

# STRATIFICATION OF LAKES

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[1] Many lakes show vertical stratification of their water masses, at least for some extended time periods. Density differences in water bodies facilitate an evolution of chemical differences with many consequences for living organisms in lakes. Temperature and dissolved substances contribute to density differences in water. The atmosphere imposes a temperature signal on the lake surface. As a result, thermal stratification can be established during the warm season if a lake is sufficiently deep. On the contrary, during the cold period, surface cooling forces vertical circulation of water masses and removal of gradients of water properties. However, gradients of dissolved substances may be sustained for periods much longer than one annual cycle. Such lakes do not experience full overturns. Gradients may be a consequence of external inflows or groundwater seepage. In addition, photosynthesis at the lake surface and subsequent decomposition of organic material in the deeper layers of a lake can sustain a gradient of dissolved substances. Three more geochemical cycles, namely, calcite precipitation, iron cycle, and manganese cycle, are known for sustaining meromixis. A limited number of lakes

do not experience a complete overturn because of pressure dependence of temperature of maximum density. Such lakes must be sufficiently deep and lie in the appropriate climate zone. Although these lakes are permanently stratified, deep waters are well ventilated, and chemical differences are small. Turbulent mixing and convective deep water renewal must be very effective. As a consequence, these lakes usually are not termed meromictic. Permanent stratification may also be created by episodic partial recharging of the deep water layer. This mechanism resembles the cycling of the ocean: horizontal gradients result from gradients at the surface, such as differential cooling or enhanced evaporation in adjacent shallow side bays. Dense water parcels can be formed which intrude the deep water layer. In the final section, stratification relevant physical properties, such as sound speed, hydrostatic pressure, electrical conductivity, and density, are discussed. The assumptions behind salinity, electrical conductance, potential density, and potential temperature are introduced. Finally, empirical and theoretical approaches for quantitative evaluation from easy to measure properties conclude this contribution.

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## 1. INTRODUCTION

[2] Humanity needs to be prepared for the changes it imposes upon the Earth, especially in times of global change and of direct human impact on the hydrological cycle. The anthropogenic impact of the last decades on our aquatic environment has shown that a responsible use of our natural resources is mandatory to guarantee sustainable conditions. Over the last decades, not only have new large water bodies, e.g., reservoirs, been created on the Earth's surface, but also entirely novel aquatic systems developed in the aftermath of mining in abandoned opencasts [e.g., Miller *et al.*, 1996; Krüger *et al.*, 2002]. In addition, the hydrological regime of many lakes has been modified to the extent that lakes have fundamentally changed their appearance, e.g., Aral Sea [Létolle and Mainguet, 1993; Crétaux *et al.*, 2005], or their stratification pattern has been altered by

human impact, e.g., Dead Sea [Gat, 1995] and Mono Lake [Jellison and Melack, 1993] or by climatic variability, e.g., Caspian Sea [Peeters *et al.*, 2000].

[3] Understanding environmental processes is the first step to forecasting such evolutions. Furthermore, a good understanding of stratification is required not only to prognosticate the future of our lakes but also to better interpret the past from lacustrine sediment cores [e.g., Kjensmo, 1988; Brauer, 2004]. Those interpretation's intent is not only to document the lake history but also to draw conclusions about the climate in the past by relating variations of the lake sediments to circulation patterns in lakes and finally climatic conditions triggering these changes.

[4] A considerable portion of the lakes on Earth are permanently stratified. Deep lakes especially show this feature. As a consequence, many of the largest lakes (e.g., Caspian Sea, Baikal, Tanganyika, Malawi-Nyasa) and many middle size and small lakes do not circulate completely in the vertical and do not show a homogenized, overturning water body at any time during the annual cycle. The permanent stratification has decisive impact on the redistri-

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bution of dissolved substances, such as nutrients or oxygen, and hence determines the biocenosis that can form in the lake [e.g., *Gaedke et al.*, 1998; *George and Hewitt*, 2006; *Thackeray et al.*, 2006].

[5] Observations of stratified lakes probably date back to long before the start of scientific literature on the environment. The Romans likely already knew about stratification. More recently, solar ponds, i.e., saltwater ponds capped with a less saline surface layer, have been utilized for breeding oysters at least since the 19th century in Norway [see *Klaveness*, 1990]. *Findenegg* [1933] brought chemical differences in connection with incomplete mixing in winter. Since then, considerable work has been published on ecological consequences of meromixis. Also, the reasons for permanent stratification have been investigated, and the understanding of involved processes has progressed. Textbooks on hydrobiology and limnology present basic knowledge about stratification and circulation of lakes [e.g., *Forel*, 1901; *Ruttner*, 1940; *Cole*, 1983; *Lampert and Sommer*, 1993; *Schwoerbel*, 1999; *Wetzel*, 2001; *Schönborn*, 2003]. Some of these books refer to permanent stratification and meromixis and mention some ecological consequences. However, the space dedicated to the physical and chemical preconditions for permanent stratification in any of these texts is limited. *Hutchinson* [1957] has attempted to summarize the knowledge on meromixis of his time, and his work is still a valid reference on meromixis at the beginning of the 21st century. In addition, there are some studies on special aspects of limnology [*Tilzer and Serruya*, 1990; *Margalef*, 1994; *Lerman et al.*, 1995; *Pourriot and Meybeck*, 1995; *Johnson and Odada*, 1996; *Taub*, 1984] which touch upon interesting aspects of physical limnology, especially *Lerman et al.* [1995], who deal with chemical peculiarities which may be linked to permanent stratification. On the contrary, the studies on physical limnology, i.e., stratification and mixing in lakes [e.g., *Fischer et al.*, 1979; *Imboden and Wüest*, 1995; *Imberger and Patterson*, 1990], and modeling currents in lakes [e.g., *Hutter*, 1987] have concentrated on thermally stratified lakes and have widely neglected the topic of meromixis. For considerations about internal waves and currents in lakes, the oceanographic literature [e.g., *Pedlosky*, 1988; *Gill*, 1982; *Pickard and Emery*, 1982; *Pond and Pickard*, 1983] has delivered the necessary theoretical knowledge for physical limnology but to date has not contributed a comprehensive work with focus on permanent stratification in adjacent seas, such as the Black Sea and the Baltic Sea.

[6] Other than in the open ocean, chemical transformations can play an important role in stratifying a lake. Lakes greatly vary in salinity and chemical composition of dissolved substances. As a consequence, freshwater assumptions and ocean water approximations may be acceptable for some considerations, while for many lakes, or more detailed investigations, they fail. Such cases require separate approaches to satisfactorily reflect given conditions.

[7] This review on stratification of lakes is also intended to be useful for limnologists of educational backgrounds different from physics. Furthermore, it shall also show

physical limnologists how biogeochemical processes determine the physical properties of stratification. In order not to interfere with the progress of reading, we tried to remove all quantifications of processes from sections 1–4. Such numerical approximations are included in section 5, which is dedicated to the evaluation of stratification relevant physical quantities.

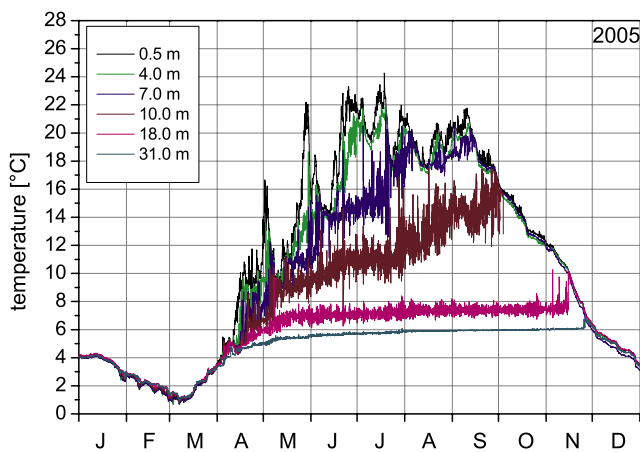
[8] Referring to the classification of ectogenic, crenogenic, and biogenic meromixis by *Hutchinson* [1957], we insert more recent observations into these classes and place lakes that stay permanently stratified because of thermobaric effect with the meromictic lakes. In addition, we include lakes with episodic partial deep water renewal, where a process can be identified that recharges and replaces part of the deep water on a regular basis. Section 2.1, introducing the annual temperature stratification cycle, briefly repeats the formation of an epilimnion and the deep recirculation, while clarifying the naming convention of mixis types and layers in a lake. Section 5 introduces and explains physical quantities which are relevant for the stratification in lakes and lists the most common quantitative expressions for their approximation for given conditions in lakes.

## 2. STRATIFICATION FEATURES OF HEAT AND DISSOLVED SUBSTANCES

### 2.1. Annual Temperature Cycle in Lakes

[9] The lake surface shows a pronounced temperature cycle over the year in most climate zones on Earth (Figure 1). Besides the thermal contact with the atmosphere, the common exposition of lakes and atmosphere to seasons, namely, the variation of incoming solar radiation, is the main reason for the parallel evolution. The depicted temperature cycle is typical for lakes of the temperate climate zone. Only lakes in close proximity of the equator (or permanently ice-covered lakes) may show little variation in the surface temperatures. A closer look reveals several processes contributing to the heat transfer through the lake surface: solar radiation, long-wave radiation of atmosphere and surface waters, sensible heat exchange, and heat flux connected with evaporation and precipitation [see, e.g., *Imboden and Wüest*, 1995]. In addition, contributions of inflow and outflow of surface and groundwater, as well as thermal contact with the lakebed, must be considered in some cases.

[10] Weather conditions set up stratification and currents in lakes. They directly and indirectly control the biological processes in the lake. Hence for the detailed investigation of physical processes in lakes, observation of the meteorological conditions above the water is mandatory (Figure 2). Waves and currents are driven by wind. For heat transfer measurements, air temperature and net radiation sensors are also required, which refer to the part of incoming solar radiation that is not reflected (albedo) but deposits its energy as heat in the water. In addition, air humidity measurements are required to estimate the evaporation from the water surface, and finally, many meteorological stations also include air pressure as a central meteorological magnitude.



**Figure 1.** Temperatures at several depths in Lake Goitsche near Bitterfeld, about 100 km southwest of Berlin in Germany during 2005 (hourly mean). Depths of temperature tracks are identified from higher temperatures to lower temperatures for increasing depth. Lake Goitsche formed in a decommissioned lignite mine pit over the years 1999 to 2002. It has a surface area of 13.5 km<sup>2</sup> and a maximum depth of about 50 m.

[11] A look at temperature time series from the surface waters of lakes within the same climate (Figure 3) reveals that lake surface temperatures reflect a climate signal. The daily average temperature is very similar between the displayed lakes. Only slight differences can be seen in spring, when the smaller lakes tend to gain heat faster, and in hot periods during summer, when the highest peaks of smaller lakes are 1° or 2° higher [see *Boehrer et al.*, 2000b]. The situation is different for temperature tracks acquired at a water depth of 7 m in the same lakes. Obviously, the mixing characteristics of each lake determine how effectively heat is being transferred vertically.

[12] While the surface water is exposed to solar radiation, heat loss by long-wave radiation, and thermal contact with the atmosphere, the deeper layers are shielded from the major sources of heat. Diffusive heat transport on a molecular level is very slow and requires a time period of the order of a month for the transport of heat over a vertical distance of 1 m. A much more efficient heat transport can be facilitated by the kinetic energy input through wind stress at the lake surface.

[13] If surface temperatures exceed the 4°C threshold, warmer and hence less dense water can only be mixed to a limited depth, which corresponds to the limited energy budget supplied by the wind. A density difference has been established, and the thermal stratification period of summer starts. In sufficiently deep lakes, thermal stratification holds until cooler autumn and winter temperatures force a circulation into deep waters. The warm surface water layer is called epilimnion, while the colder water layer beneath, which has not been mixed into the epilimnion over the stratification period, is called hypolimnion. A sharp temperature gradient (thermocline) forms in the contact zone between both layers (Figure 4).

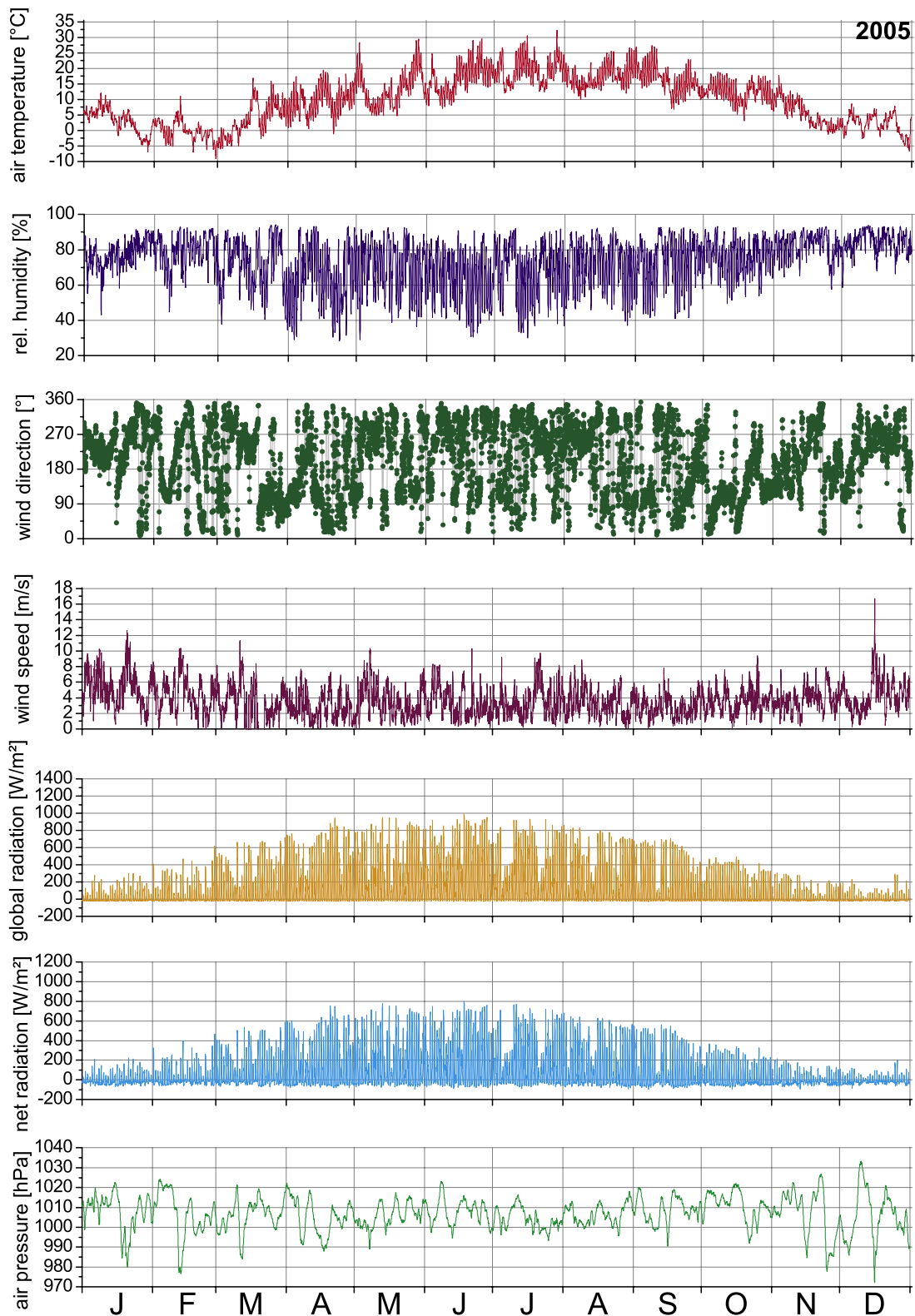
[14] Epilimnion and atmosphere exchange heat and volatile substances such as gases. In addition, the epilimnion is recirculated episodically by wind events or periods of lower temperatures during the stratification period. On the contrary, the hypolimnion is insulated from exchange with the atmosphere during the stratification period. Transport of dissolved matter across the vertical density gradient of the thermocline is usually small. In most lakes, only chemical precipitation and settling of particles are able to carry larger amounts of matter across the thermocline.

[15] In nearly all lakes, wind is the decisive factor for the thickness of the epilimnion. There are exceptions, like lakes with extremely weak winds, where other factors can evolve to be the most important. One example is a very clear lake, where light penetrates beyond the mixing depth of wind [e.g., *Simpson and Dickey*, 1981a, 1981b; *Kling*, 1988; *Mazumder and Taylor*, 1994; *Fee et al.*, 1996; *Pérez-Fuentetaja et al.*, 1999]. In addition, forming an ice cover may be regarded as a special case where molecular diffusion of heat is responsible for the thickness of the surface ice layer.

[16] The epilimnion thickness is not constant over the stratification period (Figure 4). In spring, a thin layer is formed, which throughout the summer gradually gains thickness because of wind action. It is not until autumn that colder water from the lake surface can erode the stratification to the hypolimnion at a larger rate. During this later period of thermal stratification, substances dissolved in hypolimnetic waters, such as nutrients, become available in the epilimnion again. If surface temperature falls enough, epilimnion and hypolimnion can be mixed, and the entire lake is homogenized (circulation period) to one layer, the so-called mixolimnion. As an example, the acidic residual lakes in Goitsche Mine Pit, which were neutralized by introduction of river water, are presented (see Figure 5). Chemical gradients (such as pH) could only be sustained for extended time periods if they were stabilized by density stratification (Figure 5, bottom).

[17] Epilimnion thickness is important for many organisms. Hence regressions relating epilimnion thickness with a priori known or easy to measure quantities have been proposed. Wind speed is considered through lake surface area  $A$  and a fitted exponent. *Jöhnk* [1999, 2000] collected the results of several authors (Figure 6). All data sets rely on measurements from moderate latitudes of the Northern Hemisphere, i.e., central Europe, North America, and Japan. The most central regression originates from *Patalas* [1984]  $z_{\text{epi}} = 4.6A^{0.205}$  which is close to previously used formulas from *Ventz* [1972] and *Fachbereichsstandard* [1983] (as cited by *Klapper* [1992]). The differences between fitted curves (factor 1.5) give a good impression of the accuracy at which epilimnion thickness can be parameterized with surface area only. Also, *Gorham and Boyce* [1989] confirm surface area and fetch as the most important factors for the epilimnion thickness, as does *Davies-Colley* [1988] for lakes in New Zealand.

[18] Because of its high gradients, the thermocline can form a special habitat. Organisms that can regulate their

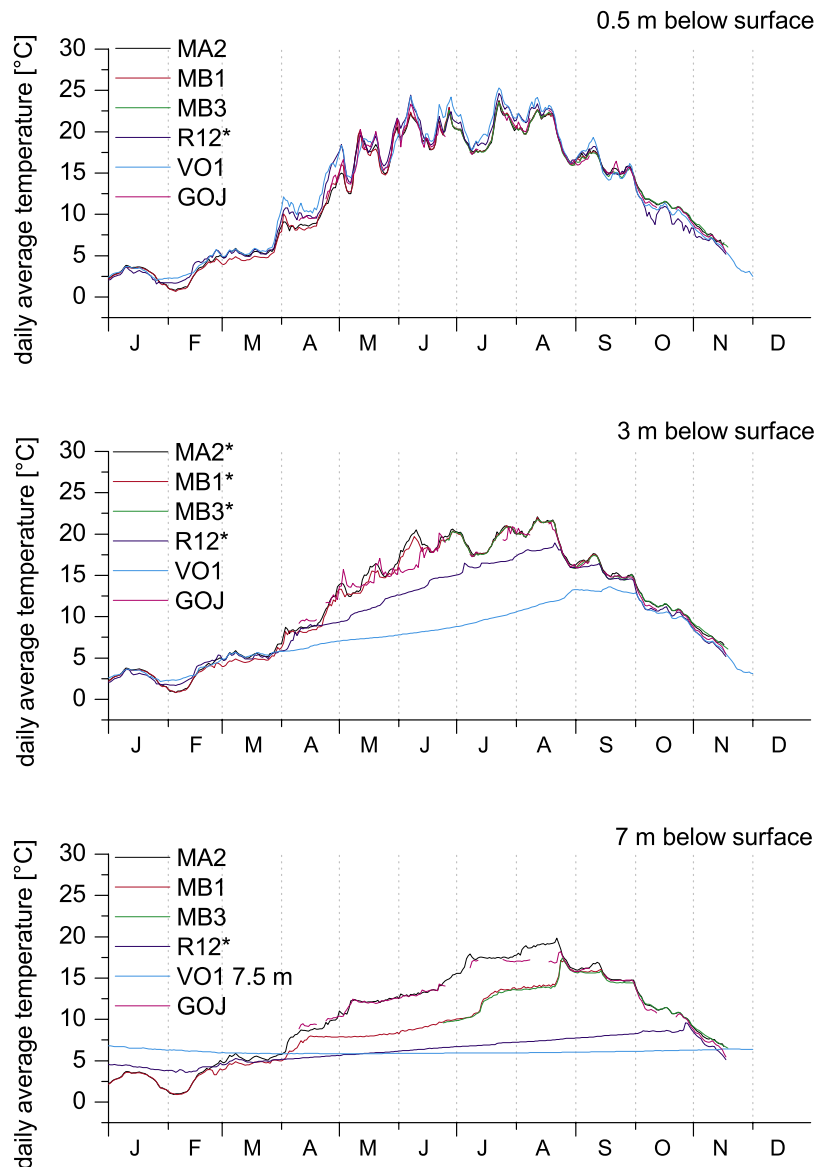


**Figure 2.** Meteorological data (hourly mean) of air pressure, net radiation, global radiation, wind speed, wind direction ( $0^\circ = 360^\circ =$  north,  $90^\circ =$  east), relative humidity, and air temperature acquired at 2.6 m above the lake surface of Lake Goitsche during 2005.

density can position themselves in the strong density gradient. Also, inanimate particles can accumulate on a level representing their density. In addition, motile organisms can dwell in the thermocline (to profit from the

advantages of both layers), epilimnion, and hypolimnion. As a consequence, a separate layer may be identified because of its own characteristic. Such a layer is called the metalimnion. Especially in nutrient-rich lakes, the

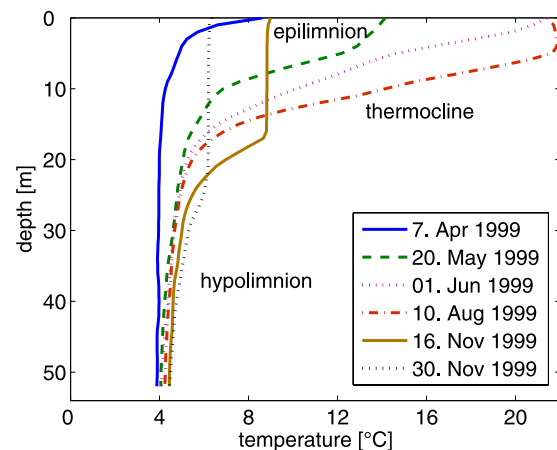




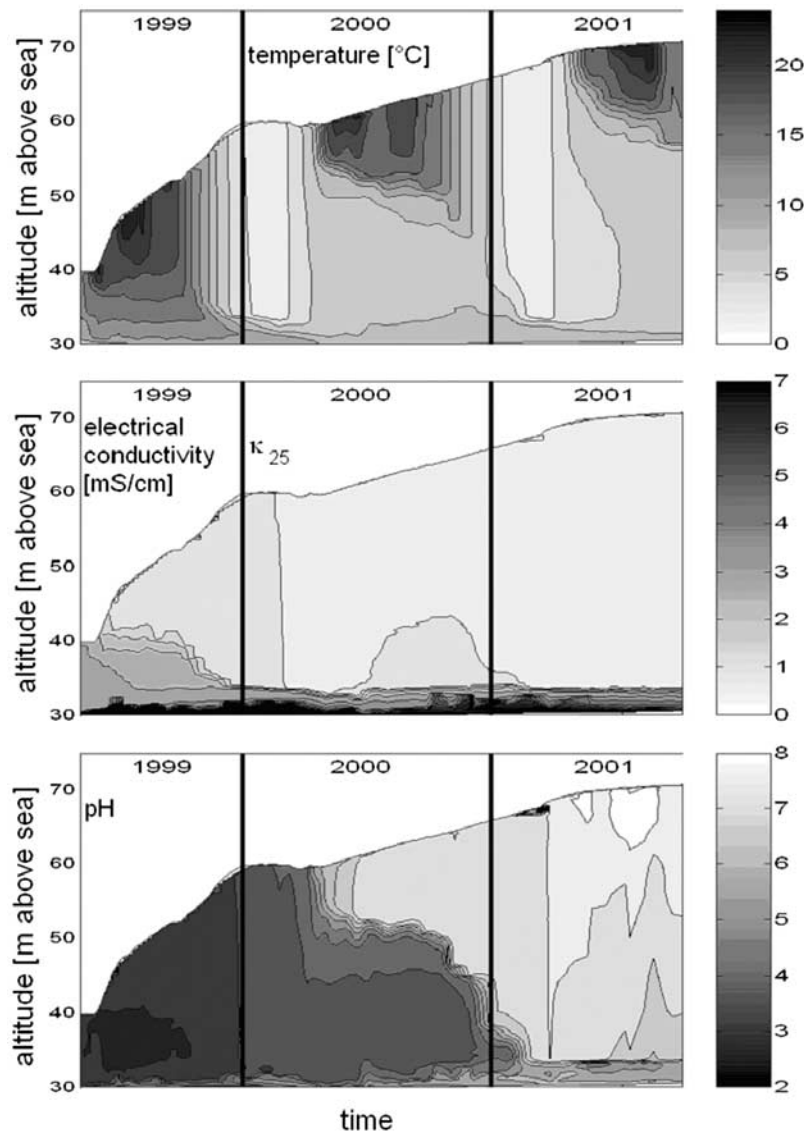
**Figure 3.** Daily average temperatures in several lakes of the northern German plain (MA2 Wallendorfer See, MB1 und MB3 Rassnitzer See, R12 Mining Lake 111, VO1 Vollert-Süd, GOJ subbasin Niemgk of Lake Goitsche) acquired at 0.5, 3, and 7 m below the surface during 1998 (modified from *Boehrer et al.* [2000b], with permission from Elsevier).

decomposition of organic material can cause a depletion of oxygen, resulting in a so-called metalimnetic oxygen minimum (Figure 7). On the contrary, if light can penetrate to the thermocline and photosynthesis can overcome the oxygen uptake in this zone, a metalimnetic oxygen maximum can be encountered. Occasionally, this phenomenon can be observed in mesotrophic lakes.

[19] Traditionally, the imagination of processes and naming convention are dominated by experiences from lakes in moderate climates. In the vicinity of the equator, some features are different, e.g., the annual temperature variation is relatively small. As a consequence, the temperature difference between epilimnion and hypolimnion is also small. Under favorable conditions, night temperatures may suffice to form cold waters in shallow areas dense enough to intrude into the hypolimnion (similar to deep water renewal,



**Figure 4.** Temperature profiles of Lake Mondsee, Austria, on several dates during the stratification period of the year 1999 (data in part from *Dokulil and Teubner* [2003]).



**Figure 5.** Contour plot of temperature, electrical conductance ( $\kappa_{25}$ ), and pH versus time since flooding started on 7 May 1999 and depth in mining lake Goitsche (station XN3) during flooding with river water [see also Boehrer *et al.*, 2003] (with permission from ecomed). The increasing shaded area reflects rising water surface.

see section 4) or even trigger a complete overturn (“polymictic” [e.g., Zauke *et al.*, 1995]). Hence some features of the stratification cycle may look different in a tropical or polar climate. We refer to Lewis [1987] for more detail on the special features of tropical lakes; for stratification features in a polar climate, see, e.g., Gibson [1999].

[20] The classification of lakes according to their circulation pattern has proven very useful within the limnology community [e.g., Hutchinson, 1957]:

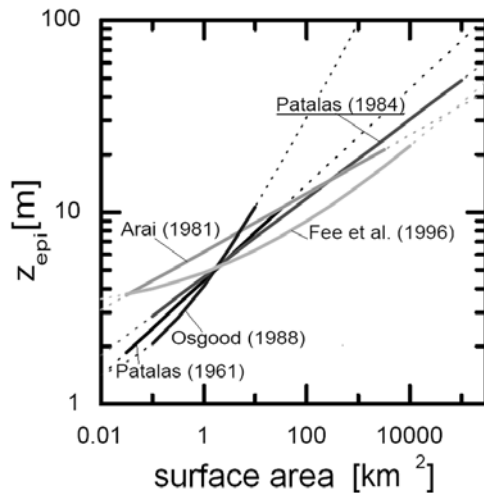
[21] 1. Holomictic lakes overturn and homogenize at least once a year.

[22] 2. Meromictic lakes are lakes in which the deep recirculation does not include the entire water body. Beyond this traditional definition, most limnologists pragmatically call a lake meromictic if a chemically different (e.g., anoxic) bottom layer, called monimolimnion, has continuously been present for at least one annual cycle.

[23] 3. Episodic partial deep water renewal lakes can have the appearance of meromictic lakes despite the presence of recycling mechanisms. In the end, it turns out that these recycling mechanisms keep the lakes stratified.

[24] 4. Amictic lakes do not experience a deep recirculation. Usually, permanently ice-covered lakes are included in this class. Lakes, however, can circulate underneath an ice sheet. Transport of both heat and matter happens because of diffusion, or double diffusion [e.g., Brandt and Fernando, 1996]. On top of this, they can also be circulated by external forcing, such as solar radiation that penetrates to the lake bed and geothermal heat flux, or by salinity gradients created when ice is forming on a salt lake.

[25] Holomictic lakes are subdivided into classes indicating the frequency and time of overturn. We refer to the most common classes, while Lewis [1983] subdivided the classes even further. “Polymictic” refers to lakes which are not



**Figure 6.** Graphical representation of several approximations of epilimnion thickness  $z_{\text{epi}}$  versus surface area of respective lakes (slightly modified from Jöhnk [2000] with author's permission).

deep enough to form a hypolimnion. The entire lake behaves like an epilimnion, which is mixed by sporadic strong wind events over the year or even on a daily basis in response to strong daily temperature cycle. “Dimictic” lakes are handled as prototypes of lakes in moderate to cold climates. A closer look at the lakes, however, reveals that in most cases an ice cover or a great maximum depth is required to guarantee a stratification period during the cold season. Intermittent between ice cover and summer stratification the lake can be circulated in the vertical, the easiest when in freshwater lakes, the temperature traverses  $4^{\circ}\text{C}$  (see Figure 1). “Monomictic” lakes possess one circulation period in addition to the stratification period. Many lakes in the temperate climate belong into this class if they do not develop an ice cover during winter. Sometimes such lakes are also referred to as warm monomictic to distinguish them from cold monomictic lakes, which show an ice cover for most of the year and circulate during the short period without ice. “Oligomictic” lakes circulate less frequently than once a year, normally at irregular intervals, triggered by extreme weather conditions such as unusually cold winters for the respective location.

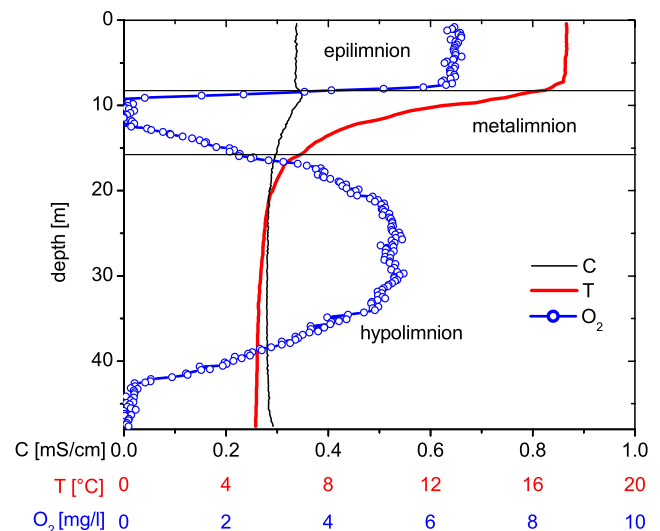
## 2.2. Salt Stratification

[26] A considerable portion of the Earth's inland water is salty [Williams, 1996]. Many large salt lakes, e.g., Caspian Sea, Issyk-Kyl, Aral Sea, Lake Van, Great Salt Lake, and the Dead Sea, are located in endorheic basins, i.e., areas on the Earth's surface without surficial hydraulic connection to the world ocean [e.g., Meybeck, 1995]. Williams [1996] presents a world map of the largest of these basins but also points out that salt lakes occur outside these areas, such as solar ponds [e.g., Kirkland et al., 1983] or basins filled with seawater that lost the connection to the sea. In addition, the high salinity of few inland salt lakes derives from underground or local salt sources, often associated with mining [e.g., Kirkland et al., 1983; Böhrer et al., 1998].

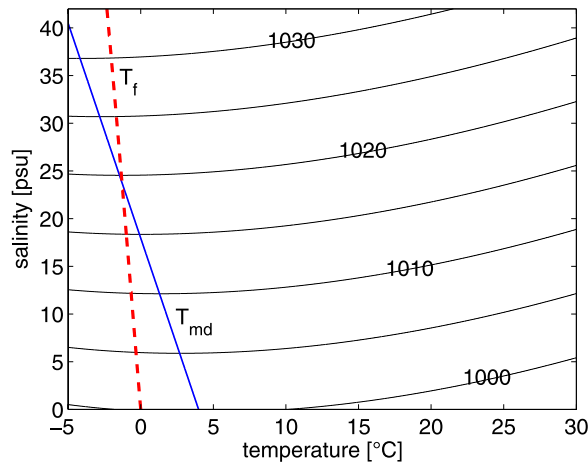
[27] In the absence of any clear boundary, lakes are referred to as salt lakes when their salt content lies above  $3 \text{ g}$  in a kilogram of lake water; i.e.,  $3 \text{ g kg}^{-1} = 3\text{‰}$ . From this concentration, humans can clearly taste the salt, and ecological consequences become obvious [e.g., Williams, 1996, 1999]. The salt content in lakes, however, can be as high as  $300 \text{ g kg}^{-1}$  [e.g., Williams, 1998]. For limnetic waters, salinity has been defined as the sum of all ion concentrations [Williams, 1994]. It is measured in  $\text{g kg}^{-1}$  or part per thousand (ppt). Direct salinity measurements hence require a full chemical analysis. By definition, freshwater lakes have a salt content below  $3 \text{ g kg}^{-1}$ . However, salinities normally lie below  $0.5 \text{ g kg}^{-1}$ . Even salinity gradients in this range can determine the circulation pattern of lakes.

[28] The composition of dissolved substances in lakes varies greatly. The surrounding geological formations release substances that are carried into the lake [e.g., Rodhe, 1949; Jellison et al., 1999]. The contributions of the chemical species to physical properties, such as electrical conductivity or density, are specific, and as a consequence, lake-specific approaches are required for numerical approximations of electrical conductance and density from conductivity-temperature-depth (CTD) probe data. Highly accurate approximations are only available for a limited number of lakes. One particular composition of salt water that is familiar and accessible to many limnologists is brackish water, i.e., water mixed from ocean water and freshwater. Lakes close to the coast that have received inflows from the open sea may well represent such conditions.

[29] As a consequence, effects of salinity on the properties of lake water will be introduced for brackish water, and



**Figure 7.** Profiles of temperature ( $T$ ), (in situ) conductivity ( $C$ ), and concentration of dissolved oxygen ( $\text{O}_2$ ) from 6 September 2000 in Arendsee, Germany (adapted from Boehrer and Schultze [2005] with permission from ecomed). The boundaries between layers were drawn along gradients in oxygen profiles. (Oxygen concentration numerically corrected for response time of 7.5 s of sensor.)



**Figure 8.** Density contours versus temperature and salinity of mixtures of seawater and pure water in  $\text{kg m}^{-3}$  at normal atmospheric pressure. The dashed line shows freezing point respective to salinity, and the thinner solid line represents temperature of maximum density at given salinity.

we selected lakes where brackish conditions are a good approximation. As such conditions can commonly be encountered, appropriate approaches and highly accurate numerical approximations have been developed and are widely accepted. Salinity in the ocean environment is calculated with the UNESCO formula from measurements of electrical conductivity and temperature (see section 5.3). The practical salinity unit (psu) follows the idea of above mentioned salt content in  $\text{g kg}^{-1} = \text{‰}$  and represents a reasonably good value for the ocean water (salinity of about 35 psu) and brackish water, though concerns about accuracy with the contained salt have resulted in debates about a new ocean standard for salinity (see also section 5.3.2). The composition of salts in lakes can, however, greatly deviate from ocean conditions. In such cases, salinity in psu can (if at all) only be used with reservation.

[30] Salt contributes to water density [e.g., *Pickard and Emery, 1982*] (Figure 8 and section 5.4). This is important, as density gradients are implied by salinity differences due to inflowing rivers or evaporation and precipitation. In many cases, density gradients due to salinity overcome

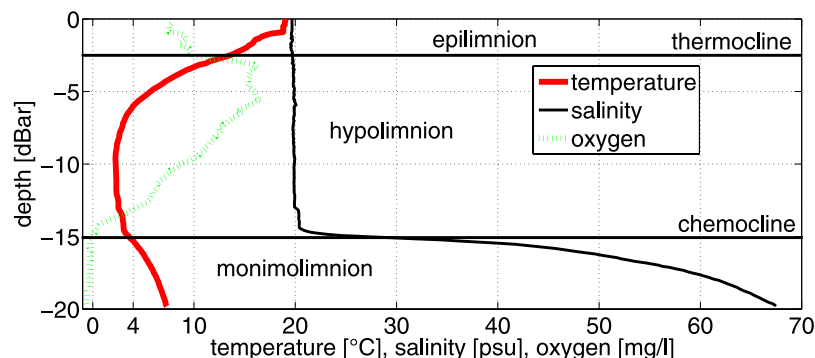
the thermal contribution. Some lakes show extreme salinity stratification between low salinity and water beyond ocean concentration. Also, in deep waters at a temperature close to the maximum density, supposedly small salinity gradients can dominate the stratification.

[31] The salinity contribution to density shows a temperature dependence. The same amount of salt increases the density of colder water slightly more than of warmer water. As a consequence, temperature of maximum density decreases with increasing salinity. At about 24 psu, temperature of maximum density intersects with the freezing point. Water of higher salinity does not show the anomaly of fresh water. This affects the hypolimnion temperature in cases close to the temperature of maximum density. In Solbad Stassfurt, the salt composition is similar to the ocean. During summer, hypolimnion temperatures stay below  $4^{\circ}\text{C}$  because of high salinity of the mixolimnion (Figure 9) [*Hausmann and Boehrer, 2006*]. For the same reason, deep ocean waters are considerably colder than  $4^{\circ}\text{C}$ .

[32] Increasing salinity also lowers the freezing point of water. Seawater freezes at  $-2^{\circ}\text{C}$  (Figure 8); lake water of even higher salinity freezes at correspondingly lower temperatures. *Kerry et al. [1977]* (as cited by *Hobbie [1996]*) report that the highly saline Deep Lake in Antarctica circulates in winter at temperatures of  $-15^{\circ}\text{C}$ , while in the southern summer, an epilimnion forms with temperatures up to  $10^{\circ}\text{C}$ .

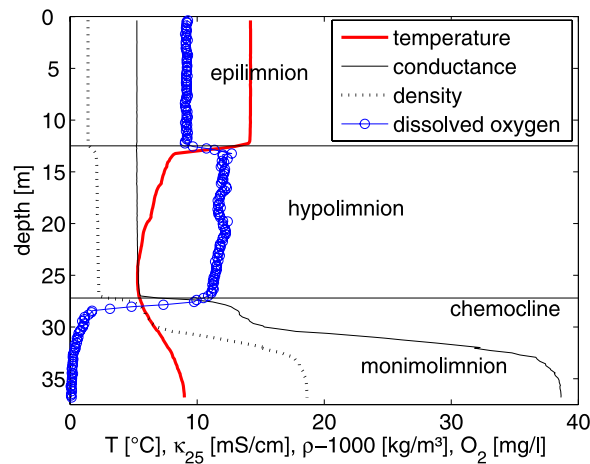
### 2.3. Suspended Matter

[33] Rivers can transport high loads of suspended material, especially at high flows. Concentrations of several hundred milligrams per liter have regularly been recorded. For settling particles, their mass adds to bulk density of the surrounding water, as long as the average speed of precipitation is constant. In some cases, suspended material is the crucial contribution to facilitate deep water renewal [e.g., *Hürzeler et al., 1996*]. Alterations in the catchment of inflows into Lake Brienz strongly reduced the amount of suspended matter, and as a consequence, incoming water could not replace the deepest water layers anymore [*Finger et al., 2006*]. An extreme example was the Mulde flood of 2002, which washed out a new river bed from fine sandy material. Concentration of suspended material was so high



**Figure 9.** Profiles of temperature, salinity, and oxygen concentration in Solbad Stassfurt on 3 May 2005 (see also *Hausmann and Boehrer [2006]* with permission from Weißensee).





**Figure 10.** Profiles of temperature, conductance, density, and dissolved oxygen from Rassnitzer See in former mining area Merseburg-Ost on 7 October 2003 (adapted from Boehrer and Schultze [2005] with permission from ecomed). (Oxygen concentration numerically corrected for response time of 7.5 s of the sensor).

( $1 \times 10^6 \text{ m}^3$  of suspended material in  $100 \times 10^6 \text{ m}^3$  of flood water) that it could replace an entire hypolimnion despite the seasonal temperature stratification [Boehrer *et al.*, 2005]. Frey [1955] even claims that suspended matter in deep waters initiated meromixis (see section 3) in Längsee, Austria.

[34] Casamitjana and Roget [1993] report fluidized lake beds where thermal springs keep sediment suspended. The resulting bulk density suffices to overcome the adverse temperature gradient. Two layers clearly separated by temperature and suspended matter are the consequence. Finally, particles that are precipitated in the contact zone of chemically different layers, e.g., mixolimnion and monimolimnion, contribute their weight to the density structure (see section 5.4). Particles can be identified through higher turbidity in profiles (see also section 5.4).

### 3. MEROMIXIS: PERMANENT STRATIFICATION

[35] Lakes are called meromictic if a chemically different bottom layer, called monimolimnion, has continuously been present for at least one annual cycle. Higher concentrations of dissolved substances have increased density sufficiently to resist deep recirculation. Exchange rates with the mixolimnion are small enough that chemically different conditions are sustained continuously.

[36] Famous examples are the lakes of Carinthia, Austria [Findenegg, 1933, 1935], but many mining lakes are also meromictic [e.g., Davis and Ashenberg, 1989; Böhrer *et al.*, 1998; Stevens and Lawrence, 1998; Rücker *et al.*, 1999; Fisher, 2002; Denimal *et al.*, 2005]. Some small and deep maar lakes [Scharf and Oehms, 1992; Scharf and Menn, 1992], as well as natural lakes in southern Norway or Finland [Ström, 1945; Merilainen, 1970; Hongve, 1997, 2002], are permanently stratified by small concentration differences between mixolimnion and monimolimnion. If

sufficiently deep, gravel pit lakes in the upper Rhine Valley tend to be meromictic [e.g., Landesanstalt für Umweltschutz Baden-Württemberg, 2000; Ilmberger and von Rohden, 2001].

[37] Monimolimnia are excluded from gas exchange with the atmosphere for years. In most cases, anoxia is established after sufficient time. Nitrate and sulphate can function as electron acceptors for microbial oxidation of organic material. In this zone, substances can be produced that would not be chemically stable in the mixolimnion and would soon be oxidized under oxic conditions. The monimolimnion is permanently exposed to the hydrostatic pressure of the water column above. Hence gases ( $\text{CO}_2$ ,  $\text{H}_2\text{S}$ , and others) can accumulate in concentrations far beyond concentrations in mixolimnia (e.g., Lake Monoun or Lake Nyos in Cameroon, Africa [Halbwachs *et al.*, 2004]).

[38] In many meromictic lakes, deep recirculation erodes the monimolimnion, leaving a sharp gradient at the end of the circulation period. The transition of all water properties happens within a few decimeters from mixolimnetic values to monimolimnetic values (see Figure 10). This sharp gradient is called halocline, chemocline, or pycnocline, depending on whether salinity gradient, chemical gradient, or density gradient is referred to. From observations of intensive colonization with only few different species, it is known that some plankton species can take advantage of such gradients (e.g., Lago Cadagno in the Swiss Alps [Camacho *et al.*, 2001; Tonolla *et al.*, 2003] and Lake Bolvod, Gek Gel, and Maral Gel [see Sorokin, 1970]).

[39] Monimolimnia are thermally locked between groundwater ( $\sim 10^\circ\text{C}$  in the depicted case of Figure 10) and the hypolimnion. While groundwater roughly represents annual average temperature, the hypolimnion conserves surface temperature of the lake during the cold season or lies close to temperature of maximum density. As a consequence, monimolimnia show temperatures within these boundaries. Usually, a continuous temperature profile with depth can be observed from the hypolimnion toward groundwater temperature.

[40] Traditionally, meromictic lakes are classified after main reasons for the creation of permanent stratification. Early studies (see Yoshimura [1937] and Findenegg [1937], as cited by Hakala [2004]) identified both external forcing and internal processes responsible for sustaining meromixis. Finally, Hutchinson [1957] proposed three classes that found broad acceptance: ectogenic, crenogenic, and biogenic meromixis. Walker and Likens [1975] assigned about 200 meromictic lakes into various classes. Other authors [e.g., Lemmin, 1995] also claimed a separate class for morphogenic meromixis, which referred to the importance of basin size and shape for sustaining meromixis. Often, monimolimnia are found in well-defined depressions in a lake bed, which are only marginally impacted by lake-wide currents of water above. Other lakes have great relative depths, which Berger [1955] already realized favor meromixis. A recent

overview of various classification schemes is given by Hakala [2004].

[41] In this contribution, we do not try to promote a new classification scheme for meromictic lakes. It is obvious that several processes can create and sustain meromixis. We discuss them in sections 3.1–3.3, well aware of the fact that a close enough investigation would reveal more processes involved in forming and sustaining stratification. In conclusion, some lakes are difficult to classify. We looked for examples that represented particular processes as clearly as possible.

[42] Lakes do not necessarily show the same circulation pattern every year. A dimictic lake, which does not freeze in a particularly warm winter, may experience a monomictic year, and a usually monomictic lake may be dimictic in a particularly cold winter. An extreme storm event may destroy a monimolimnion. Depending on conditions, a lake may turn meromictic again. In some cases, the circulation patterns of lakes have changed permanently. After dropping the water level, the Dead Sea (Israel, Jordan) has been holomictic since 1979 (see section 4), Mono Lake (California, United States) turned meromictic for several years by reconnecting the freshwater inflow (see section 3.1), and Lake Ikeda (Kyushu, Japan) turned meromictic in 1985 after a series of colder winters (H. Kikukawa, personal communication, 2005), to name just a few well-known examples.

### 3.1. Ectogenic and Crenogenic Meromixis

[43] Ectogenically meromictic lakes have become meromictic through inflowing saline surface water. Hutchinson [1957] refers to Hemmelsdorfer See, near Kiel in Germany, which was flooded by a storm from the Baltic in 1872 and remained meromictic until the 1930s. Other deep lakes have become meromictic by inflowing salt water from deicing roads (Schalkenmehrener Maar [see Scharf and Oehms, 1992]) or lost meromictic stability by inflow of road salt to the epilimnion [Kjensmo, 1997].

[44] The class of ectogenically meromictic lakes also includes saline lakes with a freshwater inflow [Walker and Likens, 1975]. After having been deprived of its tributaries for decades, Mono Lake, California, turned meromictic by reestablishing the natural hydrologic connection with freshwater runoff entering the lake [Jellison et al., 1998]. The concentration difference sufficed to resist deep recirculation for 6 years. The most spectacular example of an ectogenically meromictic lake is probably Island Copper Mine Lake (Vancouver Island, British Columbia, Canada). After decommissioning, a 330 m deep ore mine was filled with ocean water and capped with a 7 m thick freshwater layer [Fisher and Lawrence, 2006] and hence designed to be meromictic to confine undesirable substances to the deep waters for further treatment (other examples are given by Hamblin et al. [1999], Stevens and Lawrence [1997], Stottmeister et al. [1998], and Stottmeister and Weißbrodt [2000]).

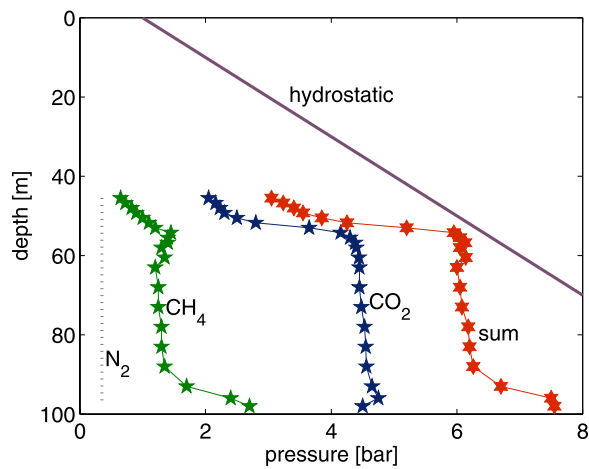
[45] We also include lakes close to the coast which receive seawater at spring tide or seasonally but otherwise have freshwater throughflow (e.g., Lower Mystic Lake in

the United States [Ludlam and Duval, 2001] or lakes in Tasmania [King and Tyler, 1981]). The episodic recharge of both freshwater and salt water sustains the stratification. Moreover, fjords were separated from the sea by falling sea level or rising terrain but still contain salt water in the deep layers [e.g., Ström, 1963]. Powell Lake in British Columbia, Canada, is a beautiful example. Sanderson et al. [1986] showed that the salt in its deep layer has been confined there for at least 10,000 years.

[46] Gibson [1999] listed 34 permanently stratified water bodies in Vestfold Hills, Antarctica. When freezing, brine exclusion raised the salinity of the water body below the ice. Salinity increased continuously and eroded progressively deeper into the stratified water body below. Maximum recirculation depth was reached at a time when the lakes were covered with ice. After ice melt, a layer of relatively fresh water protected the stratification below and inhibited deep mixing. This mechanism was also present in saline lakes in Saskatchewan, Canada. Rawson and Moore [1944] had already reported that in some of these highly saline lakes, sodium sulphate was precipitated at low water temperatures and by brine exclusion while freezing. The crystals settled to the lake bed and redissolved when temperatures rose during summer. Hammer [1994] claimed that the precipitation of sodium sulphate contributed to the stability of the meromixis in several saline lakes (e.g., Waldsea and Deadmoose in Saskatchewan).

[47] Crenogenically meromictic lakes owe their permanent stratification to groundwater inflows. The above mentioned lakes Rassnitzer See and Wallendorfer See, northern Germany (Figure 10) [Böhrer et al., 1998; Heidenreich et al., 1999], receive fresh water from a shallow aquifer, while a deep aquifer carries highly saline water from salt deposits in the deeper underground into the lake. Also, in Lago Cadagno in the Swiss Alps [Del Don et al., 2001], permanent stratification is a consequence of incoming groundwater, while in Kongressvatn, Spitsbergen, Norway, the monimolimnion is fed by mineral spring water [Bøyum and Kjensmo, 1970]. Moncur et al. [2006] present the case of a small Canadian shield lake in Manitoba (Camp Lake) which became meromictic by inflowing groundwater that had been contaminated with acid rock drainage. Lake Monoun, Lake Nyos, and Lake Kivu, central Africa [Lorke et al., 2004], can be assigned to this class, as they are located in volcanically active regions where vents carry substances into the deep waters of the lake where they dissolve and keep density stratification stable. Lac Pavin, in France, has a groundwater component to its meromictic [Aeschbach-Hertig et al., 1999, 2002] behavior, but more recent work about the distribution indicates that endogenic (see section 3.2) processes may also contribute [Bonhomme, 2008].

[48] The extent to which gases can be dissolved in a lake is displayed in Figure 11 for Lake Monoun. Depending on the chemical specification, dissolved gases contribute positively or negatively to the density of the water (see section 5.4). In Lake Nyos, the contribution of CO<sub>2</sub> is decisive for the stability of the density stratification [Schmid



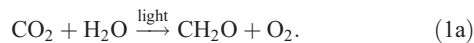
**Figure 11.** Profiles of partial pressures of dissolved gases in deep water of Lake Monoun, Cameroon, in direct comparison with hydrostatic pressure (solid line) (adapted from *Halbwachs et al.* [2004]).

*et al.*, 2004]. If more gas were advected so that the sum of the partial pressures grew higher than the hydrostatic pressure, bubbles could be created. If these bubbles form a chimney, the lake releases the dissolved gas in a limnic eruption, as happened in Lake Nyos in 1986.

### 3.2. Endogenic or Biogenic Meromixis

[49] Biogenic meromixis originates from decomposition of organic material in deep water of a lake and dissolution of its end products. *Walker and Likens* [1975] used the term endogenic meromixis, which also allows for the inclusion of geochemical processes that are not controlled by biological activity and carbon cycle. In the epilimnion, photosynthetically active plankton uses the incoming solar radiation for primary production. In addition, organic material is carried into the system through the surface or by inflowing streams (allochthonous material). A portion of this organic material settles on the lake bed. Its decomposition is facilitated by the presence of oxygen or other oxidizing agents in the deep layers of the lake (Figure 12). Equations (1a), (1b), and (2a)–(2g) show the corresponding reactions where  $\text{CH}_2\text{O}$  is used for organic substance for simplicity.

#### Photosynthesis



as performed by green algae and cyanobacteria. A similar equation is valid for the formation of organic matter by photosynthetic sulphur bacteria [*van Niel*, 1936; *Schwoerbel*, 1999; *Wetzel*, 2001]. Besides oxygen and sulphur, other substances can also act as electron/hydrogen carriers:



Decomposition of organic material is accomplished by the following processes:

#### Respiration



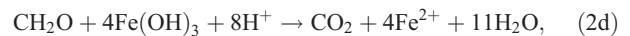
#### Denitrification



#### Manganese reduction



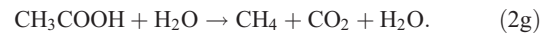
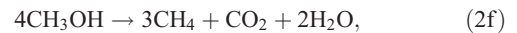
#### Iron reduction



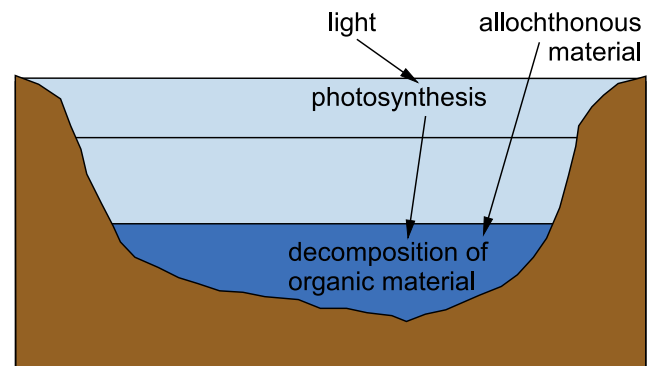
#### Sulphate reduction



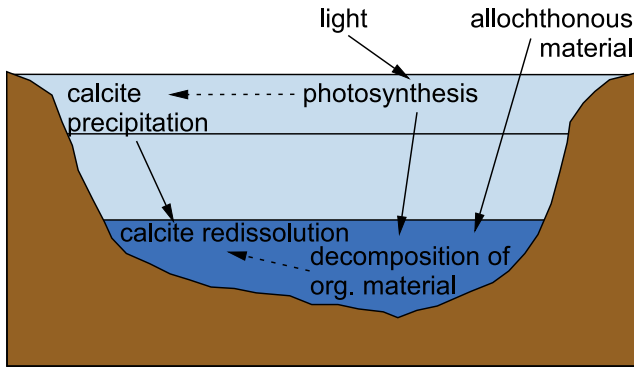
#### Methanogenesis



Equations (2a)–(2g) are gross summaries for initial and end reagents of several microbially mediated intermittent steps. More details are available in the appropriate studies [e.g., *Schlesinger*, 2005; *Wetzel*, 2001]. End products dissolve in deep lake water and contribute their part to density. The more productive a lake is, the more important this process can be for stabilizing stratification. Some of the maar lakes in Germany are typical examples for biogenic meromixis [*Scharf and Oehms*, 1992; *Stewart and Hollan*, 1975]. *Findenegg* [1935] argues that the meromictic lakes of Carinthia are permanently stratified because of dissolution of end products from decomposing organic material. Presented  $\text{CO}_2$  profiles and pH profiles with high  $\text{CO}_2$  concentrations and low pH in the monimolimnion support his arguments.

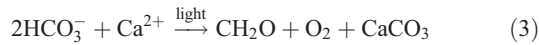


**Figure 12.** Sketch of a biogenically meromictic lake. Organic material is assembled from photosynthesis in the epilimnion; it can precipitate and decompose and dissolve in the monimolimnion.

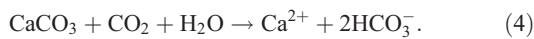


**Figure 13.** Calcite precipitation by photosynthetically increasing pH in the epilimnion and redissolution in the monimolimnion by carbon dioxide from decomposition of organic material.

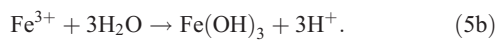
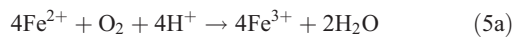
[50] Since the study by *Hutchinson* [1957], some more geochemical cycles have been shown to create or sustain meromixis [*Walker and Likens*, 1975; *Hongve*, 2004]. *Rodrigo et al.* [2001] refer to high Ca concentrations in the monimolimnion to prove that the monimolimnion in Lake La Cruz, Spain, is stabilized by calcite precipitation in the epilimnion and partial dissolution in the monimolimnion. In this geochemical cycle, calcite is precipitated out of the epilimnion by pH shift due to photosynthetic activity, while in the monimolimnion carbon dioxide is formed by the decomposition of organic material which facilitates the solution of calcite in monimolimnetic waters (see Figure 13 and equations (2a)–(2d), (3), and (4)). Photosynthesis and calcite precipitation



Calcite dissolution through carbon dioxide

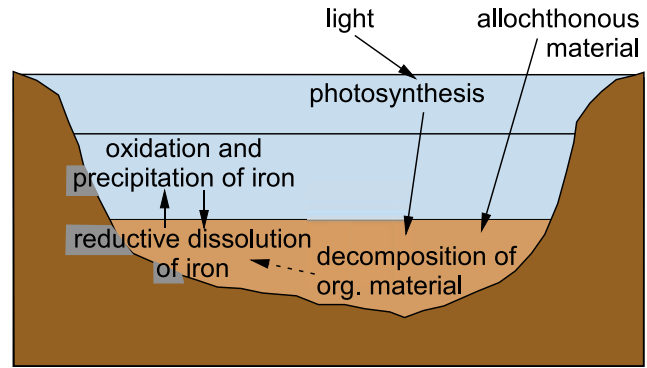


A considerable number of meromictic lakes are stabilized by iron cycling. Inflowing groundwater advects dissolved ferrous iron. This is oxidized to ferric iron and precipitated



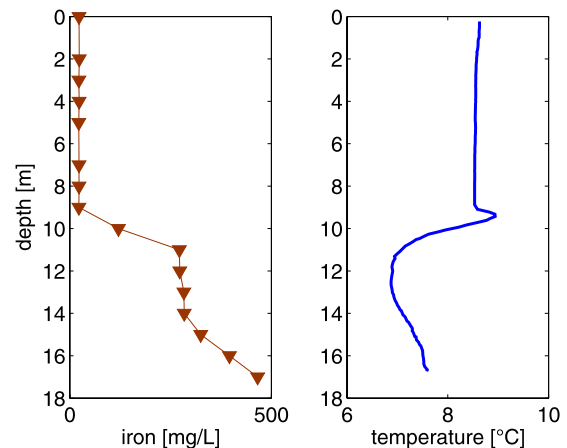
Under favorable conditions a portion of this iron can be reduced to ferrous iron again and become dissolved in monimolimnetic waters (Figure 14), where it contributes to density and stabilizes the stratification. Both oxidation and reduction are facilitated by the presence of microbial organisms. Reduction of iron was already shown in equation (2d).

[51] *Kjensmo* [1967, 1968] reported the case of an anoxic monimolimnion with high iron concentrations beneath an oxic mixolimnion. *Merilainen* [1970] has found the same layering in lakes of Finland, and *Campbell and Torgersen*



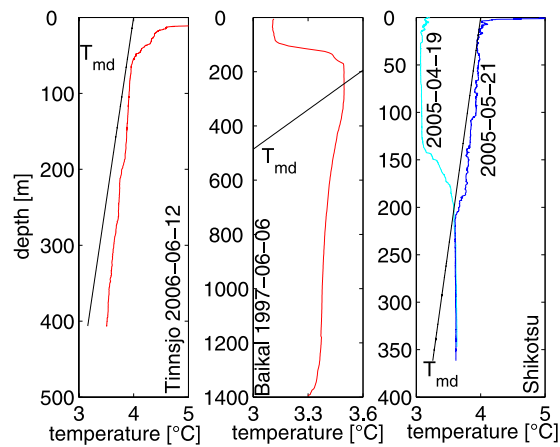
**Figure 14.** Oxidation of ferrous iron to ferric iron and subsequent precipitation in oxic layers of the lake. In the monimolimnion, reduction to ferrous iron is facilitated by using organic material as reduction agent. Diffusion across the chemocline and mixing of mixolimnetic waters with monimolimnetic waters is inefficient in terms of transport of iron, as iron is oxidized and precipitated out of the oxic layer.

[1980] saw the same feature in lakes in Ontario, Canada. More recent work by *Hongve* [1997, 2002] confirms this stratification in lakes of southern Norway. He also states that the contribution of dissolved  $\text{CO}_2$  plays an important role for stability. Most importantly, this iron cycling can be observed in many mining lakes, e.g., in eastern Germany (Waldsee near Döbern [see *Schimmele and Herzsprung*, 2000], Moritzteich [see *Stellmacher*, 2004] (Figure 15), Mining Lake 111 [see *Karakas et al.*, 2003], and Goitschee [see *Boehrer et al.*, 2003]). In Berkeley Pit Lake (Butte, Montana), even pyrite is oxidized by ferric iron in the monimolimnion, liberating additional ferrous iron, sulphate, and acidity [*Pellicori et al.*, 2005]. Reactions (5a) and (5b) release protons in sequence. A formation of local pH minima can be the consequence in the contact zones between oxic and anoxic layers, as has been observed in a number of lakes (e.g., Lake Goitsche [see *Boehrer et al.*, 2003]).



**Figure 15.** Temperature profile and concentration of iron in Moritzteich, Germany, on 28 October 2003. Oxic mixolimnion is 0–10 m, and anoxic monimolimnion is below 10 m (data from *Stellmacher* [2004]).





**Figure 16.** Temperature profiles of thermobarically stratified lakes: (left) Tinnssjø, Norway, during (early) summer stratification above temperature of maximum density  $T_{md}$ ; (middle) Lake Baikal, Siberia, Russia, during (late) winter stratification with vertical transition through  $T_{md}$ ; (right) Lake Shikotsu, Hokkaido, Japan, with nearly isothermal deep water body below the  $T_{md}$  transition. Tinnssjø data from B. Boehrer (unpublished data, 2008); Lake Baikal data from Wüest *et al.* [2005], (with permission of the American Society of Limnology and Oceanography, Inc.); and Lake Shikotsu data from B. Boehrer *et al.* (Stratification in very deep, purely thermally stratified lakes, manuscript in preparation, 2008).

[52] The same cycle between oxic and anoxic zones can be accomplished by manganese. In a few lakes, manganese can take the leading role, e.g., Lake Nordbytjernet in southern Norway [Hongve, 1997]. In most cases of observed enrichment with manganese in the monimolimnion (e.g., Berkeley Pit Lake [Pellicori *et al.*, 2005] and Lake Goitsche, Rassnitzer See, and Wallendorfer See (M. Schultze, unpublished data, 2003; B. Boehrer, unpublished data, 2003)) and in deep depressions of the Baltic Sea [Pohl *et al.*, 2004], the manganese only contributes a small portion to the density difference between mixolimnion and monimolimnion, while either salinity or iron and bicarbonate concentration is responsible for sustaining meromixis.

[53] Beyond the contribution to density stratification, cycling of iron and manganese plays an important role for the enrichment of trace elements in the monimolimnion. These substances often form coprecipitates with iron or manganese or are adsorbed by solid oxidized species of iron and manganese and hence are transferred by sedimentation into the monimolimnion. Reductive dissolution of solid oxidized species of iron and manganese results in the liberation of trace substances into monimolimnetic water. An example is given by Hongve [1997] for phosphorus.

[54] Some of the processes discussed in connection with biogenic meromixis cannot be present at the same time. If, at the base of a lake, sulphate is reduced in considerable quantities and thus high concentrations of  $H_2S$  are present (see equation (2e)), ferrous iron will be precipitated as iron sulphide and iron will continuously be removed from the system. Under such conditions, the iron cycle cannot be

responsible for forming a monimolimnion. In addition, if sulphate reduction occurs at the same time and at a similar rate as iron reduction, the formation of a microstratification of sulphur bacteria in the chemocline, as known from Sorokin [1970] and Tonolla *et al.* [2003], will be suppressed. Similarly, precipitation of iron sulphide will remove  $H_2S$ , which is required by autotrophic sulphur bacteria for photosynthesis.

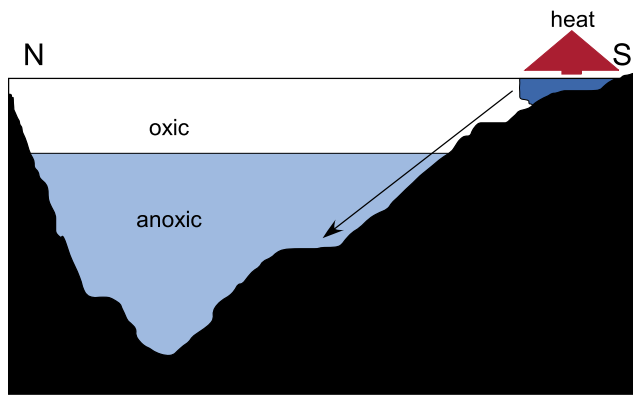
### 3.3. Thermobaric Stratification

[55] We encounter one additional barrier for the deep water recirculation in very deep lakes that show low concentrations of dissolved substances and that traverse the temperature of maximum density ( $4^\circ C$ ) during the annual cycle in their surface waters. Deep water of these lakes is close to the temperature of maximum density  $T_{md}$ . The barrier for deep water recirculation is imposed by the (small) compressibility of water and its temperature dependence: cold water is more compressible than warmer water. As a consequence, temperature of maximum density  $T_{md}$  decreases as pressure, i.e., depth, increases (by about 0.21 K over 100 m water depth [Eklund, 1963], see also equation (26) in section 5).

[56]  $T_{md}$  can be displayed as profile versus depth. To the right of the  $T_{md}$  line, usual temperature profiles with rising temperatures toward the surface can be found, while to its left, stable temperature stratification is inverse. Particularly interesting are cases where surface waters cross the temperature of maximum density during spring and autumn. Both summer and winter profiles then show peculiarities.

[57] Starting at a surface temperature above  $4^\circ C$  in summer, falling temperatures with increasing depth imply a stable stratification, as long as temperatures lie above  $T_{md}$  of corresponding pressure. At greater depth, temperatures can well be lower than  $4^\circ C$  (Figure 16, left). The first profile in a purely temperature stratified freshwater lake with a transition through  $4^\circ C$  was measured in Lake Mjøsa, Norway, on 23 June 1883 cited by Strøm [1945]. Similar temperature profiles were also found in Lake Ladoga, Russia, in 1900 by Pettersson [1902] (as cited by Strøm [1945]), in Crater Lake, United States, in 1913 by Kemmerer *et al.* [1924] (as cited by Strøm [1945]), and in Lake Shikotsu, Japan, by Yoshimura [1936a, 1936b], proving the presence of thermobaric stratification in freshwater lakes if climatic and morphometric conditions allowed for it.

[58] In winter, when surface temperatures are forced below  $4^\circ C$ , only inverse temperature stratification is stable just below the surface. At greater depths, however, water may show temperatures between  $4^\circ C$  and  $T_{md}$  at the respective pressure. Under these conditions, profiles must intersect with the  $T_{md}$  profile. Conditions are stable only for a vertical transition, i.e., no temperature gradient [Eklund, 1963, 1965]. In fact, Lake Baikal shows a temperature maximum at the intersection (see Figure 16, middle). In Lake Shikotsu (Hokkaido, Japan, Figure 16, right) the temperature gradient vanishes from the intersection downward (see also Crater Lake [Crawford and Collier, 1997, 2007]).



**Figure 17.** Sketch of deep water renewal in Lake Malawi. At the southern end “S,” surface water becomes cold enough during cold periods in southern winter (shaded dark) that its density increases sufficiently to sink and intrude into the anoxic monimolimnion.

[59] In winter, while waters warmer than  $T_{md}$  can be conserved at great depth, all waters colder than  $T_{md}$  will be removed when the surface water becomes warmer. As a consequence, some of the deep crater lakes reflect the  $T_{md}$  profile over a wide range of depths. Many of the sufficiently deep lakes in Norway, however, show a temperature profile between isothermal and the  $T_{md}$  profile. *Eklund* [1965] reasons that the lake stratification remains close to maximum stability, which he has found at a temperature profile of half the  $T_{md}$  gradient.

[60] Observations show that the water bodies of these lakes (Crater Lake, Lake Shikotsu, Lake Baikal, Lake Tinnsjø) are well supplied with oxygen. This can in part be attributed to low temperatures, as productivity and depletion of oxygen happen at a slow rate. However, it is also an indication that a considerable portion of bottom water is replaced each year and turbulent diffusive transports are highly effective because of the very weak density gradient. As a consequence, chemical gradients do not appear in these lakes, and (most) scientists have refrained from calling these lakes meromictic, despite their permanent stratification.

#### 4. EPISODIC PARTIAL DEEP WATER RENEWAL

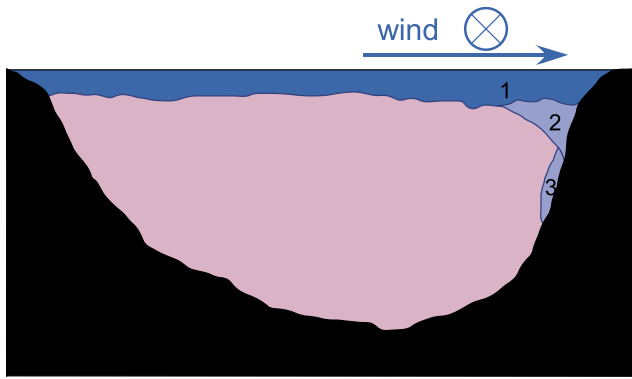
[61] A number of lakes do not fit well into the classification of holomixis and meromixis. These lakes do not experience a complete overturn; thus they are not holomictic, but neither does their circulation fit into the picture of meromictic lakes. While in meromictic lakes the circulation is confined to the mixolimnion and the water body is often homogenized from the surface to the chemocline over the entire horizontal dimension, lakes with episodic deep water renewal form water parcels of monimolimnetic density within the mixolimnion. These water parcels manage to proceed through the surrounding mixolimnetic waters down into the monimolimnion. In general, this process is similar to the ocean circulation, and hence it is not surprising that some of the largest lakes show a similar behavior.

[62] From tracer measurements (chlorofluorocarbon (CFC), helium-tritium, sulphur hexafluoride ( $\text{SF}_6$ )), an age, i.e., time since the last gas exchange with the atmosphere took place, is inferred which often lies in the range of one to a few decades. Monimolimnetic waters representing a mixture of waters of different age are assigned the apparent bulk age. In addition, tracer gases are also transported by smaller-scale turbulent transport processes through the stratified water column. This contribution also needs separate consideration when estimating deep water renewal from tracer gas data [e.g., *Ravens et al.*, 2000; *Weiss et al.*, 1991].

[63] A famous and easy to follow example for deep water renewal is 703 m deep Lake Malawi (or Nyasa) in East Africa. Lake Malawi is usually referred to as a meromictic lake, as it is permanently stratified and a clear chemocline delimits the oxic mixolimnion from the anoxic monimolimnion at a depth of 200 m. The age of monimolimnetic water has been measured with CFC as 20 to 25 years [*Vollmer et al.*, 2002a]. Lake Malawi extends from latitudes  $9^\circ\text{S}$  to  $15^\circ\text{S}$ . Thus, at least in the southern winter, it is exposed to a climatic gradient. It is suspected that the lowest water temperatures are created at the southern end during southern winter (Figure 17). Water can become cold enough to exceed the density of the surrounding mixolimnion and to flow down side slopes to the abyss. In each episode, only part of the monimolimnion is replaced [e.g., *Halfman*, 1993]. As part of the tracer gases were carried by turbulent transport through the stratified water column, the water age of 20 to 25 years gives an upper bound for the amount of annually renewed deep water.

[64] The deep water circulation in moderately saline ( $\sim 6 \text{ g kg}^{-1}$ ) Issyk-Kul (Kyrgyzstan, central Asia [*Peeters et al.*, 2003]) is similar. From the deepest point (668 m) up to 100 m depth, deep water shows a temperature between  $4.3$  and  $4.8^\circ\text{C}$ . In winter, cold winds descending from the Tien Shan (Mountains) traverse the lake from east to west. In shallow areas of the lake, water temperature falls enough to form dense water, which enters submerged river valleys to proceed into the abyss of the lake and to recharge deep waters. Tracer measurements indicate a deep water age of about 10 years [*Hofer et al.*, 2002; *Vollmer et al.*, 2002b; *Peeters et al.*, 2003]. Though the water age is only a factor of 2 or 3 smaller than in Lake Malawi, Issyk-Kul is not called meromictic, as a pronounced chemocline to a chemically different monimolimnion is missing.

[65] Deep water renewal also takes place in lakes that are commonly called holomictic/oligomictic. *Hollan* [1998, 1999], for example, argues that in years of no complete overturn, shallow areas of Lake Constance, Germany, could become cold enough to form intrusions that proceed into the deepest parts of the lake (250 m). In exceptionally cold winters, Lake Constance may stratify inversely [1987] or even freeze [1963] [*Zenger et al.*, 1990; *Bäuerle et al.*, 1998]. However, a more recent study by *Rossknecht* [2003] found that the renewal of the deepest waters was not particularly efficient during those seemingly dimictic years. This is probably connected to remaining thermobaric strat-



**Figure 18.** Sketch of deep water renewal in a deep and long lake (e.g., Lake Baikal) by wind stress and consecutive convective propagation of a dense intrusion (1, epilimnion without strong wind stress; 2, downwelling forced by wind stress on the epilimnion; 3, convective propagation after crossing temperature of maximum density). Longshore wind and onshore wind represent contribution of Coriolis deflection and direct surface draft for accumulation of surface water at the shore.

ification during inverse temperature stratification in very deep lakes (see section 3.3).

[66] Intrusions recharging monimolimnetic waters can also be driven by a higher salt concentration. Before 1979, evaporation created water of higher salt concentration in the shallow southern basin of the highly saline Dead Sea. Because of its very high density, it could flow underneath the mixolimnetic waters of the main basin [Nissenbaum, 1969; Gat, 1995]. This process kept the Dead Sea stratified for decades, perhaps centuries. However, since the southern basin has basically dried up and does not connect to the main basin as before, the lake has been holomictic since 1979. In addition, capping with fresh water from the Jordan River (ectogenic meromixis) has weakened over the same period.

[67] High-salinity waters can also be formed by freezing. The liquid phase remains with increased salinity and thus can gain enough density to form intrusions for recharging a monimolimnion [Goldman *et al.*, 1972]. Walker and Likens [1975] have listed five lakes in Antarctica which stay permanently stratified by this process [also Gibson, 1999] and proposed a separate meromixis class, as they felt that these lakes do not fit into any of the three meromixis classes of Hutchinson [1957].

[68] Lake Baikal fulfils the conditions for a classic thermobaric stratification (see section 3.3). Late during the winter, surface waters are below 4°C and “inversely” stratified to a depth of 250 m, where the temperature profile intersects vertically with the  $T_{md}$  line. However, below the intersection, temperatures fall below 3.3°C toward the bottom of the southern basin of the lake (Figure 16, middle) [Wüest *et al.*, 2005]. Hence a process recharging the deep water from the surface can be suspected. Most probably during winter periods when the lake is not covered with ice, wind pushes cold surface water deep enough that its density is higher than surrounding water (Figure 18, “2”). From there, it can proceed convectively to greater depth (Figure 18, “3”)

while it mixes with surrounding water (thermobaric instability [Carmack and Weiss, 1991]). The exact interplay of all processes, however, has not yet been deciphered. CFC and helium-tritium measurements indicate a monimolimnion age of less than 20 years [Weiss *et al.*, 1991; Hohmann *et al.*, 1998; Kodenev, 2001], where turbulent diffusive transport vertically through the stratification also contributes its part.

## 5. QUANTITATIVE APPROXIMATIONS FOR STRATIFICATION RELEVANT QUANTITIES

[69] For evaluation of stratification related magnitudes from field measurements, a number of approaches have been documented. At low salinities, the best reference is Chen and Millero [1986], while for ocean conditions the valid reference is Fofonoff and Millard [1983]. From both studies, we cite most often used approximations. However, both studies include many more physical quantities related to density, which cannot be shown within this review. Instead, we present ways to approach stratification relevant magnitudes if both ocean assumptions and freshwater assumptions fail.

### 5.1. Depth and Sound Speed

#### 5.1.1. Depth

[70] Depth of a probe could be measured from length of a rope. However, many ropes change their length considerably under tension. Other ropes shrink when absorbing water. Steel wires or metal pipes might solve some of the problems [e.g., von Rohden and Ilmberger, 2001], but currents or drift of a vessel will curve the wire and shift a probe sideways.

[71] For this reason, depth is most commonly measured as pressure. In water, hydrostatic pressure  $p$  increases with  $z$  (taken as positive downward) as

$$\frac{dp}{dz} = g\rho_{\text{in situ}} \quad (6)$$

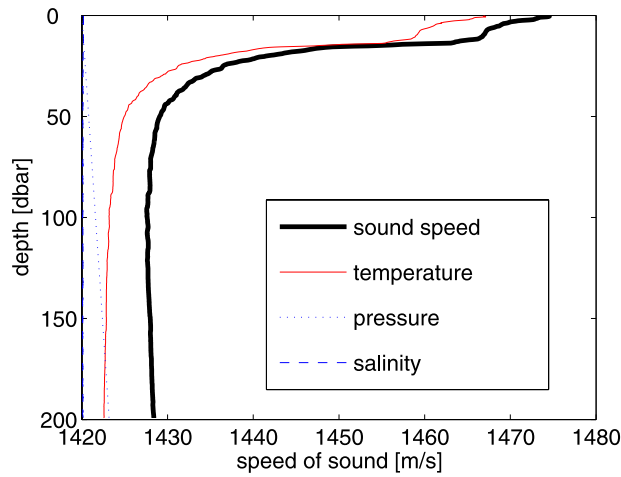
[72] Earth’s acceleration  $g \approx 9.8 \text{ m s}^{-2}$  can be assumed to be constant in nearly all lakes. In situ density  $\rho_{\text{in situ}}$  in freshwater lakes differs from  $1000 \text{ kg m}^{-3}$  as a consequence of higher temperature, dissolved substances, and pressure but only in the range of a few permille. Many probes therefore indicate pressure in dbar, which is close to m. In the ocean and in lakes of high salinity, depth needs to be evaluated properly following equation (6), which normally is facilitated by the measurements in the profile above. In all cases, atmospheric pressure, which varies with altitude and time, needs to be eliminated (“calibration”).

#### 5.1.2. Sound Speed

[73] Depth can also be measured from sound speed. An echo sounder measures the travel time  $t$  for a sound ping to complete the path of length  $s$

$$t = \int_{\text{sender}}^{\text{receiver}} \frac{1}{c} ds, \quad (7)$$





**Figure 19.** Profile of measured sound speed in Lake Constance on 13 June 2002 at a 200 m deep location. Profiles of temperature, pressure, and salinity (barely visible on the left) in roughly the magnitude of their contribution to sound speed.

where sound speed  $c$  can be calculated from density and adiabatic compressibility  $\kappa$ :

$$c^2 = 1/(\rho_{\text{in situ}} \kappa). \quad (8)$$

[74] Hence the sound propagation is dependent on temperature, salinity, and pressure (see Figure 19). In addition, it is also influenced by currents, but in most lakes this effect is a subordinate contribution and will not be discussed here. The temperature dependence in pure water was evaluated by *Del Grosso and Mader* [1972], and the approximation was extended for pressure and salinity by *Chen and Millero* [1986]:

$$c = \sum_{i=0}^5 a_i T_{\text{pot}}^i + S \sum_{i=0}^2 b_i T_{\text{pot}}^i + p \sum_{i=0}^2 c_i T_{\text{pot}}^i - 5.58 \times 10^{-5} S p + 1.593 \times 10^{-5} p^2, \quad (9)$$

with  $a_i = [1402.388; 5.0371; -5.8085 \times 10^{-2}; 3.342 \times 10^{-4}; -1.478 \times 10^{-6}; 3.146 \times 10^{-9}]$ ,  $b_i = [1.322; -7.01 \times 10^{-3}; 4.9 \times 10^{-5}]$ , and  $c_i = [0.15564; 4.046 \times 10^{-4}; -8.15 \times 10^{-7}]$ .

[75] Coefficients have been listed without units for better legibility. The unit for pressure  $p$  is bar, and for salinity it is psu. *Chen and Millero* [1986] use a slightly different definition for salinity in the ocean (see section 5.3.2) and salinity in this formula,  $S = 1.00488 S_{\text{ocean}}$  (for more about salinity, see section 5.3.2).

[76] Sound speed can vary by several percent within one lake mainly because of the temperature variation. Hence this must be considered for accurate measurements. Sound beams propagating in horizontal directions within vertical temperature and pressure gradients are deflected from straight lines and thus can take unexpected paths from sender to receiver. Depending on conditions, several different paths may be possible [e.g., *Dushaw et al.*, 2001].

[77] On the contrary, positioned senders and receivers can record the variations in the travel time for a fixed distance, and variations in the temperature field can be measured. A symmetrical setup may facilitate measurements of current speeds. A field of sensors can provide data for tomographic procedures, which can evaluate three-dimensional temperature fields and currents by noninvasive methods [e.g., *Dushaw et al.*, 2001].

## 5.2. Temperature

[78] Temperatures recorded in lakes by field equipment are so-called in situ temperatures. Without any further annotation, temperature data will be understood as in situ temperatures. Nearly all calculations refer to this value, as it is the physically, chemically, and ecologically relevant magnitude.

[79] However, if detailed considerations of stability and vertical temperature gradients are intended, potential temperature may be useful. This includes the effect of energy required for the expansion. Potential temperature is defined as temperature that a water parcel would acquire if transferred adiabatically to a reference pressure, often taken as atmospheric pressure [e.g., *Gill*, 1982]. Derivation is along lines of the Maxwell identities (see Joule-Thomson effect [e.g., *Reif*, 1987]) of the statistical mechanics. Adiabatic lapse rate is given by

$$\left(\frac{dT}{dz}\right)_{\text{ad}} = \frac{g\alpha(T + 273.15)}{c_p}, \quad (10)$$

where  $\alpha$  is thermal expansion coefficient and  $c_p \approx 4200 \text{ J(kg K)}^{-1}$  is specific heat at constant pressure. Hence potential temperature is evaluated as

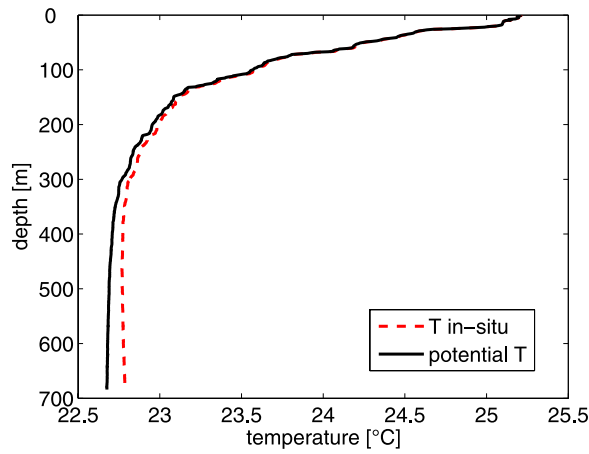
$$T_{\text{pot}} = T(z) + \int_{z'=z}^0 \left(\frac{dT}{dz}\right)_{\text{ad}} dz' = T(z) + \int_{z'=z}^0 \frac{g\alpha(T + 273.15)}{c_p} dz', \quad (11)$$

where  $z$  is positive downward. For most limnological purposes,  $\alpha$ ,  $g$ , and  $c_p$  can be considered constant over the entire depth range. Using an estimate for  $\alpha$  of

$$1 \dots 1.5 \times 10^{-5} (T - T_{\text{md}}) \frac{1}{K^2}$$

equation (10) yields an adiabatic lapse rate  $(dT/dz)_{\text{ad}}$  of the order of  $(T - T_{\text{md}}) \times 10^{-5} \text{ m}^{-1}$ . Hence in most cases the temperature stratification is much larger than the adiabatic lapse rate. The difference between in situ and potential temperature can be neglected in most cases. However, to be exact, we will add a subscript  $T_{\text{pot}}$  whenever it is appropriate in this contribution. A considerable difference between potential temperature and in situ temperature can be found in deep tropical lakes. While, for example, the in situ temperature profile in Lake Malawi is not monotonous, *Wüest et al.* [1996] showed that potential temperature rises monotonously; that is, the lake is stably density stratified in respect to temperature only. In Figure 20, we present a profile





**Figure 20.** Profiles of (in situ) temperature  $T$  and potential temperature  $T_{\text{pot}}$  near the deepest location of Lake Malawi on 13 September 1997 (data from *Vollmer et al.* [2002a]).

measured in Lake Malawi by *Vollmer et al.* [2002a]. Potential temperature was calculated following equation (11), where we assumed constant values for  $g$  and  $c_p$ . The thermal expansion coefficient

$$\alpha = \frac{1}{V} \frac{\partial V}{\partial T} = - \frac{1}{\rho_{\text{in situ}}} \frac{\partial \rho_{\text{in situ}}}{\partial T}$$

was evaluated from equation (17) (see below). The depiction is instructive in so far as it indicates what temperature is required at the surface for deep water renewal.

### 5.3. Electrical Conductivity, Electrical Conductance, and Salinity

#### 5.3.1. Electrical Conductivity and Electrical Conductance

[80] Many substances dissolve in lake water, dissociate, and form electroactive ionic species. Hence they contribute to the electrical conductivity. Electrical conductivity of a water sample depends on the dissolved constituents and temperature of the sample [e.g., *Rodhe*, 1949]. To compensate for temperature effects, electric conductivity is recorded while scanning the relevant temperature interval (Figure 21).

[81] In most cases, a linear regression  $C(T) = aT + b$  is satisfactory to define conductance, i.e., electrical conductivity  $\kappa_{\text{ref}} = C(T_{\text{ref}}) = aT_{\text{ref}} + b$  at a certain reference temperature  $T_{\text{ref}}$ . Most commonly, a reference temperature of 25°C is used. Dividing  $C(T)$  by  $\kappa_{\text{ref}}$  and solving for the latter, we find the user's formula for calculating conductance

$$\kappa_{\text{ref}} = \frac{C(T)}{\alpha_{\text{ref}}(T - T_{\text{ref}}) + 1}, \quad (12)$$

where

$$\alpha_{\text{ref}} = [T_{\text{ref}} + (b/a)]^{-1}.$$

[82] For the hypolimnion water of highly acidic Mining Lake 111 (5 m sample), *Karakas et al.* [2003] evaluated

$$b/a = 35.2$$

and a very small value  $\alpha_{\text{ref}} = 0.0166$  (see also section 5.3.1) at a reference temperature  $T_{\text{ref}} = 25^\circ\text{C}$ :

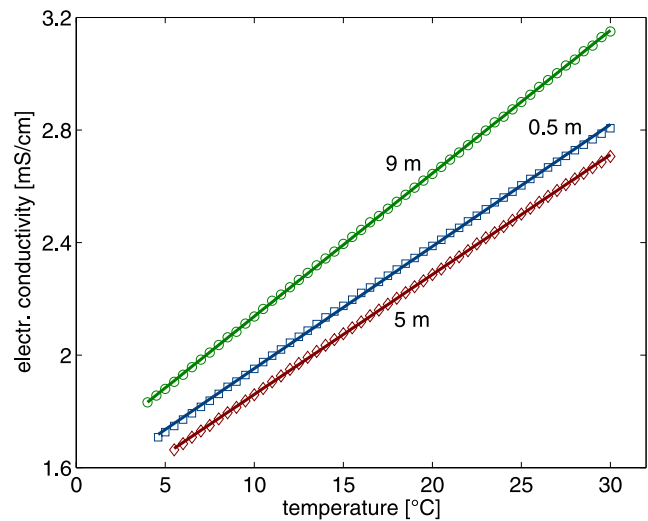
$$\kappa_{25} = \frac{C(T)}{0.0166(T - 25^\circ\text{C}) + 1}. \quad (13)$$

[83] For better clarity, we dropped the unit  $\text{K}^{-1}$ . Equation (13) was applied to CTD profiles of years 1996–2001 (see Figure 22). The result can be interpreted as an indication for concentration of dissolved substances. The depiction clearly shows the high concentration in the monimolimnion and its intermittent absence. In Mining Lake 111, the precipitation-evaporation deficit during summer months increases the conductance in the epilimnion [*Karakas et al.*, 2003]. *Heinz et al.* [1990], for example, used changes in conductance as an indication of vertical transport in Lake Constance and calculated transport coefficients from it.

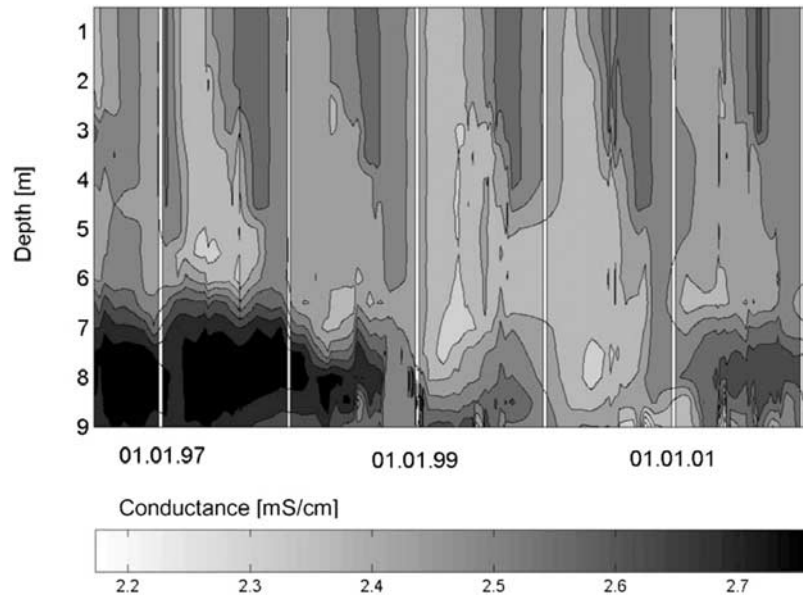
[84] The advantage of restricting oneself to a linear regression for conductance bears the advantage of only one parameter

$$(T_{\text{ref}} + (b/a))^{-1} = \alpha_{\text{ref}}$$

defining the calculation entirely. This facilitates the comparison of data of different reference temperatures. Measurements of various years would directly indicate to what extent changes between various layers and years must be considered. As in Mining Lake 111, such differences can be found, especially within other meromictic lakes. *Karakas et al.* [2003] decided to use the hypolimnion regression for the entire lake, as the epilimnetic fit turned out to be very



**Figure 21.** Electrical conductivity of three water samples of Mining Lake 111 versus temperature [see *Karakas et al.*, 2003]. Symbols represent measured conductivity. Solid lines show linear regression (with permission from Birkhäuser).



**Figure 22.** Electrical conductance  $\kappa_{25}$  in Mining Lake 111 versus depth and time (adapted from Karakas *et al.* [2003] with permission from Birkhäuser).

similar, while in the monimolimnion, gradients were very high anyway; hence no particularly high accuracy for conductance was required.

[85] In highly acidic Mining Lake 111, a very small value  $\alpha_{25} = 0.0166$  was found. In most surface water, a value close to  $\alpha_{25} = 0.02$  is appropriate [Wissenschaftlich-technische Werkstätten, 1993]. For the neutral mixolimnion of lake Goitsche,  $\alpha_{25} = 0.0197$  was evaluated, while epilimnion, hypolimnion, and monimolimnion of salt lake Ransitzsee showed values  $0.0191 < \alpha_{25} < 0.0199$ . In acidic Roter See near Burgkennitz, Germany,  $\alpha_{25} = 0.0185$ , while several samples of the Elbe river near Magdeburg, Germany, showed  $0.0196 < \alpha_{25} < 0.0208$  (G. Götz, personal communication, 2000).

[86] This approach follows the *International Organization for Standardization (ISO)* [1985], in that only the factor  $\alpha_{25}$  is adjusted for varying composition of constituents contributing to electrical conductivity. A close look at Figure 21 reveals that data points show a curvature against temperature, which cannot be reproduced by a linear regression. Hence a better parameterization has been looked for. Sorensen and Glass [1987] consider various approaches for fitting temperature dependence of electrical conductivity. They conclude that the ISO [1985] delivers good results, compared to other methods. However, they also prove that an approximation of the form

$$\kappa_{25} = C(T) \left( \frac{\eta(T)}{\eta_{25}} \right)^K, \quad (14)$$

would be better. The term  $\eta(T)/\eta_{25}$  represents the ratio of viscosity of water at the temperature of measurement and the reference temperature at 25°C. Equation (14) requires only one constant  $K$ , which ISO supplies in a table for the most common ions. One shortcoming of equation (14) lies in the fact that  $\eta(T)/\eta_{25}$  is not necessarily available, at least not for the considered lake water [e.g., Jellison *et al.*, 1999].

Most scientists still use a temperature compensation following ISO [1985] or variations thereof.

[87] If the lake water properties cannot be evaluated in a conductivity measurement versus temperature, coefficients for the most common ions can be taken from Sorensen and Glass [1987]. The coefficients of most of the common ions are close to each other; only  $H^+$  shows a different behavior. Consequently, a good approximation is gained by replacing  $K$  in equation (14) by  $\bar{K} = 0.941 - 0.00639[H^+]$ , where  $H^+$  is given in  $\mu\text{mol L}^{-1}$ . Their data relied on Canadian lakes with low concentration of other dissolved salts. How well this regression would suit highly mineralized waters has not been checked.

[88] From oceanography it is known that electrical conductivity is also influenced by pressure. Wüest *et al.* [1996] assume that the electrical conductivity is increased by a factor of 1.0066 at 700 m depth in Lake Malawi in comparison to surface pressure by extrapolating ocean assumption to zero salinity. While in Lake Malawi this effect makes up for nearly the entire conductivity gradient in the monimolimnion, this effect is negligible in most other lakes.

### 5.3.2. Total Dissolved Substances and Salinity

[89] For various purposes, depiction and numerical evaluation of the quantity of total dissolved substances (TDS) may be useful. In some applications, this value is used as an intermediate step to calculating density in limnic waters [Chen and Millero, 1986; McManus *et al.*, 1992]. Evaluating TDS involves the full chemical analysis of all dissolved substances including gases and organic material, both of which can pose sampling and analytical difficulty.

[90] The easiest approach connects TDS with conductance by a regression factor [e.g., Snoeyink and Jenkins, 1980] (as cited by McManus *et al.* [1992]). But there are more elaborate approaches as given by higher-order approx-

imations of *McManus et al.* [1992]. If the nonconductive dissolved species do not show a parallel profile to conductive species, their contribution must be added in separate (e.g., *Wüest et al.* [1996] in their approach for Lake Malawi).

[91] Oceanography uses electrical conductivity and temperature to calculate salinity in psu, which gives a good indication for dissolved salt in  $\text{g kg}^{-1}$  for ocean water and brackish water:

$$S = S(C, T, p). \quad (15a)$$

Salinity is evaluated over several steps:

$$R = C/42.914 \text{ mS cm}^{-1}, \quad (15b)$$

$$r_t = \sum_{i=0}^4 c_i T_{\text{pot}}^i, \quad (15c)$$

$$R_p = 1 + \sum_{i=0}^3 e_i p^i / \left[ (d_0 + d_1 T_{\text{pot}} + d_2 T_{\text{pot}}^2 + (d_3 + d_4 T_{\text{pot}}) R) \right], \quad (15d)$$

$$R_t = R / (R_p r_t), \quad (15e)$$

and

$$S_{\text{ocean}} = \sum_{i=0}^5 a_i R_t^{i/2} + \frac{T_{\text{pot}} - 15}{1 + k(T_{\text{pot}} - 15)} \sum_{i=0}^5 b_i R_t^{i/2}. \quad (15f)$$

Coefficients have been evaluated at high accuracy:

$$\begin{aligned} a_i &= [0.0080; -0.1692; 25.3851; 14.0941; -7.0261; 2.7081], \\ b_i &= [0.0005; -0.0056; -0.0066; -0.0375; 0.0636; -0.0144], \\ c_i &= [0.6766097; 2.00564 \times 10^{-2}; 1.104259 \times 10^{-4}; -6.9698 \\ &\quad \times 10^{-7}; 1.0031 \times 10^{-9}], \\ d_i &= [1; 3.426 \times 10^{-2}; 4.464 \times 10^{-4}; 0.4215; -3.107 \times 10^{-3}], \\ e_i &= [0; 2.070 \times 10^{-5}; -6.370 \times 10^{-10}; 3.989 \times 10^{-15}], \\ k &= 0.0162. \end{aligned}$$

[92] At the time of writing this review, UNESCO offered an online calculator (<http://ioc.unesco.org/>) for properties of ocean water using coefficients from *Fofonoff* [1985]. For low-salinity limnetic water (<0.6 psu), *Chen and Millero* [1986] introduced a correction factor to evaluate salinity for weakly conductive limnetic waters:

$$S = 1.00488 S_{\text{ocean}}. \quad (15g)$$

[93] In limnetic systems, the composition of dissolved substances differs from the ocean. In some cases, there are

even pronounced vertical gradients within the same lake. Also, there may be substances contributing considerably to density but not to electrical conductivity (e.g., suspended particles, dissolved organic compounds, silicate). As a consequence, salinity can only be used with reservation and when chemical composition of lake water is close enough to ocean conditions. Annotation of the psu or referring to salinity according to UNESCO indicates reference to a numerical value and not salt content.

[94] Oceanography has long been aware of limitations of current definitions of salinity and psu scale. Measurements during the last decades showed that an approximation of dissolved substances could be achieved at a much better accuracy. The idea is now to compile a Gibb's function, from which thermodynamic properties of seawater can be derived, including deviations from standard compositions of salts [e.g., *Feistel and Hagen*, 1995]. The novel approach is based on recent measurements and is planned to be agreed upon and issued in 2008 [see *Feistel and Marion*, 2007].

#### 5.4. Density

[95] Density  $\rho$  is the crucial magnitude for the stability of stratification. Density differences drive currents. Whether and how stable a stratification is can be concluded from comparing density of water parcels under the same pressure conditions. Usually, the reference of normal conditions (atmospheric pressure) is used, and the term potential density may be more correct. Hence density is a magnitude of central importance in limnophysics, but direct measurement in the field is not feasible at the required accuracy [see *Gräfe et al.*, 2002]. As a consequence, the most practical way is to evaluate density from measurements of electrical conductivity and temperature.

[96] In lakes of a composition of dissolved substances similar to the ocean, the so-called UNESCO formula by *Fofonoff and Millard* [1983] may be applied (e.g., *Rassnitzer See* in Figure 10, lakes close to the coast [see *Ludlam*, 1996]), which according to *Fofonoff and Millard*, is only applicable for salinities between 2 and 42 psu. Hence it does not apply for hyperhaline lagoon waters, though the salt is imported from the sea [e.g., *Por*, 1972]. In limnology, depiction of density as “density minus  $1000 \text{ kg m}^{-3}$ ” may be used and referred to as sigma ( $\sigma$ ).

$$\begin{aligned} \rho &= \rho(S_{\text{ocean}}, T_{\text{pot}}) \\ &= \sum_{i=0}^5 a_i T_{\text{pot}}^i + S_{\text{ocean}} \sum_{i=0}^4 b_i T_{\text{pot}}^i + S_{\text{ocean}}^{3/2} \sum_{i=0}^2 c_i T_{\text{pot}}^i + S_{\text{ocean}}^2 d_0, \end{aligned} \quad (16)$$

$$\begin{aligned} a_i &= [999.842594; 6.793952 \times 10^{-2}; -9.095290 \times 10^{-3}; \\ &\quad \cdot 1.001685 \times 10^{-4}; -1.120083 \times 10^{-6}; 6.536332 \times 10^{-9}], \\ b_i &= [0.824493; -4.0899 \times 10^{-3}; 7.6438 \times 10^{-5}; -8.2467 \\ &\quad \times 10^{-7}; 5.3875 \times 10^{-9}], \\ c_i &= [-5.72466 \times 10^{-3}; 1.0227 \times 10^{-4}; -1.6546 \times 10^{-6}] \\ d_0 &= 4.8314 \times 10^{-4} \end{aligned}$$

[97] For low salinities (<0.6 psu), the following formula by *Chen and Millero* [1986] may be used for density calculation:

$$\rho = \rho(S, T_{\text{pot}}) = \sum_{i=0}^6 a_i T_{\text{pot}}^i + S \sum_{i=0}^2 b_i T_{\text{pot}}^i, \quad (17)$$

using

$$\begin{aligned} a_i = & [999.8395; 6.7914 \times 10^{-2}; -9.0894 \times 10^{-3}; \\ & \cdot 1.0171 \times 10^{-4}; -1.2846 \times 10^{-6} \cdot 1.1592 \times 10^{-8}; -5.0125 \\ & \times 10^{-11}] \\ b_i = & [0.8181; -3.85 \times 10^{-3}; 4.96 \times 10^{-5}] \end{aligned}$$

[98] In cases where salinity cannot be used, calculation of density may directly be based on measurements of temperature and conductivity. *Bührer and Ambühl* [1975], for example, propose a formula for Lake Constance where they separate temperature dependence  $\rho_T$  from contribution of dissolved substances  $\Gamma$ . *Heinz et al.* [1990] and *Bäuerle et al.* [1998] used the following variation for Upper Lake Constance (Obersee):

$$\rho = \rho_T + \Gamma = 999.8429 + 10^{-3} \cdot (0.059385 T_{\text{pot}}^3 - 8.56272 T_{\text{pot}}^2 + 65.4891 T_{\text{pot}}) + \Gamma. \quad (18)$$

They used 20°C as reference for the conductance and evaluated the coefficient connecting  $\kappa_{20}$  with density empirically:

$$\Gamma = \gamma \kappa_{20} \text{ and } \gamma = 0.67 \times 10^{-3} \text{ kg m}^{-3} (\mu \text{ S cm}^{-1})^{-1}. \quad (19)$$

[99] An approach that separates temperature and conductivity contributions cannot reflect the effect of lowering temperature of maximum density by dissolved substances (see section 2.2). Hence *Karakas et al.* [2003] allow for a temperature-dependent  $\gamma$ . They use 25°C as reference temperature in the highly acidic Mining Lake 111:

$$\Gamma = \gamma(T) \kappa_{25}, \quad (20)$$

where

$$\begin{aligned} \gamma(T_{\text{pot}}) = & -7.522 \times 10^{-6} T_{\text{pot}}^3 + 2.613 \times 10^{-4} T_{\text{pot}}^2 - 5.146 \\ & \times 10^{-3} T_{\text{pot}} + 0.7918. \end{aligned}$$

[100] In general, many samples can be used for measuring temperature dependence of density. In a second step, a regression curve is created over the two-dimensional field of data. This was done for Lake Goitsche in 1999, for example, when river water was deviated into an abandoned mine pit, where *Gräfe and Boehrer* [2001] included all terms to third order:

$$\rho = \rho(\kappa_{25}, T_{\text{pot}}) = \sum_{i,k \geq 0}^{i+k \leq 3} a_{ik} \kappa_{25}^i T_{\text{pot}}^k. \quad (21)$$

**TABLE 1. Contribution of Dissolved Substances to the Density of Water<sup>a</sup>**

Substance	$\beta_n, \text{ g kg}^{-1}$
Ca(HCO <sub>3</sub> ) <sub>2</sub>	$0.807 \times 10^{-3}$
Mg(HCO <sub>3</sub> ) <sub>2</sub>	$0.861 \times 10^{-3}$
Na(HCO <sub>3</sub> )	$0.727 \times 10^{-3}$
K(HCO <sub>3</sub> )	$0.669 \times 10^{-3}$
Fe(HCO <sub>3</sub> ) <sub>2</sub>	$0.838 \times 10^{-3}$
NH <sub>4</sub> (HCO <sub>3</sub> )	$0.462 \times 10^{-3}$
CO <sub>2</sub>	$0.273 \times 10^{-3}$
CH <sub>4</sub>	$-1.250 \times 10^{-3}$
Air	$-0.090 \times 10^{-3}$

<sup>a</sup>Data from *Imboden and Wüest* [1995].

The resulting 10 coefficients  $a_{ik}$  are specific for the case of the flooding of Lake Goitsche in 1999 and can be retrieved from *Gräfe and Boehrer* [2001].

[101] An alternative method is applicable if the composition of dissolved substances is known. Density can be calculated by using a temperature curve and adding substance-specific contributions to the density temperature curve.

$$\rho = \rho_T(T_{\text{pot}}) \left( 1 + \sum_n \beta_n c_n \right) \quad (22)$$

$$\beta_n = \frac{1}{\rho} \frac{\partial \rho}{\partial c_n} \quad (23)$$

represent coefficients correlating density contribution with concentration of a dissolved substance;  $c_n$  is the concentration of substance  $n$ . Values of  $\beta_n$  are given in Table 1 for common salts and gases. As the assumption is made that the ions do not interfere with each other, given values are only valid for diluted solutions [*Imboden and Wüest*, 1995; see also *Wüest et al.*, 1996].

[102] In addition, suspended material also contributes to density. If no volume effect is considered, the bulk density  $\rho_b$  can be calculated by numerically filling the space of suspended material with its density  $\rho_s$  in the surrounding liquid density  $\rho_l$  at given properties

$$\rho_b = \rho_l + c \times (\rho_s - \rho_l) / \rho_s, \quad (24)$$

where  $c$  is the mass of suspended material per unit water volume ( $\text{kg m}^{-3}$ ).

#### 5.4.1. In Situ Density

[103] Besides stability considerations, density was already mentioned in section 5.1.2. There, however, the term used was in situ density  $\rho_{\text{in situ}}$ , which is the value for density under given environmental conditions, e.g., higher pressure. Density (or potential density) and in situ density differ by the contribution of (little) compressibility from each other. For most considerations, except for stability, in situ density is the proper reference. However, also in cases



of complex stratification, e.g., quantitative considerations about the thermobaric stratification, the concept of potential density cannot reflect conditions appropriately.

[104] Adiabatic compressibility of water lies around  $5 \times 10^{-10} \text{ Pa}^{-1}$ . This means at 200 m depth (potential) density and in situ density differ by about  $10^{-3}$ . This is the same magnitude as weak thermal or weak salinity stratification. Hence the compressibility is the leading term for the increase of in situ density in weakly stratified deep waters of very deep lakes.

[105] *Fofonoff and Millard* [1983] present a system of equations to evaluate  $\rho_{\text{in situ}}$  for waters under ocean conditions. This involves a number of constants which can be extracted from the respective paper. As under limnological conditions (smaller pressure range than the ocean, limited validity of the salinity concept), the in situ density is of limited importance for lake waters of high salinity.

[106] On the contrary, the evaluation of  $\rho_{\text{in situ}}$  is useful in connection with the thermobaric stratification. Starting with (potential) density evaluated from equation (16),  $\rho_{\text{in situ}}$  under given pressure  $p$  bar is calculated as

$$\rho_{\text{insitu}} = \rho(1 - p/K), \quad (25)$$

where for freshwater applications [*Chen and Millero*, 1977, 1986]

$$K[\text{bar}] = \sum_{i=0}^4 c_i T_{\text{pot}}^i + p \sum_{i=0}^2 d_i T_{\text{pot}}^i + S \sum_{i=0}^1 e_i T_{\text{pot}}^i + f p S,$$

$$c_i = [19652.17; 148.113; -2.293; 1.256 \times 10^{-2}; -4.18 \times 10^{-5}],$$

$$d_i = [3.2726; -2.147 \times 10^{-4}; 1.128 \times 10^{-4}],$$

$$e_i = [53.238; -0.313], f = 5.728 \times 10^{-3}.$$

[107] From this, *Chen and Millero* [1977, 1986] present an equation to calculate the temperature of maximum density for low-salinity water; that is, the temperature dependence of density at constant pressure and constant salinity vanishes:  $(\partial \rho_{\text{in situ}} / \partial T_{\text{pot}})_{p,S} = 0$ :

$$T_{\text{md}} = 3.9839 - 1.9911 \times 10^{-2} p - 5.822 \times 10^{-6} p^2 - (0.2219 + 1.106 \times 10^{-4} p) S. \quad (26)$$

#### 5.4.2. Stability

[108] Stability of a water column derives from density increase along the vertical coordinate. The usual quantity for stability is

$$N^2 = \frac{g}{\rho} \frac{d\rho}{dz}. \quad (27)$$

[109] The quantity  $N$  is also called stability frequency or Brunt-Väisälä-frequency (unit  $\text{s}^{-1}$ ), which also indicates the

maximum frequency ( $\omega$ ) for internal waves that can propagate in respective stratification.  $N^2$  indicates how much energy is required to exchange water parcels vertically [e.g., *Boehrer et al.*, 2000a]. Especially in cases where the variability of the thermal expansion coefficient becomes obvious (see sections 3.3 and 5.4.1), derivation of stability as potential temperature gradient finds its limitations [see also *Gill*, 1982].

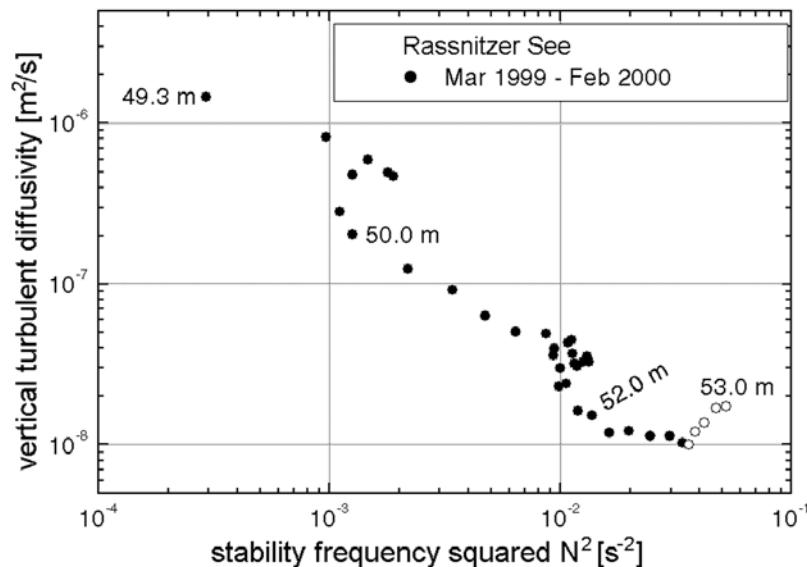
[110] As a consequence, chemical gradients can only persist for longer time periods where density gradients limit vertical transport of dissolved substances (see Figure 23). Lake basin Niemegk of Lake Goitsche (Germany) has been neutralized by introducing buffering river water to the epilimnion (Figure 5). During summer 2000, vertical transport through the temperature stratification was limited, and a chemical gradient in pH could be sustained. However, in winter, temperature stratification vanished, vertical transport was enhanced, and consequently chemical gradients were removed.

## 6. CONCLUDING REMARKS AND PERSPECTIVES

[111] This review presents an overview of lacustrine stratification features. It is designed to enable a limnologist to correctly interpret a stratification that he/she may encounter in field measurements. We hope that after reading this contribution, limnologists and geoscientists are able to understand what processes create stratification and which approaches can be taken to evaluate fundamental quantities such as conductance or density. References are given to provide the reader with access to seminal publications in the field.

[112] Though actual stratification in a lake results from a balance between stratifying processes and mixing processes, the former have been investigated less quantitatively than the latter. This may have been a consequence of the similarities of mixing processes in atmosphere, ocean, and lakes, while many stratifying processes do not occur in atmosphere or ocean, or at least do not play an important role, e.g., geochemical processes or thermobaric stratification close to the temperature of maximum density. As a consequence, this paper has focused on stratifying processes in lakes. Contributions of heat, dissolved or suspended substances, and pressure on the density have been discussed. Also, most widely accepted approaches for quantitatively evaluating stability of stratification have been introduced.

[113] Reliable numerical simulations of annual temperature cycles of lakes deceive us into believing that problems concerning stratification prognostications have all been solved. However, not much quantitative material is available on transport in strongly stratified deep waters. Prognosticating the evolution of meromixis even in the (simple) case of conservative salts remains a challenge. Thermobaric stratification can be observed in lakes which are deep enough and located in the corresponding climate zone. The stratification features encountered in these lakes can



**Figure 23.** Turbulent diffusive transport of an artificial tracer ( $\text{SF}_6$ ) in the strongly stratified monimolimnion of Rassnitzer See versus density gradient,  $N^2 = -g/\rho(d\rho/dz)$  (adapted from von Rohden and Ilmberger [2001] with permission from Birkhäuser).

be explained with the current knowledge. Numerical considerations show that deep water renewal can be facilitated by wind action in Lake Baikal, for example. However, a demonstration that a stratification model is capable of reproducing temperature profiles as found in these lakes and possibly illuminating the complex conditions when replacing deep waters is still outstanding.

[114] A number of geophysical processes have been identified which play a role in sustaining stratification. Regardless of whether more geochemical cycles need considering, quantifying precipitation and coprecipitation as well as dissolution and liberation of substances in the monimolimnetic environment has just started. This is especially true for the effect on lake stratification at the accuracy needed for prognostications. In some cases, such processes can be decisive factors for whether a lake turns meromictic or not. Aware of many limitations, better prognostication instruments are nevertheless required. Even in complex situations, uncertainties can be limited to a smaller number of factors. A large number of lakes would profit from improving numerical tools.

[115] A healthy aquatic environment is a key component for a good quality of life. In the perspective of global change, predictive tools for the evolution of ecological function of lacustrine ecosystems are indispensable. Climate change, changing land use, decommissioning disrupted mining or industrial landscapes, or the creation of artificial water bodies poses new questions to limnology. If timely prognosticated, some undesirable evolutions may be avoided, while others may have at least been forecast in time to initiate legal and social action to prevent the worst. Understanding and communicating the causal connection is fundamental for public acceptance of legislative steering to

responsibly compromise between use of the environment and the conservation of its sound ecological status.

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