studies dealing with the transport of contaminants in volcanic rocks are scarce, especially for reactive contaminants and those which may decay. It seems that absorption isotherms and ion-exchange rules can be applied to some extent, but actual properties are poorly known. Studies on pesticide transport in Oahu island, Hawaii (United States of America) show that transport in soils can be described conventionally (Oki and Giambelluca, 1989), when there are thick soils (Loague, 1994). In other cases soils are thin or absent. In this case little knowledge is available.

16.6 Groundwater quality issues in volcanic formations

Groundwater quality issues in volcanic rocks are not essentially different from other rocks, but there are some aspects that deserve consideration:

- the easy weathering of the rock means that before being released to the atmosphere soil CO_2 is used to increase total dissolved solids in a larger proportion than in other rocks under similar conditions;

- the contribution of volcanic CO_2 in areas with relatively recent volcanism or magma emplacement episodes may produce water rich in dissolved solids, rich in sodium and/or magnesium bicarbonate, sometimes sparkling at atmospheric pressure, and sometimes with rather high dissolved Fe^{++}. These waters are unsuitable for domestic supply, may be highly corrosive to metal and fibro-cement pipes, or highly incrusting, depending on circumstances, and most unsuitable for irrigation.
- it is not rare for fluor ions, F, to attain concentrations well above the upper limit permitted for drinking water, especially in acidic, soft water.

Occasionally, problems are caused by high concentrations in boron or in some heavy semi-metals, such as vanadium and arsenic but most problems of related to the effects of late stages of volcanism on some springs. Problems of high V and As concentrations in pore waters in the semi-arid plains (Pampa) of Argentina are linked to glassy volcanic dust from the Andean volcanoes that becomes incorporated into the aeolian-fluvial deposits; a process still active. In Mexico, arsenic is a problem in some rhyolite aquifers. It is well known that large metallic mineralizations are associated to important ancient volcanic events, mainly in subduction areas. This is well characterised in the Pyrite belt and Ossa-Morena areas of the Iberian Peninsula. This is the subject of classical research and mapping studies. The erosion, transport and subsequent sedimentation, adsorption or precipitation of some components may be the origin of noxious concentration of some elements in groundwater, although in many cases the geochemical conditions are complex and poorly known.

16.7 Groundwater exploitation in volcanic formations

Groundwater exploitation from volcanic formations relies on the same methods commonly used in other formations, especially in hard rocks, although there are specific issues.

In lava formations far from the eruption centres, springs outflowing at valley bottoms are tapped. Often these valleys correspond to tectonic features in which not only erosion acts preferentially but permeability is also locally greater. The most frequent form of groundwater abstraction is by means of wells. Large-diameter dug wells, 2 to 8 m in diameter and a few tens of metres deep are common in the Deccan traps. The large storage capacity of water inside the well allows for relatively high pumping rates during short periods, though the yield of the aquifer is often low.

Currently, drilled wells are common. They are 0.1 to 0.6 m in diameter and up to several hundred metres deep. Drilling with a pneumatic down-the-hole hammer rig is currently the most common drilling method, although not without problems of soft sections being unstable or caving. Cable tool percussion is also used, but penetrating thick massive and abrasive lava flows is a slow and costly process. Almost any drilling procedure for hard rock has been used with variable success, depending on local circumstances and driller's experience. Wells are cased when soft sections cave in, or to protect the pump. Torch or machine slotted tube is generally sufficient and preferred to more elaborate screen sections.

The groundwater yielding formations are mainly the breccia and scoria between lava flows, especially the less altered and thicker flows. In some cases it is possible to assign water-yielding properties to identifiable individual lava flows. Data for the Deccan traps show that permeability decreases with depth.

In the Paraná basin basalt outcrops the situation is not so clear, but water-yielding properties decrease with depth, too. In this case, large fissures are also a main source of groundwater, but they are not easily intersected by wells, since these fissures tend to be vertical. There is a greater probability of encountering fissures close to tectonic features, which in some cases show up as lineaments in aerial pictures and as deep valleys on the ground.

In volcanics at or near eruption areas matters are more complex due to the variable internal structure of the formations. They may present a wide range of groundwater-yielding properties. As these areas are generally of pronounced relief, relatively high altitude springs may appear, as in many volcanic islands receiving relatively good rainfall recharge. The direct tapping of the streams that originate from them allow water to be transported by gravity to urban areas or to irrigated fields at lower elevations, and even to generate some electricity if there is enough flow and head difference, as in the SE of Réunion island. Water that is not used is lost downstream or to the sea. In water-scarce areas, reservoirs have been built to store excess water for periods of high water demands. Surface water storage reservoirs may be built by damming the valleys when these are in poorly permeable volcanic material, as in Gran Canaria; otherwise, artificial impermeable linings must be constructed.

In some situations spring tapping is 'improved' in order to collect much of the otherwise dispersed outflow in one point. This is done by digging horizontal water galleries at or below the spring outflow point. After excavation the water flow often increases dramatically for some time, due to the drainage of the local groundwater storage. When this is depleted, the final steady flow may be considered to be too small, requiring the gallery to be extended. This process has produced in Tenerife (Canary Islands) hundreds of water galleries up to 6 km long and up to 1 km below ground level at the front. Some of them have reached the high-temperature, sodium bicarbonate rich waters of the island core. Water galleries are efficient groundwater abstraction works if they penetrate perpendicular to dykes and major vertical fissures, but the final steady flow depends on the fraction of the recharge they intercept. This may be quite small, even for a long gallery, as most of the yield is produced at the head section when much of gallery's length may become unsaturated zone due to depletion and the effect of other water galleries at the same or at a lower altitude. Bores drilled horizontally from the gallery sides help to increase the yield temporarily, in the same way as in large diameter wells. Transient periods may last from weeks to many years. As in the case of springs, water that is not used is lost downstream. This is of special concern if this water is one time storage water. Locally bulk-heads have been installed deep inside the galleries. A few examples can be found in Mount Etna (Sicily), in Tenerife and in Hierro (Canary Islands). Bulkheads are difficult to construct, due to the high pressure that may develop behind them. A massive dyke may be a good place to anchor them. Water may penetrate through most of the active length of the gallery but most often it appears in short sections separated by long, poorly yielding and 'dry' sections. Some sections have to be lined to avoid caving in.

Wells are the most commonly used groundwater abstraction works, although water has to be pumped out, expending energy and using costly machinery. This was a major problem in the past when steam or oil driven motors were needed to move piston pumps to push up the water. Currently, submersible pumps and electrical energy have simplified many of the old problems. In remote areas electrical energy is locally generated by means of oil-driven motors. The appropriate solutions depend on the area (Kulkarni et al., 1997).

Drilled wells using the methods commented on before are now common, but groundwater-yielding sections are often separated by long sections of non-yielding rock. This means that deep wells have to be drilled, with the risk of bypassing good quality, short turnover groundwater below the water table and instead tapping more mineralised, deep groundwater, or mixing the two at the pump inlet. Drilling deep below the water table is a problem for air-driven bottom hammers, since air pressure has to overcome the pressure created by the water column in the bore. High pressure air compressors or supplementary airlift is needed. Caving-in of loose material is a major problem, producing serious failures. Casing is a method to avoid problems of caving-in but makes drilling slower and more expensive. Vertically drilled wells are not efficient groundwater abstraction works in areas where vertical fissures are important water-yielding features that help in draining large volumes of rock.

In the Canary Islands, and especially in Gran Canaria, the solution to accommodate the old bulky pumping machinery was to construct large diameter wells (about 3 m), excavated like a mine shaft with lining and reinforcement of soft sections. At the same time the large diameter of the well allowed horizontal water galleries to be excavated from the well bottom in order to intersect vertical groundwater-yielding internal features, thus greatly increasing the well yield. More recently the dangerous, expensive and risky gallery excavation method has been replaced by horizontally drilled bores ('catas') from the well bottom, using especially adapted, electrically powered drilling machinery. The 'catas' are often 50 to 200 m long, but some are longer than 500 m to increase the yield of the expensive excavated wells, which are often up to 500 m deep. The drilling machine has to be placed below the static water table and the well shaft has to be continuously dewatered by pumping.

Groundwater production from the well drains water stored in the rock and the discharge decreases as this storage is depleted. When well yield becomes too small to meet needs, the well has to be deepened and new 'catas' drilled. This is an expensive process that need frequent or continuous work inside the well. Moreover, more energy is needed to pump out the water. Whether redrilling is advisable depends on the price of water in the market, in combination with water quantity issues and administrative regulations from the Water Authority. Existing wells are maintained and deepened, but new wells are rarely constructed, due to the large investment required to reach the deep saturated zone. Currently these large diameter shafts are extended down by means of small diameter, mechanical drilling, substituting lateral drainage by deep penetration, but this currently cheaper method is hydraulically less efficient in draining large rock volumes.

In strongly sloping ground, side galleries are constructed to take out the water at a lower elevation, thus saving energy. Something similar is done in long water galleries, from which outflowing water has to be pumped prior to distribution. By using a closed pipe, instead of an open canal to tap the water at the gallery bottom where most of it is produced, saves part of the hydraulic head above the gallery mouth due to the slope of the gallery (about 0.02).

16.8 Groundwater balance in volcanic formations

The groundwater balance in volcanic formations follows the same principles as in any other aquifer system (Custodio et al., 1997), and will not be described here, except for some specific comments.

Groundwater recharge by infiltrating rainfall and snowmelt is a major term which is difficult to calculate with accuracy. This is not new. In many volcanic areas, soil is very thin and permeable, and vegetation is shallow rooted. Consequently, in wet climates a large part of rainfall becomes recharge. Values greater than 50% of local rainfall are possible. But in arid climates, even in sparsely vegetated areas, most of the rainfall returns to the atmosphere, partly by direct soil evaporation. This is reflected in the isotopic composition of the water. Studies carried out in southeast Gran Canaria (Canary Islands), in a dry area receiving between 100 and 200 mm of rainfall annually, showed a mean recharge of 2 to 10 mm/year with a salinity of up to 5 g/l total dissolved solids due to the rainfall taking up airborne marine salts (Custodio, 1992).Recharge is zero in most of the years, except the wettest ones. As a consequence, a thick unsaturated zone and the thick saturated zone underneath results in old radiocarbon ages for pumped water, even though this water comes from a water table aquifer receiving some recharge.

Many volcanic areas may develop fertile soils on which a dense vegetation cover can develop. This cover often uses rainfall water very efficiently in moderate precipitation areas, thus dramatically reducing recharge. Some of this vegetation may satisfy part of its water needs from capturing air moisture in foggy areas.

Recent volcanic formations, still devoid of significant vegetation, highly porous and with a very irregular land surface (badlands or 'malpais') seem prone to transform most of the rainfall into recharge. This is true for high rainfall areas (in which weathering proceeds fast) but not necessarily for arid areas. Air may flow through the upper few metres of the ground due to wind effects and temperature gradients, taking away most of the rain-water held in the large specific surface lapilli, scoria and breccia.

Although often poorly permeable, volcanic formations may conserve connected porosity, some permeability and open fissures down to large depths. This has been shown in studies for possible nuclear waste repositories in North America and also in the long water galleries of Tenerife (Canary Islands). In the latter there is some groundwater permeability in the first thousand metres and probably deeper. This has to be taken into account when putting boundaries to a groundwater balance domain.

Low permeability, the possibility of preferential drainage through fissures intersected by groundwater abstraction works and regionalised systems due to the large active thickness means that any hydraulic disturbance is followed by a long transient period which can be measured in terms of

$$L^2 S/T \equiv L^2 S_v/k$$

in which:

L	=	linear dimension of the system	*[L]*
S	=	storage coefficient	*[dimensionless]*
S_v	=	specific storage coefficient	*[L^{-1}]*
T	=	transmissivity	*[L^2T^{-1}]*
k	=	permeability	*[LT^{-1}]*

In 'core' situations, characteristic times in the order of one year are common values for a deep well or gallery, due to the effect of the low, local S_v value (elastic behaviour), but for the whole formation they may attain hundreds of years for a long-term drainable porosity of 0.02 to 0.06. This mean that storage terms may play an important role in the groundwater balance of the system. This is something often ignored, especially in volcanic islands, where the groundwater outflow to the sea is not measurable and mostly unnoticed. Thus, even when groundwater abstraction is less than recharge, continuous large drawdowns may be observed -larger around individual wells and water galleries- since the steady state can only be attained by reducing the outflow to the sea or a main river valley. When groundwater abstraction is far from the coast or the valley bottom, this is only possible with a large total drawdown.

Groundwater management issues are complex ones and they depend on local circumstances and rules. When water is scarce due to aridity and/or high agricultural demand, prices are high. These rules are set, although they do not get necessarily an efficient use of water and a rational development of groundwater (Cabrera et al., 1997). It is not rare that private groundwater markets develop.

16.9 Groundwater monitoring in volcanic formations

Groundwater monitoring in volcanic formations covers three main topics:

- water levels,
- water physico-chemical and quality characteristics, and
- water abstraction.

In thick, poorly permeable formations, the three-dimensional nature of groundwater flow and chemical characteristics has to be taken into account. Water table slopes may be greater than 0.01 and sometimes exceed 0.1 (often under land surface control). In these cases, vertical flow components are important. This is clearly shown by the drawdown of the static water level in wells in highlands when they are deepened. Some authors describe groundwater flow in the massif as a series of numerous perched aquifers on top of a deep regional level corresponding to what has been called a 'basal aquifer'. This is generally incorrect and the very permeable deep layer cannot exist except if it lies on top of other formations such as fractured limestones. The conceptual model is a thick saturated poorly permeable formation with large vertical head gradients. This comes from the classic description of aquifers in Oahu, Hawaii, in which pervious recent basalts in the coastal plain have a gently sloping water table (the basal aquifer). This aquifer is at the foot of the much less permeable, densely dyke-intruded island core, with high altitude perennial springs. This is a common situation in many other volcanic islands (e.g. Tahiti, Réunion, Martinique), but it does not mean that the basal (coastal) aquifer groundwater levels continues inland below a thick series of perched aquifers.

Groundwater monitoring of pumping wells may give distorted results, due to local storage depletion. Moreover, in large-diameter wells the water levels measured when the pump is idle may show poor water level recovery, due to the large well capacity. Thus, the monitoring network has to be carefully designed and some data may need corrections. A deep open borehole shows a water head which is a composite value of the point heads along its length. Inside water galleries the locally influenced water table position can be seen when it is penetrated, but since there are large, poorly permeable sections and the ambient air is almost water vapour saturated, often the exact position cannot be located. Furthermore, the gallery is continuously yielding water and thus the intersection with the water table is locally distorted.

Groundwater level monitoring is best done by means of drilled observation boreholes, but this is expensive, especially when the unsaturated zone is thick. The ideal situation is when various separated piezometers (different bores or various isolated tubes in a large diameter borehole) are installed at different depths at a site, each with a short open section, grouting the rest. Since permeable features may be separated by long, poorly permeable sections, the placing of the open sections has to be decided after drilling and testing the borehole. Open sections in very poorly permeable rock may need weeks to show the true water level, and water head changes are greatly dampened and delayed. In such circumstances isolated water pressure transducers may be a solution for monitoring.

Groundwater quality monitoring can be carried out by periodically sampling wells and springs, generally once or twice a year. Consideration has to be given to what represents the water sample. When the well is not pumped, the water in it comes from the highest water head level or is dominated by water falling from possible perched aquifers. During pumping a mixture of the different yielding levels is produced, which varies with time and with pumping discharge, and in large-diameter wells it is influenced by water in storage. It is not easy to obtain good water samples which are representative of particular situations. Pumped water is a mixture of many different origins and residence times. Seasonal changes are sometimes observed, but they are more the result of variable mixtures due to seasonally affected groundwater levels than of a true seasonal change in water quality. To study agricultural pollution, the upper part of the water table aquifer must be sampled.

In coastal aquifers the existence of sea water and the mixing zone has to be considered. Salinity stratification can be studied by logging the electrical conductivity (EC) of the water along long open boreholes, but it is not rare for small vertical flows along the borehole to modify salinity distribution and the mixing zone thickness. Water temperature logging is very useful in identifying anomalous situations. Young volcanics far from the eruption centres and near the coast are easier to monitor, since mean permeability may be more homogeneous, and vertical flow negligible.

Monitoring groundwater abstraction seems an easy task in principle but it is very difficult in practice. Often, wells are operated discontinuously, pump discharge varies with water table elevation and the efficiency changes with time. Generally there is neither a flow meter nor an hour meter, or if they have been installed they do not work properly or at all due to lack of maintenance. Excess water from springs and galleries is seldom gauged; the same applies to water in streams. Field survey and interpretation of aerial and satellite pictures is a gook combination, but direct involvement of groundwater users through their representative associations is of paramount importance.

Recharge is seldom measured due to its complexity and the restricted value of local results. It is preferentially calculated from meteorological data and some soil observations and calibrated against water level changes or spring discharge if possible. This is a difficult task in volcanic formations with thin or discontinuous soil cover.

16.10 Geothermal effects in volcanic formations

Geothermal energy will not be dealt with here. Since volcanic formations may contain magma intrusions such as magma chambers, sills or buried necks, the vertical thermal gradient may be increased and convection of hot fluids may take place, affecting groundwater resources. The effect of CO_2 release has already been discussed.

The interaction of water with hot rock affects the equilibrium between water and silicate. This decreases the Na/K ratio and produces a $\delta^{18}O$ shift to more positive values. If there is some boiling the shift also effects deuterium. Condensed steam may produce small seeps that are characterised by lighter isotope composition and a larger deuterium excess.

Groundwater involved in geothermal convection is dominantly of meteoric origin, but on small islands seawater is involved as well (Henry et al., 1996).

16.11 References and additional reading

Araña, V; López-Ruiz, J., 1974. *Volcanismo: dinámica y petrología de sus productos.* Ediciones Istmo, Madrid, 481 pp.

Araña, V.; Ortíz, R., 1984. *Volcanología*. Ed. Rueda. Madrid, 510 pp.

Aubouin, J.; Brousse, R.; Lehman, J. P., 1981. *Précis de géologie: pétrologie*. Bordas, Paris. Translated, Tratado de geología: petrología. Ed. Omega, Barcelona, 602 pp.

Cabrera, M. C.; Custodio, E.; Roque, F., 1997. *Las aguas subterráneas y la planificación hidrológica en las Islas Canarias*. Asoc. Intern. Hidrogeólogos, Grupo Español, 266 pp.

Cabrera, M. C.; Nuñez, J. A.; Custodio, E., 1992. Contribución al conocimiento geológico del subsuelo de Telde (Gran Canaria, Islas Canarias). *Actas del III Congreso Geológico de España/ VII Congreso Latinoamericano de Geología, Salamanca,* Vol. 2, pp. 256–60.

Carracedo, J. C.; Day, S. J.; Guillou, H.; Rodriguez-Badiola, E.; Canas, J. S.; Pérez-Torrado, F. J., 1997. Origen y evolución del volcanismo de las Islas Canarias. In: *Ciencia y Cultura en Canarias.* Ed. Museo de la Ciencia y del Cosmos, Canarias, pp. 67–89.

CEDEX, 1987. *Simposio internacional sobre hidrología de terrenos volcánicos. Arrecife de Lanzarote, 1974.* CEDEX. Madrid, 1296 pp.

Courteaud, M., 1996. Étude des structures géologiques et hydrogéologiques du massif de La Fournaise par la méthode audiomagnetotellurique. Doc. Thesis. Université de la Réunion. Saint Denis, 205 pp.

CUSTODIO, E., 1978. *Geohidrogeología de terrenos e islas volcánicas*. Instituto de Hidrología/Centro de Estudios Hidrográficos, Pub. No. 128, Madrid, 303 pp.

——, 1989. Groundwater Characteristics and Problems in Volcanic Rock Terrain. In: *Isotope Techniques in the Study of the Hydrology of Fractured and Fissured Rocks.* STI/PUB/790, Intern. Atomic Energy Agency, Vienna, pp. 87–137.

——, 1992. Coastal Aquifer Salinisation as a Consequence of Aridity: The Case or Amurga Phonolitic Massif, Gran Canaria Island. *Study and Modelling of Salt Water Intrusion.* CIMNE-UPC, Barcelona, pp. 81–98.

——, 2000. Utilización de la relación Cl/Br como trazador hidrogeoquímico en hidrología subterránea. B*ol. Geolog. y Minero,* Vol. 11, Madrid, pp. 49–68.

CUSTODIO, E.; LLAMAS, M. R., 1983. *Hidrología subterránea (2 vol)*. Ed. Omega, Barcelona, 2350 pp.

CUSTODIO, E.; LLAMAS, M. R.; SAMPER, J., 1997. *La evaluación de la recarga a los acuíferos en la planificación hidrológica.* Asoc. Intern. Hidrogeólogos, Grupo Español-Inst. Tecnol. Geominero de España. Madrid. 455 pp.

DAVIS, S. N.; DE WIEST, R. J. M., 1966. *Hydrogeology*. Wiley, 463 pp.

DAVIS, S. N.; WHITTEMORE, D. O.; FABRYKA-MARTIN, J., 1998. Uses of Chloride/Bromide Ratios in Studies of Potable Water. *Ground Water,* Vol. 36, pp. 338–50.

DELMELLE, P.; BERNARD, A., 1994. Geochemistry, Mineralogy and Chemical Modelling of the Acid Crater Lake of Kawah Ijen Volcano, Indonesia. *Geochim. et Cosmochim. Acta*, Vol. 58, pp. 2445–60.

DUFF, P. MCL. D., 1993. *Holme's Principles of Physical Geology*. Chapman, 791 pp.

ELSWORTH, D.; CARRACEDO, J. C.; DAY, S. J., (eds.), 1999. Deformation and Flank Instability of Oceanic Island Volcanoes: A Comparison of Hawaii with Atlantic Island Volcanoes. *J. Volcanology and Geothermal Research,* Vol. 94, 340 pp.

FALKLAND, A.; CUSTODIO, E., 1991. *Guide on the Hydrology of Small Islands*. Studies and Reports in Hydrology, No. 49, UNESCO, Paris, 435 pp.

FAYAS, J. A.; DOMÉNECH, J., 1974. *Morfología volcánica de Olot y su interés hidrológico*. Documentos de Investigación Hidrológica, Centro de Estudios, Investigación y Aplicaciones del Agua, Barcelona.

FISHER, R. V.; SCHMINCKE, H.-U., 1984. *Pyroclastic Rocks*. Springer Verlag, 472 pp.

FRANCIS, P., 1993. *Volcanoes: A Planetary Perspective*. Oxford Univ./Clarendon Press, 435 pp.

HENRY, P.; GUY, C.; CATTIN, R.; DUDOIGNON, P.; SORNEIN, J. F.; CARISTAN, Y., 1996. A Convective Model of Water Flow in Mururoa Basalts. *Geochim et. Cosmochim. Acta,* Vol. 60, pp. 2087–109.

HERRERA, CH.; CUSTODIO, E., 2000. Saline Water in Central Fuerteventura Island, Canary Islands, Spain. *XVI Salt Water Intrusion Meeting, Miedzyzdroje, Poland.* (In press.)

HINDS, J. J.; GE, SH.; FRIDRICH, CH. J., 1999. Numerical Modelling of Perched Water under Yucca Mountain, Nevada. *Ground Water,* Vol. 37, pp. 498–504.

HUANG, K.; TZANG, Y. W.; BODVARSSON, G. S., 1999. Simultaneous Inversion of Air-Injection Tests in Fractured Unsaturated Tuff at Yucca Mountain. *Water Resources Research,* Vol. 35, pp. 2375–86.

HUNT ET AL., 1988. The Geology of North America. *The Geological Society of America,* Vol. 30, pp. 255–62.

JAWAHEER, K.; PROAG, V., 1988. Groundwater Aquifers of Mauritius. *Hydrogéologie,* Vol. 2, pp. 125–34.

JOIN, J. L.; COUDRAY, J., 1992. Exemple d'un écoulement souterrain chenalisé dans les brèches volcaniques peu perméables: la ravine Saint-Gilles, île de la Réunion. *Hydrogéologie,* Vol. 3, pp. 163–72.

——, 1993. Charactérisation géostructurale des émergences et typologie des nappes d'altitude en milieu volcanique insulaire (Île de la Réunion). *Geodinamica Acta,* Vol. 6, Paris, pp. 243–54.

KILKARNI, H.; LALWANI, A.; DEOLANKAR, S. B., 1997. Selection of Appropriate Pumping Systems for Bore Wells in the Deccan Basalt of India. *Hydrogeology Journal,* Vol. 5, pp. 75–81.

KULKARNI, H.; DEOLANKAR, S. B.; LALWANI, J. B.; PAWAS, S., 2000. Hydrogeological Framework of the Deccan Basalt Groundwater Systems, West-Central India. *Hydrogeology Journal,* Vol. 8, pp. 368–78.

LOAGUE, K., 1994. Regional Scale Ground-Water Vulnerability Estimates: Impact of Reducing Data Uncertainties for Assessments in Hawaii. *Ground Water,* Vol. 32, pp. 605–16.

LOUVAT, P.; ALLÈGRE, C. J., 1997. Present Denudation Rates on the Island of Réunion Determined by River Geochemistry: Basalt Weathering and Mass Budget between Chemical and Mechanical Erosion. *Geochim. et Cosmochim. Acta,* Vol. 61, pp. 3645–69.

MACDOUGALL, J. D., 1988. *Continental Flood Basalts.* Kluwer, 341 pp.

MIDDLEMOST, E. A. K., 1985. *Magmas and Magmatic Rocks: An Introduction to Igneous Petrology*. Longman, 266 pp.

OKI, D. S.; GIAMBELLUCA, TH. W., 1989. Groundwater Contamination by Nematicides: Influence of Recharge Timing under Pineaple Crop. *Water Resources Bull.,* Vol. 25, pp. 285–94.

PENNISI, M.; LEEMAN, N. P.; TONANINI, S.; PENNISI, A.; NABELEK, P., 2000. Boron, Sr, O, and H Isotope Geochemistry of Groundwater from Mt. Etna (Sicily): Hydrologic Implications. *Geochim. et Cosmochim. Acta,* Vol. 64, pp. 961–74.

ROBINEAU, B.; RITZ, M.; COURTEAUD, M.; DESCLOITRES, M., 1997. Electromagnetic Investigation of Aquifers in the Gran Brulé Coastal Area of Piton de la Fournaise Volcano, Réunion Island. *Ground Water,* Vol. 35, pp. 585–92.

ROBOCK, A., 2000. Volcanic Eruptions and Climate. *Reviews of Geophysics,* Vol. 38, pp. 191–219.

ROSE, T. P.; DAVISSON, M. L., 1996. Radiocarbon in Hydrologic Systems Containing Dissolved Magmatic Carbon Dioxide. *Science,* Vol. 273, pp. 1367–70.

SMELLIE, J. L., 1994. *Volcanism Associated with Extension at Consuming Plate Margins*. Geolog. Soc. London, 292 pp.

SONNENTHAL, E. L.; BODVARSSON, G. S., 1999. Constraints on the Hydrology of the Unsaturated Zone at Yucca Mountain, NV, from Three-Dimensional Models of Chloride and Strontium Geochemistry. *J. Contam. Hydrol.,* Vol. 38, pp. 107–56.

TAKASAKI, K. J.; MINK, J. F., 1985. *Evaluation of Major Dike-Impounded Ground-Water Reservoirs, Island of Oahu.* U.S. Geological Survey, Water-Supply Paper 2217, Washington D.C., 77 pp.

VIOLETTE, S.; LEDOUX, E.; GOBLET, P.; CARBONNEL, J. P., 1997. Hydrologic and Thermal Modelling of an Active Volcano: The Piton de la Fournaise, Réunion. *J. of Hydrology,* Vol. 191, pp. 37–63.

WATTS, A. G.; PEIRCE, C.; COLLIER, J.; DALWOOD, R.; CANALES, J. P.; HEUSTOCK, T. J., 1997. A Seismic Study of Lithospheric Flexure in the Vicinity of Tenerife, Canary Islands. *Earth and Planetary Sci. Letters,* Vol. 196, pp. 431–47.

APPENDICES

Appendix A-1 Conversion factors for physical data

To convert	*To*	*Multiply by*
Grams	Ounces (avoirdupois)	0.03527
Ounces (Avoirdupois)	Grams	28.35
Tons per acre	Metric tons per hectare	2.2417
Grains per gallon	Milligrams per litre	17.12
Milligrams per litre	Grains per gallon	0.05841
Milligrams per litre	Tons per acre-foot	0.001360
Gallons (Imperial)	Gallons (U.S.)	1.2009
Gallons (U.S.)	Litres	4.55
Litres	Gallons (U.S.)	0.219
Cubic-feet/second (cfs)	Gallons (U.S.) per minute	448.8
Cubic-feet/second/day*	Acre-feet	1.983471
Cubic-feet/second/day	Gallons (U.S.) per day	646,317
Acre-feet	Gallons (U.S.)	325,851
Acre-feet	Cubic feet	43,560
Acre-feet	Cubic metres	1,232
Cubic feet	Cubic metres	0.028317
Cubic feet	Gallons (U.S.)	7.481

* 1 sec-ft day = 1cfs for 24 hr.

Appendix A-2 Conversion factors for hydrochemical data

milligrams/litre (mg/l) * F1 = milliequivalents/litre (meq/l);
milligrams/litre * F2 = millimoles/litre (mmol/l).

Element and reported species	*F1*	*F2*
Aluminium (Al^{+++})	0.11119	0.03715
Ammonium (NH_4^+)	0.05544	0.05544
Barium (Ba^{++})	0.01456	0.00728
Beryllium (Be^{++})	0.22192	0.11096
Bicarbonate (HCO_3–)	0.01639	0.01639
Boron (B)		0.09250
Bromide (Br^-)	0.01251	0.01261
Cadmium (Cd^{++})	0.01779	0.00890
Calcium (Ca^{++})	0.04990	0.02495
Carbonate (CO_3^{--})	0.03333	0.01666
Chloride (Cl^-)	0.02821	0.02821
Chromium (Cr)		0.01923
Cobalt (Co^{++})	0.03394	0.01697
Copper (Cu^{++})	0.03148	0.01574
Fluoride (F^-)	0.05264	0.05264
Germanium (Ge)	0.01378	
Gallium (Ga)	0.01434	
Gold (Au)	0.00511	
Hydrogen (H^+)	0.99209	0.99209
Hydroxide (OH^-)	0.05880	0.05880
Iodide (I^-)	0.00788	0.00788
Iron (Fe^{++})	0.03581	0.01791
Iron (Fe^{+++})	0.05372	0.01791
Lead (Pb)	0.00483	
Lithium (Li^+)	0.14411	0.14411
Magnesium (Mg^{++})	0.08226	0.04113
Manganese (Mn^{++})	0.03640	0.01820
Molybdenum (Mo)	0.01042	
Nickel (Ni)	0.01703	
Nitrate (NO_3^-)	0.01613	0.01613
Nitrite (NO_2^-)	0.02174	0.02174
Phosphate (PO_4^{--})	0.03159	0.01053
Phosphate (HPO_4^{--})	0.02084 .	0.01042
Phosphate ($H_2PO_4^-$)	0.01031	0.01031
Potassium (K^+)	0.02667	0.02557
Rubidium (Rb^+)	0.01170	0.01170
Silica (SiO_2)	0.01664	
Silver (Ag)	0.00927	
Sodium (Na^+)	0.04350	0.04350
Strontium (Sr^{++})	0.02283	0.01141
Sulphate (SO_4^{--})	0.02082	0.01041
Sulphide (S^{--})	0.06238	0.03119
Titanium (Ti)	0.02088	
Uranium (U)		0.00420
Zinc (Zn^{++})	0.03060	0.01530

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