# 18 Alpine tectonics of the Alps and Western Carpathians

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The Alps and Western Carpathians constitute that part of the Alpine-Mediterranean orogenic belt which advances furthest to the north into Central Europe. They were formed by a series of Jurassic to Tertiary subduction and collision events affecting several Mesozoic ocean basins, continental margins, and continental fragments. The Western Alps form a pronounced, westward-convex arc around which the strike of the tectonic units changes by almost 180° (Fig. 18.1). The Western Carpathians are a northward-convex arc of similar size but with minor curvature. The two arcs are connected by an almost straight, WSW–ENE striking portion including the Eastern Alps

Stresses produced by tectonic processes in the Alps also influenced the tectonics of large parts of central and northern Europe, leading, for example, to basin inversion and strike-slip faulting. In this chapter, we will discuss the present-day structure of the different tectonic units in the Alps and Western Carpathians in relation to their palaeotectonic history in order to illustrate the plate tectonic evolution using geological data. Many tectonic problems of the Alps and Western Carpathians are still unsolved, although dramatic progress has been made, especially over the last *c.* 20 years. Therefore, some of the interpretations presented below are still controversial and do not always express the opinion of all three authors. Given that the main theme of this book is Central Europe, the Southern and Western Alps are discussed in less detail than those parts of the Alps which belong to Central Europe: the Central Alps, the Eastern Alps and the Western Carpathians.

## The Alps (N.F., R.S.)

### *Geographic outline*

The Alps form an arcuate mountain chain *c.* 1000 km long and between 120 and 250 km wide. The eastern end of the Alps follows approximately a north–south orientated line from Vienna towards the south. Along this line, the Alpine tectonic units disappear beneath the sediment fill of Cenozoic basins, the Vienna Basin to the north and the Styrian Basin further south. Additionally, various Alpine tectonic units reappear in the Western Carpathians and this chain represents the continuation of the northern part of the Alps. To the SE, there is no clear limit between the Southern Alps and the Dinarides, either morphologically or geologically. Hence the Dinarides represent the continuation of the Southern Alps, just as the Western Carpathians prolongate the northern part of the Alps. South of the Alps, on the inner side of the arc, is the Cenozoic Po Basin. This is the foreland basin of the Southern Alps, since this part of the Alps forms a south-directed thrust-and-fold belt. The Po Basin is closed in the west by the westward-convex arc of the Western Alps. At the southern end of this arc, the Alps continue into the Apennines located south of the Po Basin. A complex north–south orientated fault zone, the Sestri-Voltaggio Zone, is taken as the boundary between the Alps and the Apennines. Across this line, the direction of tectonic transport changes: the main thrusts on the Alpine side are SW-directed, whereas the younger thrusts on the Apenninic side (i.e. from late Middle Eocene onward; Marroni *et al.* 2002) are NE-directed, so that the Po Basin also serves as the foreland basin for the Apennines. On its outer, western side, the arc of the Western Alps is in contact with the Provençal chains, a series of east–west striking ridges which represent, in a structural sense, an eastern continuation of the Pyrenees. North of these chains, the Alps are bordered by the Cenozoic fill of the Rhone-Bresse graben, and still further north, the chain of the Jura Mountains diverges in a northwesterly direction from the westernmost part of the Alps. This arcuate, outward-vergent fold-and-thrust belt is often treated as part of the Alps because the folds branch off from the external zone of the Western Alps, and because its basal thrust is probably rooted at depth under the Alps. In a topographic sense, however, the Jura Mountains are distinct from the Alps and form an independent range, enclosing, together with the Alps, the Swiss Molasse Basin, which is the westernmost part of the northern foreland basin of the Alps.

### *Geographic subdivision of the Alps*

A large-scale geographic subdivision of the Alps distinguishes between the Western, Central, Eastern, and Southern Alps (Fig. 18.1). (In the German and Austrian literature, the Central Alps are included in the Western Alps.) The boundary between the Southern Alps and the other parts of the Alps is defined by a system of important east–west orientated valleys (e.g. Valtellina, Pustertal, Gailtal). These valleys are the morphological expression of the Periadriatic Fault, a major fault of Tertiary age. Therefore, this boundary is both a geographic and a structural one.

The geographic boundary between the Eastern and Central

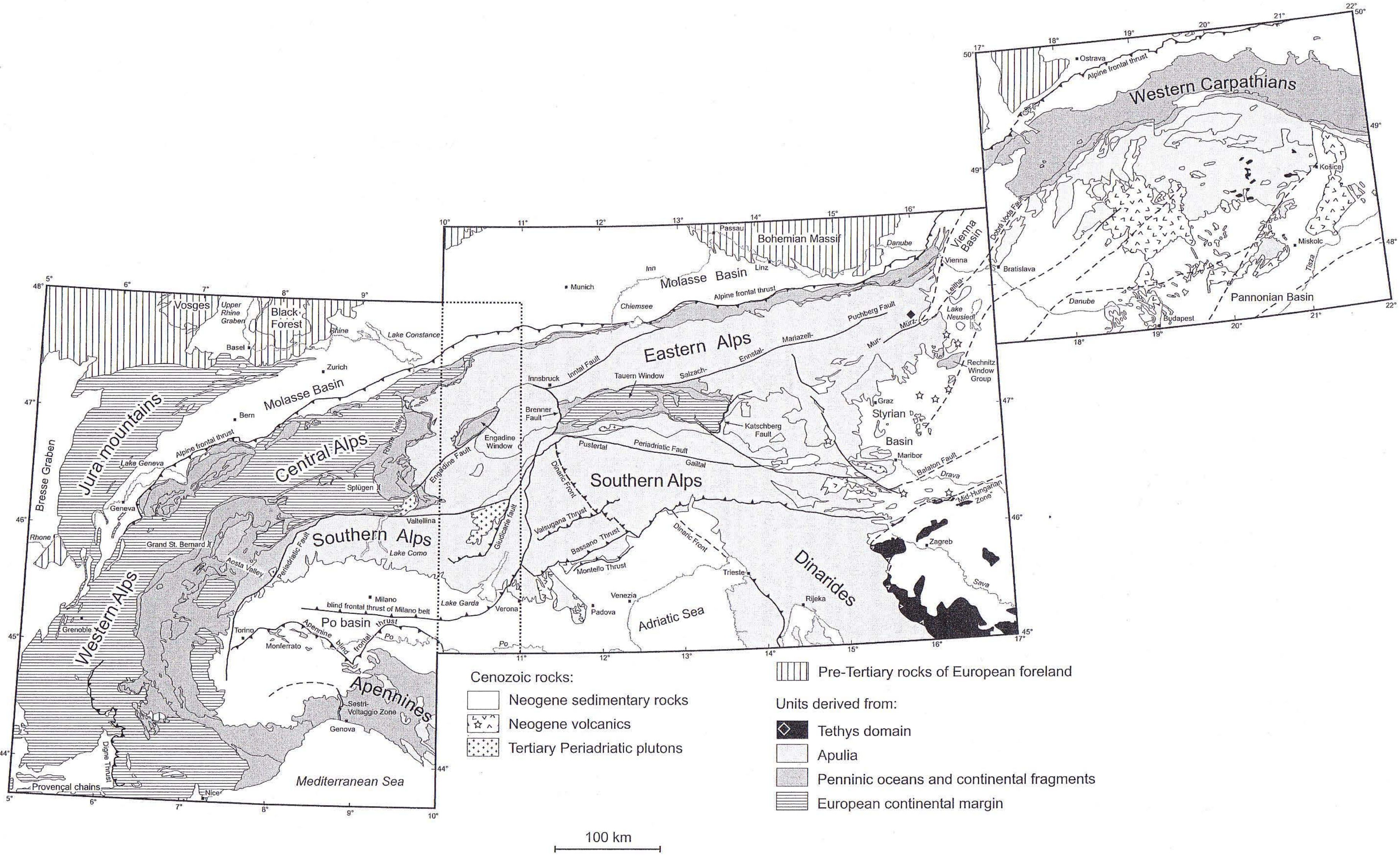



**Fig. 18.1.** Tectonic map of the Alps and Western Carpathians. After Schmid *et al.* (2004*a*) and other sources.

Alps is drawn from Lake Constance southward along the Rhine Valley and over the Splügen Pass to Lake Como, the boundary between the Central and Western Alps from Lake Geneva south along the Rhone Valley, over the Grand St. Bernard Pass, and along the Aosta Valley to the Po Plain.

## *Tectonic subdivision*

In terms of tectonics, the Alps are subdivided into structural zones, termed Helvetic, Penninic, Austro-Alpine and South Alpine, each of which is again subdivided into smaller units (e.g. Sub-Penninic, Lower, Middle and Upper Penninic; Fig. 18.2). This subdivision coincides with the geographical one only in the case of the South Alpine Zone which is identical with the Southern Alps. The South Alpine Zone is the part of the Alpine edifice to the south of the Periadriatic Line. It forms a south-directed thrust-and-fold belt (a 'retro-wedge' since the formation of the Alps is related to southward subduction). The South Alpine Zone is characterized by the almost complete absence of an Alpine metamorphic overprint. In contrast, the Austro-Alpine, Penninic and Helvetic zones are partly metamorphosed and constitute the north-directed (or, in the Western Alps, west-directed) 'pro-wedge' of the Alps. They form three groups of thrust sheets or nappes, of which the Austro-Alpine is structurally highest, the Penninic is in a middle position, and the Helvetic is at the base. Thrusting and metamorphism in the Austro-Alpine Zone are mainly Cretaceous in age whereas they are Tertiary in the Penninic and Helvetic zones. All of these zones have distinct sedimentary facies in the Mesozoic. The facies boundaries, however, are not always strictly parallel to the later tectonic boundaries (thrusts). Furthermore, these terms were introduced long before plate tectonics and, hence, do not denote plate tectonic entities such as oceans or continental margins. Only the Austro-Alpine–Penninic boundary coincides with a plate tectonic boundary: the boundary between the Apulian continental crust and the Piemont-Ligurian oceanic crust.

## *Palaeogeographic subdivision*

In terms of Mesozoic palaeogeography, the Alps were constructed by units which developed from a range of continental and oceanic domains. From NW to SE, these include (Fig. 18.3): (1) the European shelf and continental margin (Helvetic nappes and External massifs, Sub-Penninic nappes); (2) a northern ocean basin, the Valais Ocean (Lower Penninic nappes); (3) the Briançonnais Terrane, representing a northeastern prolongation of Iberia (Middle Penninic nappes); (4) a southern ocean basin, the Piemont-Liguria Ocean (Upper Penninic ophiolite nappes); (5) the Cervinia terrane, a narrow, elongate continental fragment within the Piemont-Liguria Ocean (Upper Penninic continental nappes); (6) the Apulian continental margin (Austro-Alpine nappes and South Alpine units); and (7) a southeastern ocean basin, the Meliata Ocean (only present as small slivers in the Austro-Alpine Zone but more broadly exposed in the Western Carpathians).

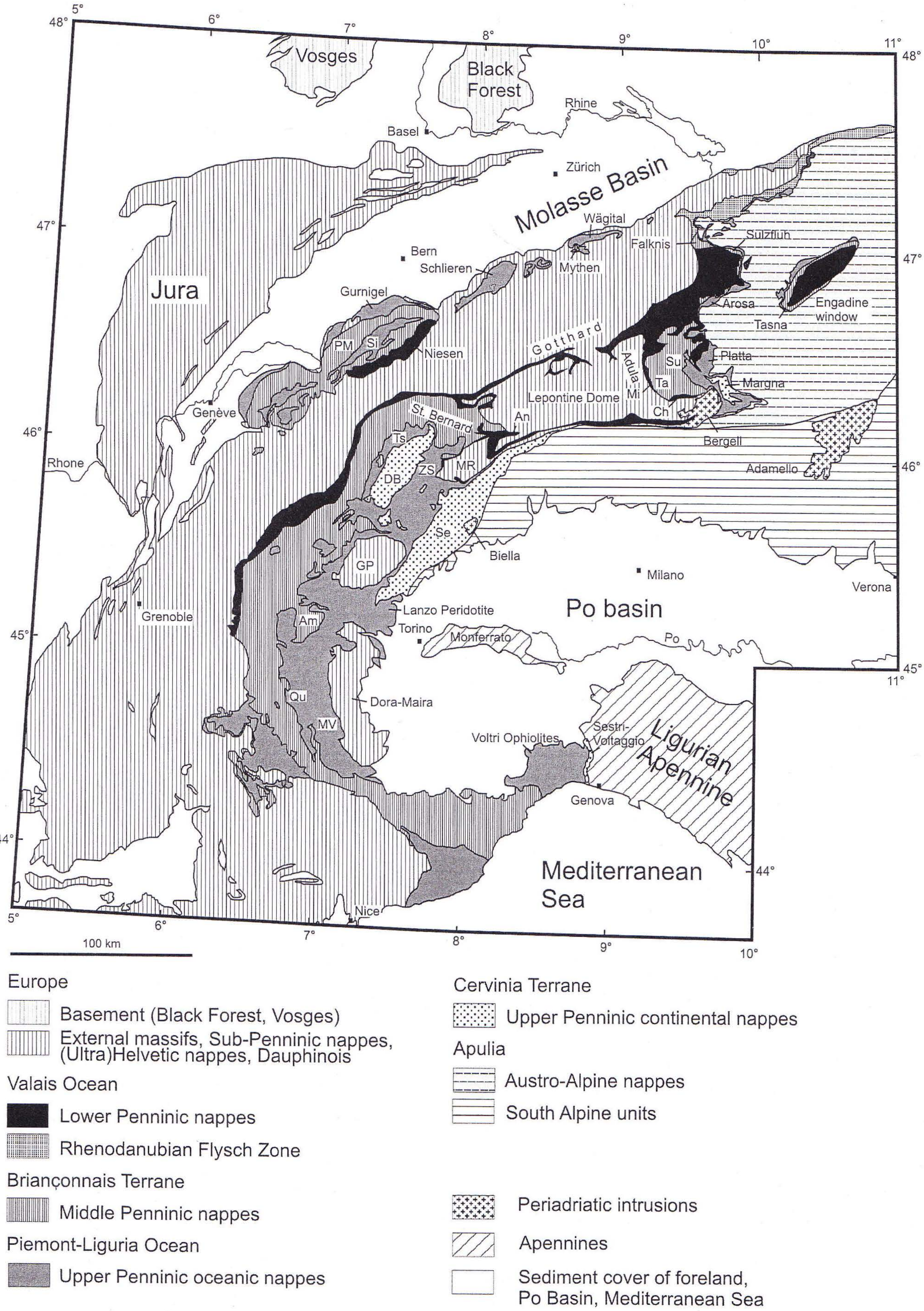
Both the opening and closure of the oceanic basins, as indicated by radiometric ages of oceanic magmatites and high-pressure metamorphism, respectively, show a broad trend towards younger ages from SE to NW (Fig. 18.4).

## *Outline of palaeogeographic evolution*

During the Triassic, a sedimentary prism representing a typical passive continental margin developed in the future Alps. A wide carbonate platform developed in the Austro-Alpine and South Alpine zones and was bordered to the NW by more terrestrially influenced sedimentation in the Penninic and Helvetic zones. The ocean to which this continental margin belongs must have been located to the east or SE of the Alpine area. An association of deep-marine Ladinian- to Carnian-age sediments (radiolarite), basalts partly overprinted by blueschist-facies metamorphism of middle Jurassic age, and Jurassic deep-marine clastics cropping out at Meliata (Slovakia) in the southern foothills of the Western Carpathians, is interpreted to be a remnant of this Triassic ocean which is therefore called the Meliata Ocean (Kozur & Mock 1973). Similar associations are present in the mélange zones of the Dinarides and occur as slivers in the Eastern Alps, within the Austro-Alpine nappe pile (Kozur & Mostler 1992; Mandl & Ondrejickova 1993). The Meliata Ocean is interpreted as a marginal basin of Tethys (Stampfli & Borel 2004). The exact geometry of the Mesozoic-age continents and oceans in south-eastern Europe, however, is still controversial.

The closure of the Triassic Meliata Ocean commenced at the end of the Triassic (Kozur 1991), probably at a SE-dipping, intra-oceanic subduction zone. Backarc spreading above the subduction zone created Jurassic oceanic crust (Vardar Ocean). The collision of the northwestern (Apulian) continental margin with the arc, above the subduction zone, during the Jurassic led to obduction of ophiolite sheets onto the Apulian margin in the Dinarides and probably also in the Austro-Alpine Zone. By that time the Piemont-Ligurian Ocean had opened farther to the NW. Ophiolites from this basin can be traced along the strike of the Alps. The first oceanic crust in the Piemont-Ligurian Ocean was formed during the Middle Jurassic. The Piemont-Ligurian Ocean was kinematically linked to the Middle Atlantic that opened at the same time, in that the Middle Atlantic continental breakup propagated northward only as far as the latitude of Gibraltar from where a strike-slip or transtensional zone extended eastward into the future western Mediterranean region, transferring the east–west separation from the Middle Atlantic into the SW–NE striking Piemont-Ligurian Ocean. It is not clear if and how the Piemont-Ligurian Ocean continued east of the Alps and into which basin of the SE European Tethyan realm it finally merged. The Cervinia terrane (Pleuger *et al.* 2007) was an elongate fragment of continental crust, separated from the Apulian continental margin by serpentinitic and basaltic ocean floor (Fig. 18.3). It has been proposed that this fragment is an extensional allochthon which rifted away from the Apulian margin by top-to-the-NW detachment faulting (Froitzheim & Manatschal 1996). Alternatively, the fragment may have resulted from a ridge jump or the simultaneous opening of two sub-basins during an early phase of opening of the Piemont-Ligurian Ocean.

The second oceanic basin of the Penninic Zone, the Valais Ocean, opened in the Early Cretaceous to the NW of the Piemont-Ligurian Ocean (Fig. 18.3). Ophiolites and sediments from this basin are well developed in Switzerland and can be traced to the SW into France as far as the town of Moutiers, from whence they disappear to the south. The continental fragment between the Valais and Piemont-Liguria ocean basins is termed the Briançonnais fragment, after a system of cover nappes derived from this continent cropping out around the town of Briançon in the Western Alps. In the Eastern Alps the ophiolites of the Valais and Piemont-Ligurian oceans are separated by nappes of the Briançonnais fragment only in the Engadine Window. Towards the east, in the Tauern Window and the Rechnitz Window Group, the Briançonnais fragment is missing. This suggests that the Briançonnais fragment wedged out towards the east and that the two oceanic basins merged into


Vosges
Black Forest
Rhine
Basel
Zürich
Molasse Basin
Wägital
Falknis
Sulzfluh
Bern
Schlieren
Mythen
Jura
Gurnigel
PM
Si
Niesen
Gotthard
Arosa
Engadine window
Tasna
Adula
Su
Platta
Ta
Margna
Mi
Lepontine Dome
Ch
Genève
St. Bernard
An
Bergell
Rhone
Ts
ZS
MR
DB
Adamello
Se
Biella
GP
Milano
Verona
Lanzo Peridotite
Po basin
Grenoble
Am
Torino
Monferrato
Po
Qu
Dora-Maira
MV
Voltri Ophiolites
Sestri-Voltaggio
Ligurian Apennine
Genova
Mediterranean Sea
Nice
100 km
5°
6°
7°
8°
9°
10°
11°
44°
45°
46°
47°
48°
Europe
Basement (Black Forest, Vosges)
External massifs, Sub-Penninic nappes, (Ultra)Helvetic nappes, Dauphinois
Valais Ocean
Lower Penninic nappes
Rhenodanubian Flysch Zone
Briançonnais Terrane
Middle Penninic nappes
Piemont-Liguria Ocean
Upper Penninic oceanic nappes
Cervinia Terrane
Upper Penninic continental nappes
Apulia
Austro-Alpine nappes
South Alpine units
Periadriatic intrusions
Apennines
Sediment cover of foreland, Po Basin, Mediterranean Sea

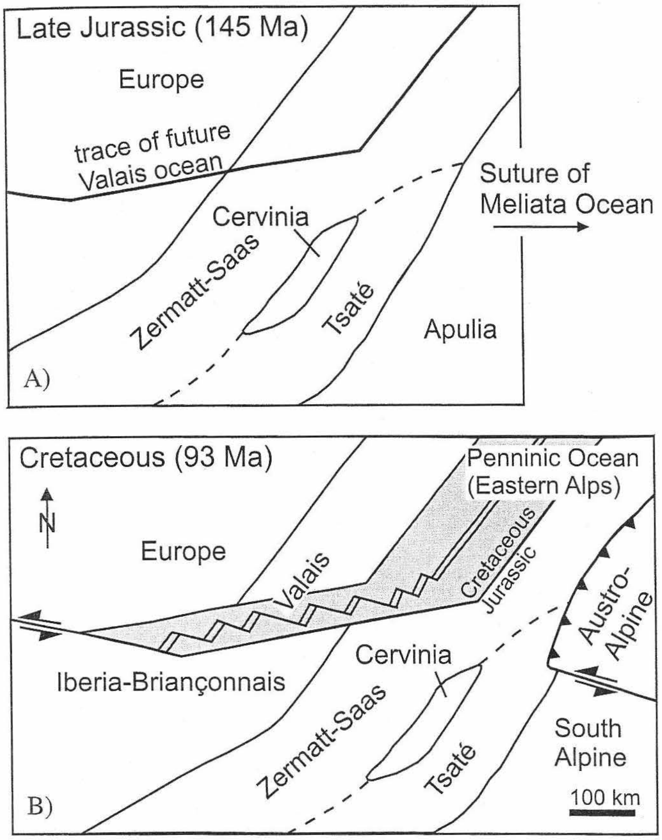



**Fig. 18.3.** Palaeogeographic reconstruction of Penninic oceans in Late Jurassic (A) and Late Cretaceous time (B). (**A**) Jurassic crust formed in the Piemont-Ligurian Ocean; the Cervinia terrane represents either a microcontinent or an extensional allochthon isolated from the Apulian margin during rifting. (**B**) Sinistrally transtensive opening of the Valais Ocean between Iberia and Europe during the Cretaceous. Towards the east Jurassic crust was re-rifted in the Cretaceous. Note that the Valais Ocean contains both Cretaceous crust (dark grey) and Jurassic crust (light grey) originally formed in the Piemont-Ligurian ocean.

one. However, in these areas the subdivision of the Valais and Piemont-Ligurian basins is problematic and several interpretations exist (e.g. Tollmann 1987; Froitzheim *et al.* 1996).

The Valais Ocean is related to Atlantic Ocean opening in a similar way to the Piemont-Ligurian Ocean (Frisch 1979; Stampfli 1993). The continental breakup in the Atlantic propagated towards the north during the Lower Cretaceous as far as the Bay of Biscay, from where a transtensional basin extended eastwards through the later Pyrenees and joined the Valais Ocean of the Alps, a connection which was later obliterated by Alpine deformation. Accordingly, the Briançonnais fragment formed a peninsular prolongation of Iberia into the Alpine realm. Recent dating of metagabbroic rocks from the Penninic Zone (Liati *et al.* 2003; Liati & Froitzheim 2006) yielded ages as young as 93 Ma for oceanic crust of the Valais Ocean, but also Jurassic ages of around 155 to 160 Ma, suggesting that Cretaceous opening of the Valais Ocean involved renewed rifting of Jurassic oceanic crust (Figs 18.3 & 18.4).

The closure of the Penninic oceans began in the Cretaceous with the southeastward subduction of Piemont-Ligurian lithosphere beneath the southern continental margin represented by the Austro-Alpine and South Alpine zones. The precise timing is, however, still uncertain. According to Wagreich (2001*a*) subduction began in the Late Aptian/Albian (*c.* 112 Ma). Thrust imbrication in the most internal parts of the Austro-Alpine Zone had commenced earlier in the Late Jurassic and continued prograding towards the NW during the Early Cretaceous. Thrusting within the Austro-Alpine Zone can, therefore, not have resulted from Penninic subduction, but was related to the collisional closure of the Meliata Ocean and post-collisional shortening. In this collision, parts of the Austro-Alpine Zone represented the lower plate. These were affected during the Cretaceous by thrusting and a metamorphic cycle including locally high-pressure (Hoinkes *et al.* 1999) and ultrahigh-pressure (Janák *et al.* 2004, 2006) metamorphism. During the Late Cretaceous, the Austro-Alpine nappe stack was extended in a west–east to NW–SE direction, leading to ductile deformation and normal faulting. Large parts subsided below sea level and were covered by clastic marine sediments (deposition of the Gosau Group from Turonian to Paleogene). At the same time, Piemont-Ligurian oceanic lithosphere was subducted from the north and NW beneath the Austro-Alpine and South Alpine zones of the Apulian continent.

The Penninic oceans were completely subducted by Eocene times (*c.* 40 Ma). An uncertainty exists about the kinematics of consumption of the Penninic oceans. Some authors assume that there was only one slab descending towards the south, which brought first the Piemont-Ligurian, then the Briançonnais, and finally the Valaisan crust to the foot of the accretionary wedge (Schmid *et al.* 1996). Others, however, assume that there were two subduction zones, one in the Piemont-Ligurian and one in the Valais Ocean (Frisch 1979; Froitzheim *et al.* 2003).

## *Tectonic units derived from the European margin*

The European margin of the Penninic oceans was affected by Alpine tectonics in a progression from internal to external zones. It is represented by basement and cover units that have been deformed and metamorphosed to varying degrees, from moderate folding and thrusting in the external parts to high-grade metamorphism and complete transposition in the most internal parts. The sedimentary units have been partly detached from their basement and accumulated in the external parts. The following units are, in our view, derived from the European margin: the External massifs and their cover, the Dauphinois Zone, and the Helvetic, Ultra-Helvetic and Sub-Penninic nappes.

### *External massifs*

The External massifs (Fig. 18.5) are structural culminations where Variscan basement of the European plate is exposed in the external part of the Alps. The basement consists predominantly of various gneisses, amphibolites and granitoids. The following massifs occur from SW to NE along the strike of the Alps: Argentera, Pelvoux, Belledonne, Aiguilles Rouges, Mont Blanc and Aare. In addition, some smaller basement lamellae occur on the internal side of the larger massifs, such as the Mont Chetif Massif to the SE of Mont Blanc and the Tavetsch massifs to the

**Fig. 18.2.** Palaeogeographic units of the Western and Central Alps (after Froitzheim *et al.* (1996) and Schmid *et al.* (2004*a*), modified). Abbreviations: Am, Ambin; An, Antrona Ophiolites; Ch, Chiavenna Ophiolites; DB, Dent Blanche Nappe; GP, Gran Paradiso Nappe; MV, Monte Viso; PM, Préalpes Médianes Nappe; Qu, Queyras; Se, Sesia Nappe; Si, Simme Nappe; Su, Suretta Nappe; Ta, Tambo Nappe; Ts, Tsaté Nappe; ZS, Zermatt-Saas Zone.

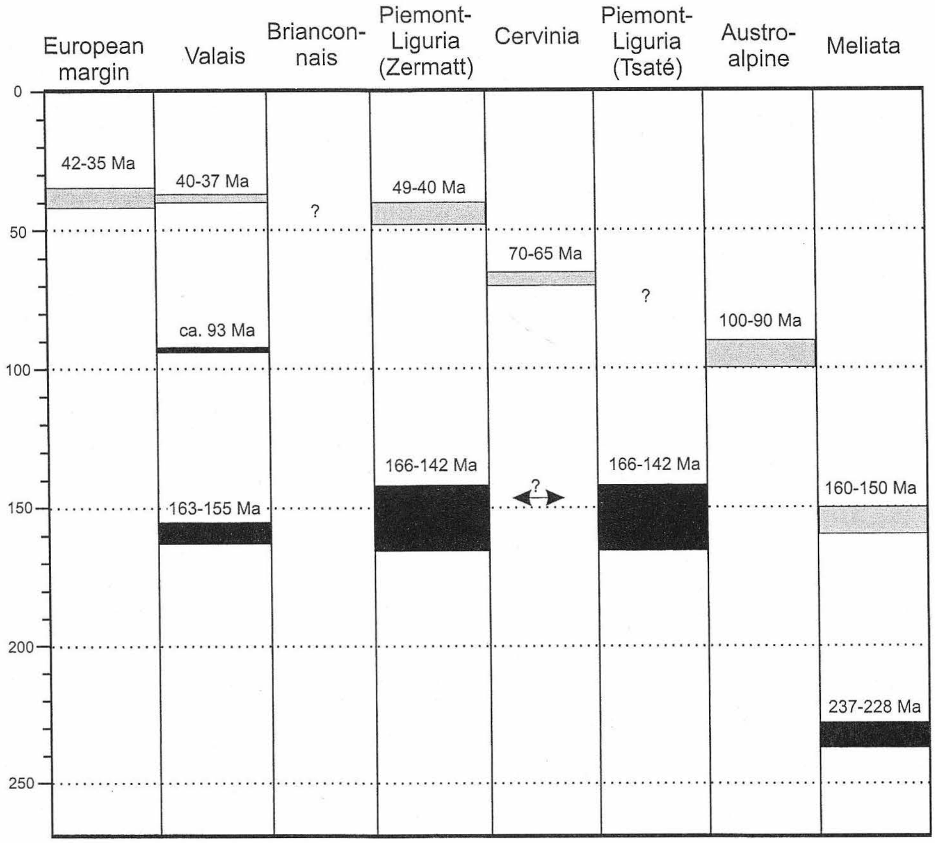



**Fig. 18.4.** Age data for oceanic spreading (black bars) and high-pressure metamorphism during subduction (grey bars) of Alpine palaeogeographic units. All ages are radiometric data, except for spreading of the Meliata ocean which is based on stratigraphic data (Channell & Kozur 1997). For oceanic spreading ages of the Piemont-Ligurian ophiolites, affiliation to the Zermatt-Saas or Tsaté sub-basin is in some cases unclear. If possible, preference was given to U–Pb, Sm–Nd and Lu–Hf ages. Sources: Dallmeyer *et al.* (1996), Duchêne *et al.* (1997), Faryad & Henjes-Kunst (1997), Gebauer (1996), Gebauer *et al.* (1997), Lapen *et al.* (2003, 2007), Liati *et al.* (2003, 2005), Liati & Froitzheim (2006), Miller *et al.* (2005), Rubatto *et al.* (1998, 1999), Stucki *et al.* (2003), Tilton *et al.* (1991).

SE of the Aare Massif. The Gotthard Nappe (Figs 18.6 & 18.7; Gotthard Massif of older authors) is now generally included in the Sub-Penninic nappes (as suggested by Milnes 1974) and is no longer treated as an External massif. The Alpine metamorphism of the External massifs reaches the greenschist facies in the more internal parts.

The position of the massifs is controlled by thrust geometry. The Aare, Mont Blanc and Aiguilles Rouges massifs are ramp anticlines above thrust faults cutting the upper crustal basement (Fig. 18.6 & 18.7). These thrusts formed during the Miocene, leading to uplift and cooling of the external massif basement and to antiformal bending of the earlier (Oligocene to Miocene) basal thrusts of the Helvetic nappes. On the external sides of the massifs, the outward dip of thrusts that results from this bending has often been misinterpreted as indicating gravitational emplacement of the Helvetic nappes. The position of the Pelvoux Massif (Fig. 18.5), in contrast, results from an interference of Oligocene–Miocene west-directed thrusting with structures formed during two older, pre-Priabonian, thrusting and folding events that are related to the Pyrenean collision (Ford 1996; Sue *et al.* 1997). The northern limit of the area affected by the Pyrenean shortening runs along the NW border of the Pelvoux Massif (Fig. 18.5). The geometry of the Pelvoux and Belledonne massifs was also strongly predetermined by Jurassic rift tectonics. Half-grabens formed during Jurassic east–west directed extension were transformed into pinched cover synclines during Alpine shortening, and the east-dipping master normal faults of these grabens were partly transformed into thrust faults.

*Cover of the External massifs, Dauphinois Zone*

The autochthonous and parautochthonous sedimentary cover crops out in Switzerland only in the vicinity of the External massifs, below the Helvetic nappes. Towards the west, into France, the displacement of the overlying Helvetic nappes decreases and they become indistinct, so that the area constituted by the autochthonous and parautochthonous cover units widens considerably. In this sense, the Dauphinois Zone in France (Fig. 18.5), also known as the Chaînes Subalpines, is the continuation of the autochthonous and parautochthonous cover units in Switzerland. However, the shortening of the Dauphinois Zone is important and is accommodated by outward-directed thrust systems, for example, the Digne Thrust (Fig. 18.5).

In Switzerland, the autochthonous/parautochthonous cover comprises locally Upper Carboniferous and Permian graben fill, followed by Triassic to Eocene, predominantly shallow-water sediments, and by the late Eocene to Oligocene North Helvetic Flysch (Fig. 18.7). An important member of the latter is the 32–29 Ma (Boyet *et al.* 2001) Taveyannaz Sandstone with detritus including 32 Ma andesite, providing evidence for the existence of andesitic volcanoes at that time. This sandstone also occurs in

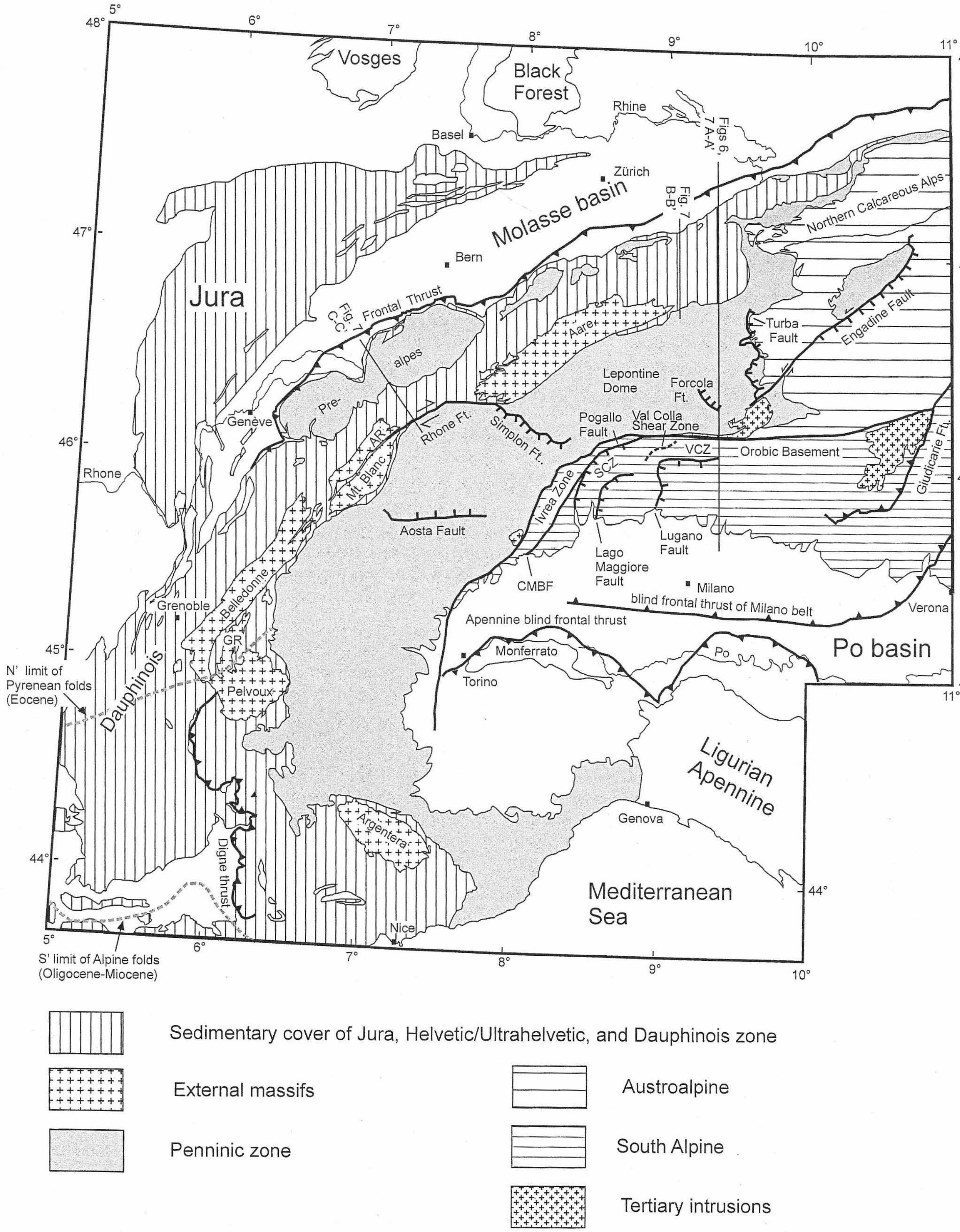



**Fig. 18.5.** Map of the Western and Central Alps with some important faults indicated. Abbreviations: AR, Aiguilles Rouges Massif; CMBF, Cossato-Mergozzo-Brissago Fault; GR, Grandes Rousses Massif; SCZ, Strona-Ceneri Zone; VCZ, Val Colla Zone.

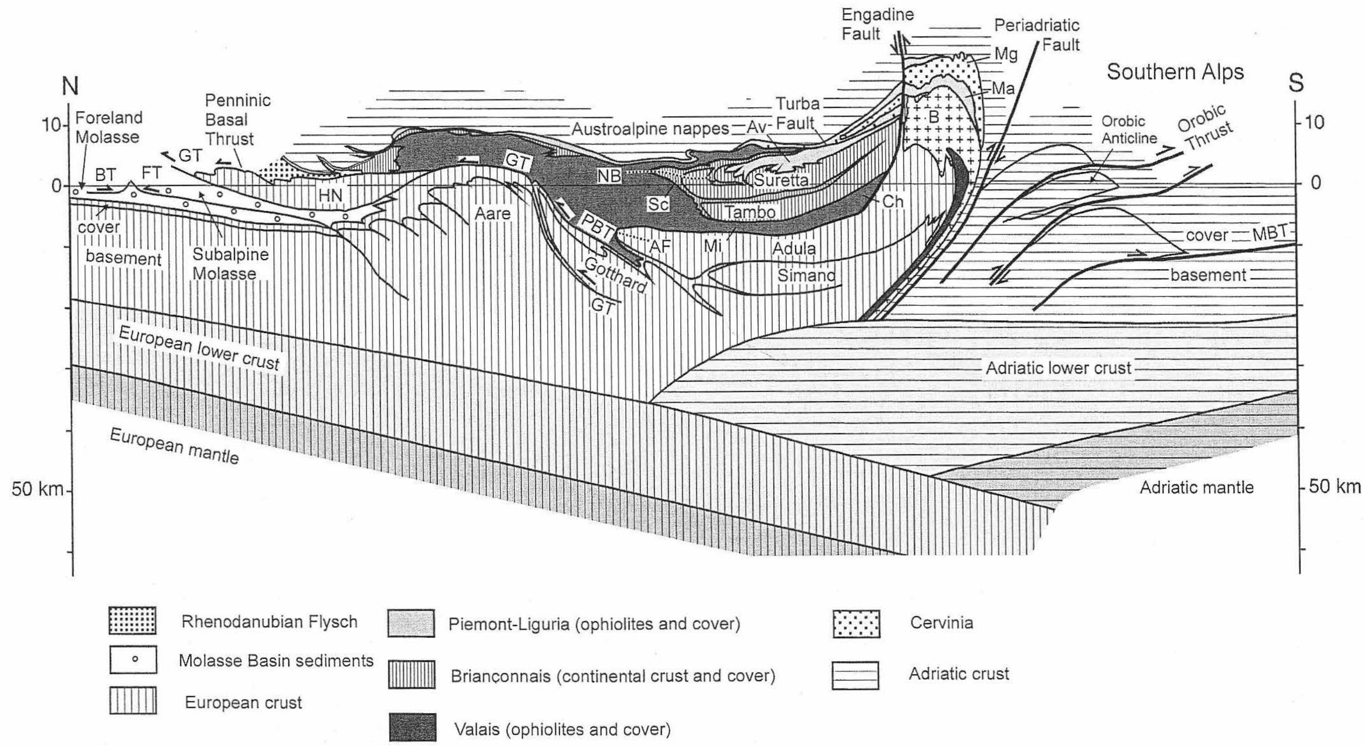



**Fig. 18.6.** Cross-section through the Central Alps (modified after Schmid *et al.* 1996). Abbreviations: AF, axial trace of the frontal fold of the Adula Nappe; Av, Avers Bündnerschiefer Nappe; B, Bergell granitoids; BT, backthrust in the Molasse Basin; Ch, Chiavenna Ophiolite; FT, frontal thrust of the Alps; GT, Glarus Thrust; HN, Helvetic Nappes; Ma, Malenco Ultramafic Complex; Mg, Margna Nappe; MBT, basal thrust of Milano belt (Miocene); Mi, Misox Zone; NB, axial trace of the Niemet-Beverin Fold; PBT, Penninic Basal Thrust; Sc, Schams Nappes. See Figure 18.5 for location of cross-section.

the Helvetic Nappes (Wildhorn Nappe in western Switzerland). The equivalents of the North Helvetic Flysch in the Western Alps are the Champsaur (also andesite-bearing) and Annot sandstones in the Dauphinois Zone and the Aiguilles d'Arves Flysch in a position similar to the Helvetic nappes.

The Morcles Nappe in western Switzerland represents the cover of the external part of the Mont Blanc Massif, partly detached from the massif and folded into a recumbent anticline with a thick normal limb and an extremely thinned inverted limb. The thin Ardon Nappe on top of the Morcles Nappe is the parautochthonous cover of the internal Mont Blanc Massif (Fig. 18.7, profile C–C′).

In the northern foreland of the Eastern Alps, the eastward continuation of the autochthonous and parautochthonous sedimentary cover units is buried beneath sediments of the Molasse Basin. In the Waschberg Zone near Vienna, however, shallow-marine Malm-age limestones are visible at the surface.

*Helvetic nappes*

The Helvetic nappes are classically developed in the Central Alps of Switzerland. They are cover nappes without a pre-Permian basement. The sedimentary succession locally comprises the clastic and volcanic fill of Permian rift basins (e.g. Glarus Verrucano in Eastern Switzerland; Fig. 18.7, profile B–B′), followed by thin Triassic sediments in a Germanic facies, Jurassic deposits with variable thickness due to rift tectonics, Cretaceous to Palaeogene shelf deposits, and locally Eocene to Early Oligocene deep-marine clastics. Some of the Helvetic nappes are thrust nappes lacking an inverted limb, for example, the Glarus Nappe in eastern Switzerland. This nappe is floored by the most famous thrust fault of the Alps, the Glarus Thrust (Figs 18.6 & 18.7), along which Permian-age Verrucano sedimentary and volcanic rocks overlie Jurassic- to Tertiary-age sediments of the Infra-Helvetic Complex (see below). Others are fold nappes with a well-developed inverted limb such as the Wildhorn Nappe in western Switzerland. The sequence of the Helvetic nappes in eastern Switzerland comprises (from base to top) the Glarus Nappe, the Mürtschen Nappe, the Axen Nappe and the Säntis-Drusberg Nappe (Fig. 18.7, profile B–B′). The first three of these are not independent nappes; their bounding thrusts become less important towards the south and die out in the internal parts of the Helvetic nappes. In western Switzerland, the Helvetic nappes are represented by the lower Diablerets Nappe and the higher Wildhorn Nappe. The Ardon and Morcles nappes, underlying the Diablerets Nappe, are not Helvetic nappes in the strict sense but are parautochthonous cover units of the internal and external parts, respectively, of the Mont Blanc Massif (see above).

The basement from which the eastern Helvetic nappes were detached is more internal than the Aare Massif since the latter has its own sediment cover in an autochthonous to parautochthonous position. The basement for the eastern Helvetic nappes is represented by the Tavetsch 'Zwischenmassif' which is a narrow, strongly deformed basement complex SE of the Aare Massif, and by the Sub-Penninic Gotthard Nappe (Fig. 18.7, profile B–B′). Other parts of this basement have disappeared at depth, between the Aare Massif and the Tavetsch Massif, and between the latter and the Gotthard Nappe.

In the Western Alps, the Aiguilles d'Arves Unit (Ceriani *et al.* 2001; Ceriani & Schmid 2004) occupies a similar position

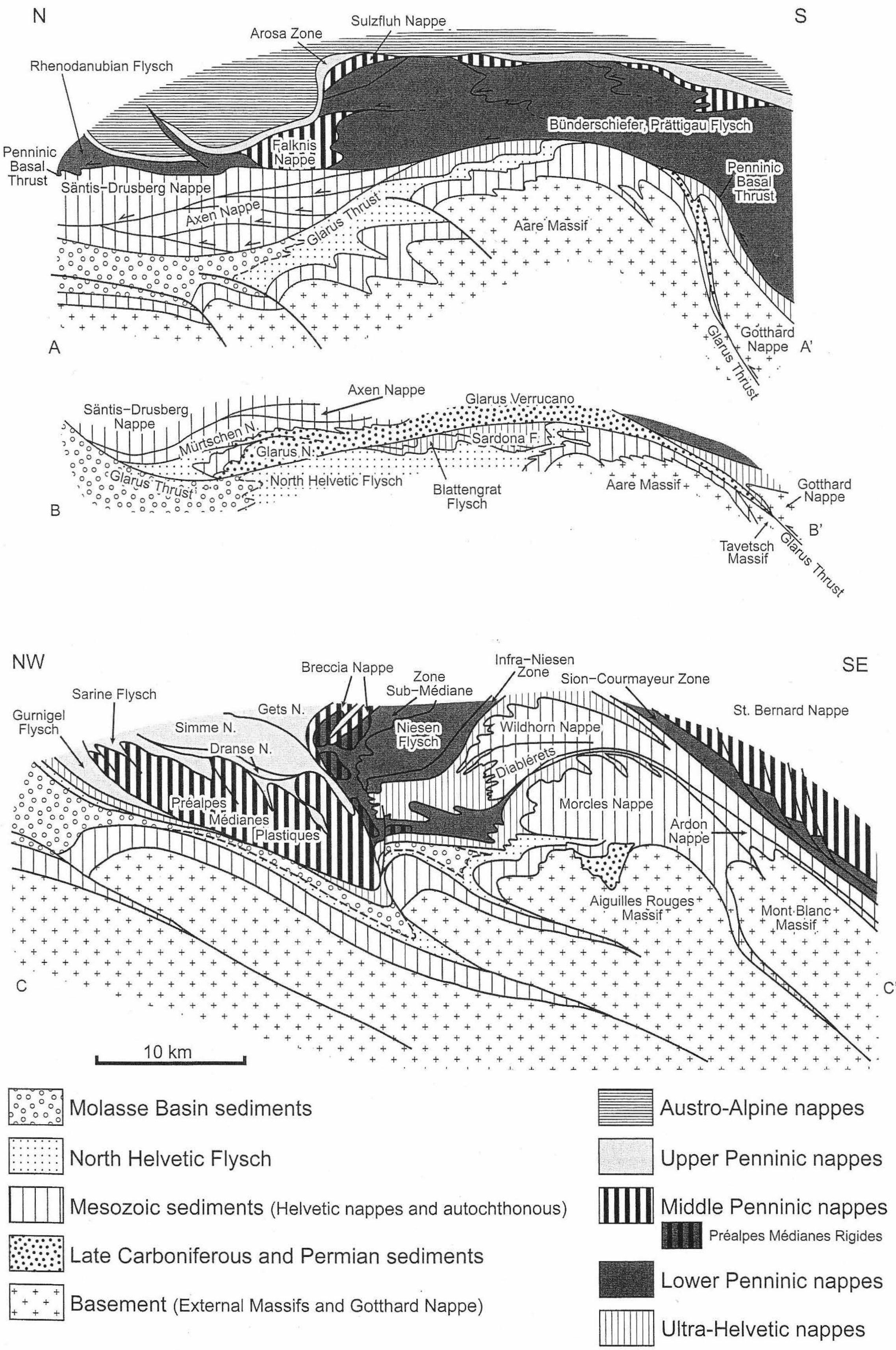



**Fig. 18.7.** Three profiles through the Helvetic Zone in Switzerland. After Schmid *et al.* (1996; profile A–A′), Schmid (1975; profile B–B′) and Escher *et al.* (1997, profile C–C′), modified. See Figure 18.5 for locations of cross-sections.

to the Helvetic nappes. It comprises basement (Combeynot Massif) and Triassic and Jurassic carbonates, unconformably overlying nummulitic limestone and Late Eocene (Priabonian) deep-marine clastics (Aiguilles d'Arves Flysch). In the older literature, the Aiguilles d'Arves Unit and the lowermost of the Lower Penninic nappes, the Cheval Noir Unit, were termed 'Ultradauphinois'.

*Ultra-Helvetic nappes*

This term encompasses a group of tectonic units which were structurally situated above the Helvetic nappes after initial thrusting, but which have locally become enveloped under Helvetic nappes in the course of out-of-sequence thrusting and nappe refolding. The palaeogeographic origin of these nappes was the Cretaceous continental rise between the Valais Ocean to the SE and the European shelf (Helvetic units) to the NW. Therefore, the facies of the Cretaceous sediments in the Ultra-Helvetic nappes is, in general, more pelagic (marly) than in the Helvetic nappes. Their youngest sediments are Eocene to Early Oligocene deep-marine clastics. Locally, for example, in western Switzerland, the Ultra-Helvetic units represent a mélange of various blocks of older rocks in a matrix of these clastic sediments. The Ultra-Helvetic units were emplaced on the Helvetic realm, that is, the later Helvetic nappes to the south and the autochthonous/parautochthonous cover of the External massifs to the north, by early thin-skinned thrusting during the Oligocene, before the Helvetic nappes formed. When these formed, the Ultra-Helvetic units acted like the 'youngest sediment layer' and became enveloped between and below the Helvetic nappes. In eastern Switzerland, such enveloping beneath the Glarus Thrust (i.e. the basal thrusts plane of the Helvetic Glarus Nappe) affected not only the Ultra-Helvetic Sardona Flysch but also the Blattengrat Flysch, the origin of which is assumed to be in the southern part of the subsequently-formed Helvetic nappes (Fig. 18.7, profile B–B′).

The Gotthard Nappe has only thin remnants of its own cover. The main part of the present sedimentary cover of this nappe is allochthonous and in an upside-down position (Etter 1987). This allochthonous Gotthard cover can be regarded as belonging to the Ultra-Helvetic nappes (Fig. 18.7, profile B–B′). It was emplaced on the Gotthard Nappe in an event of north-directed 'cover substitution' thrusting, when the original cover of the nappe, which is now part of the Helvetic nappes, was detached and replaced by the Ultra-Helvetic sediments.

*Helvetic and Ultra-Helvetic nappes along the front of the Eastern Alps*

Helvetic units cropping out along the northern front of the Alps in western Austria (Vorarlberg) and Bavaria form the eastward continuation of the Säntis-Drusberg Nappe of eastern Switzerland. Deeper Helvetic nappes are not exposed but were encountered in the Hindelang 1 borehole in western Bavaria (Schwerd *et al.* 1995), namely the Hindelang Nappe and the still deeper Hohenems Nappe. In addition, another Helvetic imbricate, the Grünten Nappe, locally occurs at the surface above the Säntis-Drusberg Nappe. The sequence is then, from base to top: Hohenems, Hindelang, Säntis-Drusberg and Grünten nappes (Schwerd *et al.* 1995). The Helvetic units comprise sediments of Upper Jurassic to Middle Eocene age, detached from the older rocks and thrust northward over the fill of the Molasse Basin. In Vorarlberg and western Bavaria, the Helvetic Säntis-Drusberg Nappe is overlain by the Ultra-Helvetic Liebenstein Nappe which includes sediments of Aptian to Middle Eocene age, predominantly marls and pelagic limestones. The basal thrust of the Liebenstein Nappe is folded together with the sedimentary succession of the Säntis-Drusberg Nappe.

True Helvetic units with the shelf facies occur approximately as far east as Salzburg,. Further to the east, the shelf facies is replaced in the Helvetic units by a more basinal facies resembling the Ultra-Helvetic nappes of Switzerland, with pelagic marls (Buntmergel) of Late Cretaceous to Early Tertiary age. These units have the structural position of the Helvetic nappes but exhibit the facies of the Ultra-Helvetic nappes. This situation arises because the thrust front of the Helvetic nappes is oblique to the facies boundaries. To the west, the thrust front formed on the shelf, and to the east, on the continental rise. The pre-Late Cretaceous succession of these units is developed in the so-called Gresten Facies with sandy, shallow-marine sediments of Jurassic age.

Between Salzburg and Vienna, the Helvetic units – in 'Ultra-Helvetic' facies – form an imbricate fan together with the tectonically overlying nappes of the Rhenodanubian Flysch. Initially, the Rhenodanubian Flysch was emplaced from the south onto the Helvetic units, and then onto the imbricate fan formed by northward thrusting over the fill of the Molasse Basin. The Helvetic units are exposed in the immediate hanging-walls of the individual thrusts, from whence they disappear towards the south under the Rhenodanubian Flysch.

*Sub-Penninic nappes*

The Sub-Penninic nappes (Milnes 1974) comprise elements of the distal European continental margin, dominated by basement but also including Permo-Mesozoic cover rocks. In contrast to the External massifs, the Sub-Penninic nappes have been completely detached from the lithospheric mantle. Sub-Penninic nappes occur in the Lepontine area of the Central Alps and in the Tauern Window of the Eastern Alps. In contrast to the original definition of Milnes (1974) which excluded the Adula Nappe, this nappe is now also termed 'Sub-Penninic' (Schmid *et al.* 2004*a*). We also include the Monte Rosa, Gran Paradiso and Dora-Maira nappes because they are in a similar structural position to the Adula Nappe, representing the most distal part of the European margin (Gebauer 1999; Froitzheim 2001).

The most external parts of the Sub-Penninic units, e.g. the Gotthard Nappe, formed the basement from which the Ultra-Helvetic and parts of the Helvetic cover nappes were detached. The lower Sub-Penninic nappes are eclogite-free whereas the uppermost and most internally derived nappes (Adula, Monte Rosa, Dora-Maira) were affected by eclogite-facies, partly ultra-high-pressure metamorphism during the Eocene.

The classical gneiss nappes of the Lepontine Dome (Fig. 18.2) belong to the Sub-Penninic nappes. The bulk of these nappes contain no eclogites. The uppermost tectonic unit, the Adula Nappe, however, includes Alpine eclogites and garnet peridotites (Heinrich 1986). The lower, eclogite-free nappes are the Simano, Lucomagno-Leventina, Maggia, Antigorio, Lebendun and Monte Leone nappes. The Gotthard Nappe (former Gotthard Massif) is the structurally deepest Sub-Penninic nappe in the area and crops out to the north of the other units (Figs 18.2 & 18.6). The Maggia Nappe has sometimes been correlated with the Briançonnais units because it locally overlies the Adula Nappe (e.g. Froitzheim *et al.* 1996; Schmid *et al.* 2004*a*), but other authors (Grujic & Mancktelow 1996; Maxelon 2004) explained this geometry as being related to post-nappe recumbent folding and they correlated the Maggia Nappe with the Sub-Penninic nappes, an interpretation followed herein. Towards the SW, the Moncucco and Camughera (Fig. 18.8) nappes also belong, in our view, to the Sub-Penninic system.

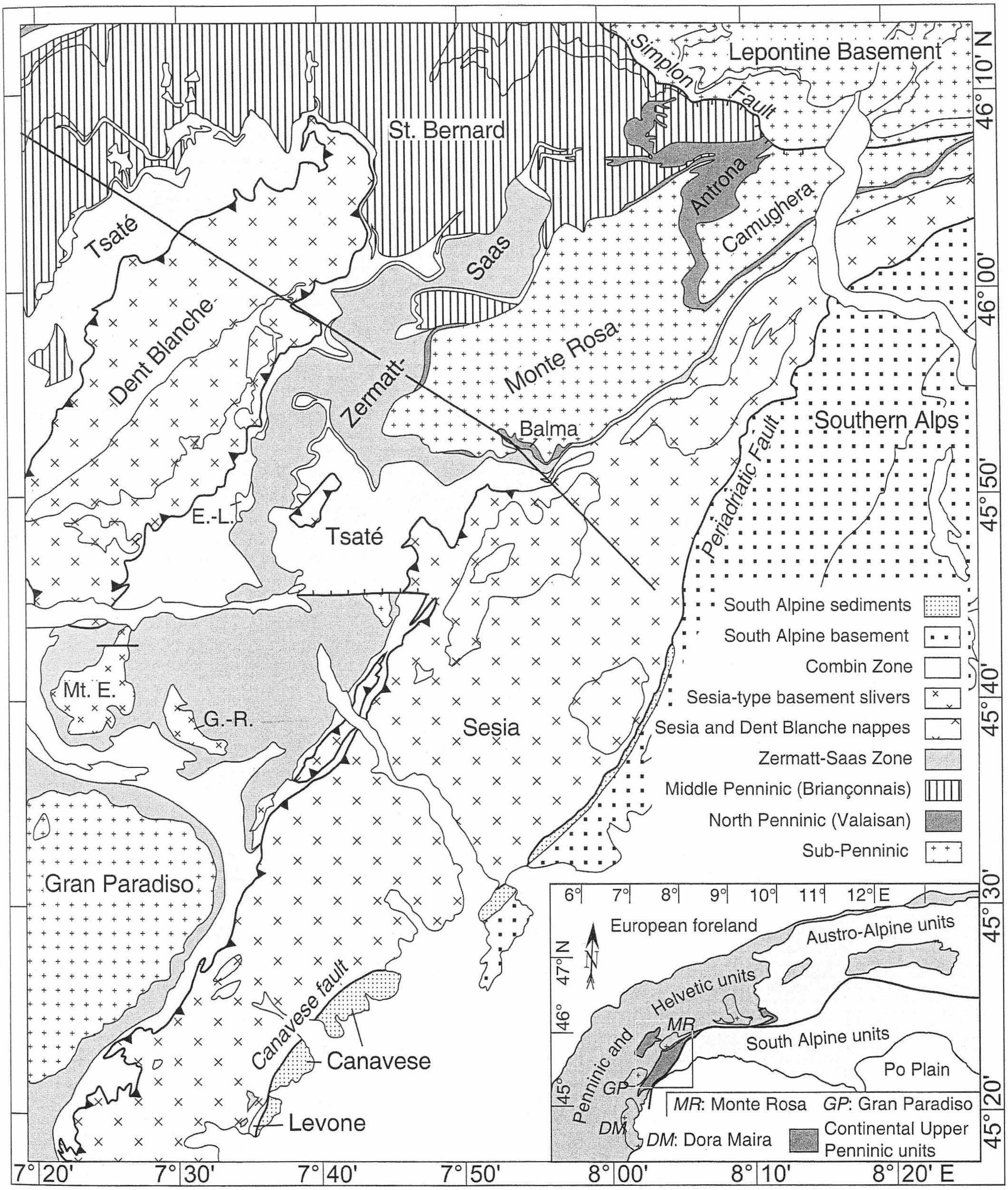



**Fig. 18.8.** Tectonic map of the Penninic Alps. Ophiolites from the Piemont-Ligurian Ocean occur in a deeper level (Zermatt-Saas Zone) and a higher level (Tsaté Nappe). In between are slivers of basement (E.-L., Etirol-Levaz; G.-R., Glacier-Rafray; Mt. E., Monte Emilius) and sedimentary rocks (C.B., Cimes Blanches Nappe), interpreted here to have been derived from the Cervinia terrane. A second, much thicker thrust sheet derived from Cervinia is on top of the Tsaté nappe, forming the Dent Blanche and Sesia nappes. It was emplaced by a NW-directed out-of-sequence thrust. Trace of cross-section in Figure 18.9 is indicated. After Steck *et al.* (1999) and Pleuger *et al.* (2007).

The eclogite-free Sub-Penninic nappes of the area are recumbent folds with a basement core (orthogneiss, paragneiss, amphibolite) and a thin sedimentary cover. The latter forms nappe-separating synclines pinching out mostly towards the south. Ophiolites and associated 'Bündnerschiefer' derived from the Valais Ocean also locally occur in these synclines. They were thrust over the Sub-Penninic units before the recumbent folds formed. As the folding occurred after initial thrusting, the stacking order of the recumbent folds, in many cases, does not reflect their original palaeogeographic arrangement, which led to the controversy concerning the origin of the Maggia Nappe.

The Adula Nappe (Figs 18.2 & 18.6) is the uppermost Sub-Penninic unit of the Lepontine Dome. It is not a coherent basement sheet but a stack of thin basement sheets separated by even thinner layers of Mesozoic quarzites, marbles and calcareous schists. Abundant lenses and boudins of eclogite and, in the southern part, garnet peridotite are also found in the Adula Nappe. For the Cima Lunga Unit, i.e. the southwestern part of the Adula Nappe, an Alpine age of high-pressure equilibration (40 to 35 Ma) is well established (Becker 1993; Gebauer 1996). A garnet peridotite body at Monte Duria in the southeastern Adula Nappe was rapidly exhumed from eclogite-facies conditions at 34 to 33 Ma (Hermann *et al.* 2006). An additional, pre-Alpine high-pressure event may have affected part of the Adula Nappe (Biino *et al.* 1997). An early Palaeozoic age was determined for some eclogite protoliths in the Adula Nappe (Santini 1992).

*Monte Rosa, Gran Paradiso and Dora-Maira nappes*

The palaeogeographic origin of these units is controversial. We have assigned them to the Sub-Penninic nappes (Gebauer 1999; Froitzheim 2001) whereas many other authors have favoured an origin from Briançonnais crust (Escher *et al.* 1997; Schmid *et al.* 2004*a*). The Monte Rosa Nappe (Figs 18.8 & 18.9) comprises para- and orthogneisses, the latter representing Carboniferous and Permian intrusions into the paragneisses. Albite-rich micaschists at the northern rim of the Monte Rosa Nappe may represent remnants of the Permo-Carbonferous cover, although generally, the cover was sheared off during subduction. Metabasic boudins in the paragneiss of the Monte Rosa Nappe record eclogite-facies metamorphism, the age of which is Eocene (*c.* 42 Ma; Lapen *et al.* 2007). In the interpretation outlined here, the Monte Rosa Nappe is a post-nappe recumbent anticline similar to the other Sub-Penninic nappes, and the Antrona Ophiolites that occur below the Monte Rosa Nappe do not represent an oceanic suture but are situated in a post-thrusting synform, the Antrona synform (Fig. 18.9; Froitzheim 2001). The geometric analysis of the Lepontine nappes by Maxelon (2004) demonstrates that the Monte Rosa Nappe is at the same structural level of the nappe stack as the Adula Nappe, which would support our interpretation.

The Gran Paradiso Nappe is very similar to the Monte Rosa Nappe in most respects. It was affected by Alpine eclogite-facies metamorphism in the Eocene (Meffan-Main *et al.* 2004). In contrast to the Monte Rosa Nappe, the Gran Paradiso Nappe has a Mesozoic metasedimentary cover. The Dora-Maira Nappe is composed of several sheets of Variscan basement (ortho- and paragneisses and metabasites) separated by Permo-Mesozoic metasediments and locally minor ophiolites. The nappe experienced blueschist- to eclogite-facies metamorphism, and one of the basement sheets even reached ultrahigh-pressure (UHP) conditions. Indeed, this is the famous coesite locality (Chopin 1984). UHP metamorphism is dated at *c.* 35 Ma (Gebauer *et al.* 1997). The Pinerolo Unit is exposed in a window beneath the Dora-Maira Nappe and contains metamorphic Permo-Carboniferous sediments in a facies similar to the Briançonnais units, and late Variscan granitic to dioritic bodies. The Pinerolo Unit underwent only blueschist metamorphism. The same lithological association, as in the Pinerolo Unit, also crops out in the Money Window, a small tectonic window through the Gran Paradiso Nappe (Compagnoni *et al.* 1974). If these rocks indeed belong to the Middle Penninic (Briançonnais) nappes, our interpretation implies major out-of-sequence thrusting of the 'European' Gran Paradiso and Dora-Maira nappes over the more internally derived Briançonnais nappes.

*Sub-Penninic nappes in the Tauern window*

Of the three large tectonic windows where Penninic units crop out in the Eastern Alps (Engadine Window, Tauern Window,

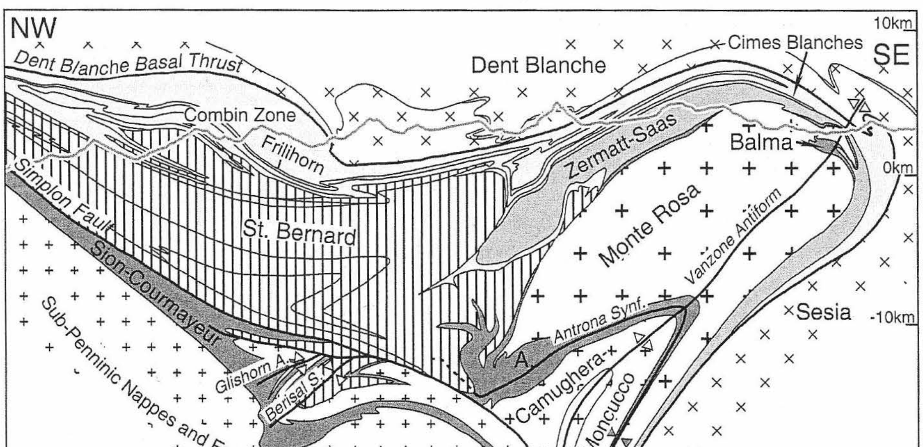



**Fig. 18.9.** Cross-section of the Penninic Alps. Note intensive, polyphase refolding of nappes and out-of-sequence thrusting, leading to the situation that originally deeper, more externally derived nappes (Monte Rosa Nappe, Dent Blanche Nappe) overlie originally higher, more internally derived nappes (Antrona Ophiolites and Tsaté Nappe, respectively). Glishorn A., Glishorn Antiform; Berisal S., Berisal Synform. For legend and location of section see Figure 18.8. After Escher *et al.* (1993) and Pleuger *et al.* (2007).

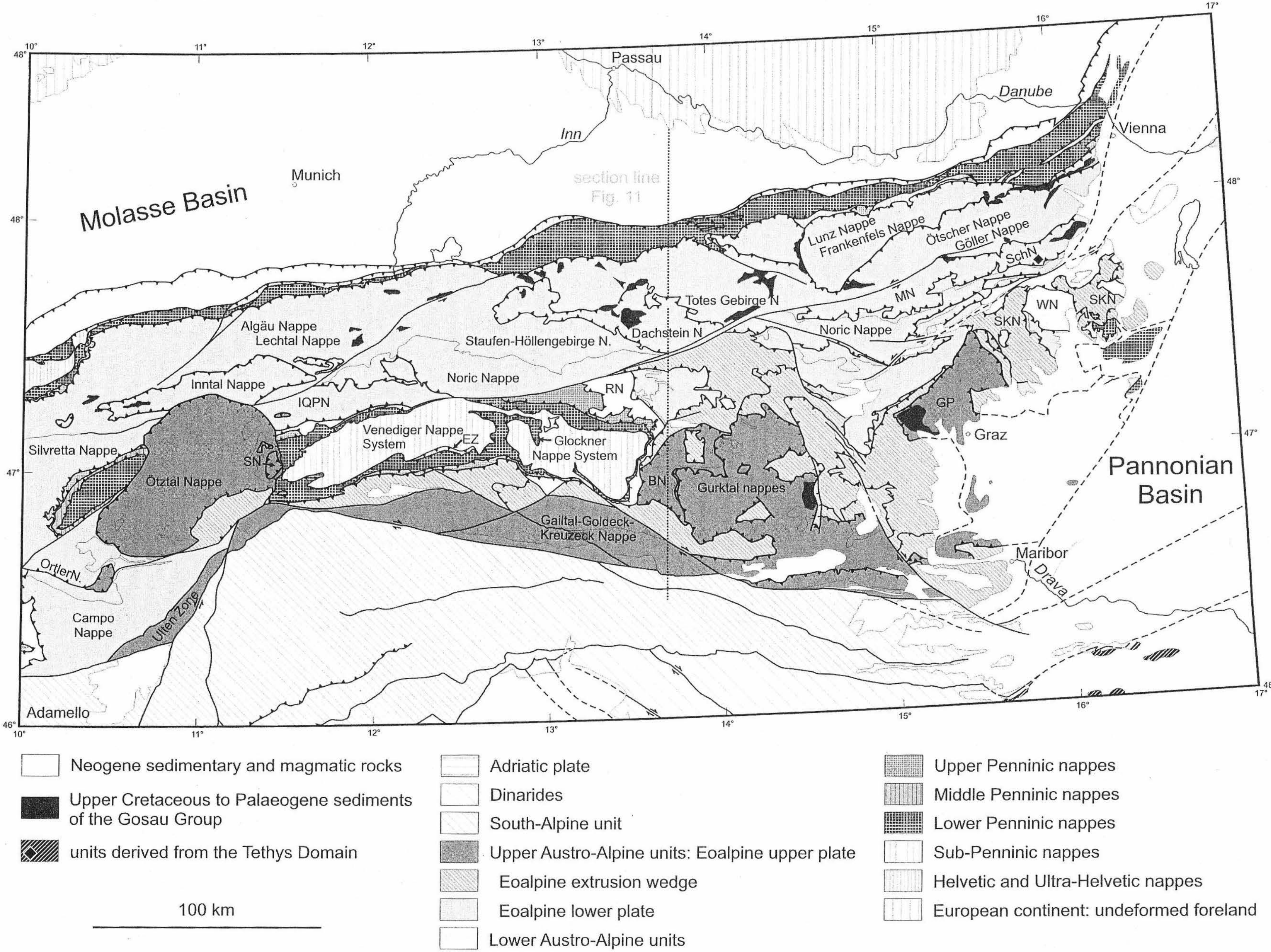



**Fig. 18.10.** Tectonic units of the Eastern Alps. Abbreviations: BN, Bundschuh Nappe; EZ, Eclogite Zone; GP, Graz Palaeozoic nappes; IQPN, Innbruck Quartz Phyllite Nappe; MN, Mürzalpen Nappe; RN, Radstadt nappes; SchN, Schneeberg Nappe; SKN, Stuhleck-Kirchberg Nappe; SN, Steinach Nappe; WN, Wechsel Nappe.

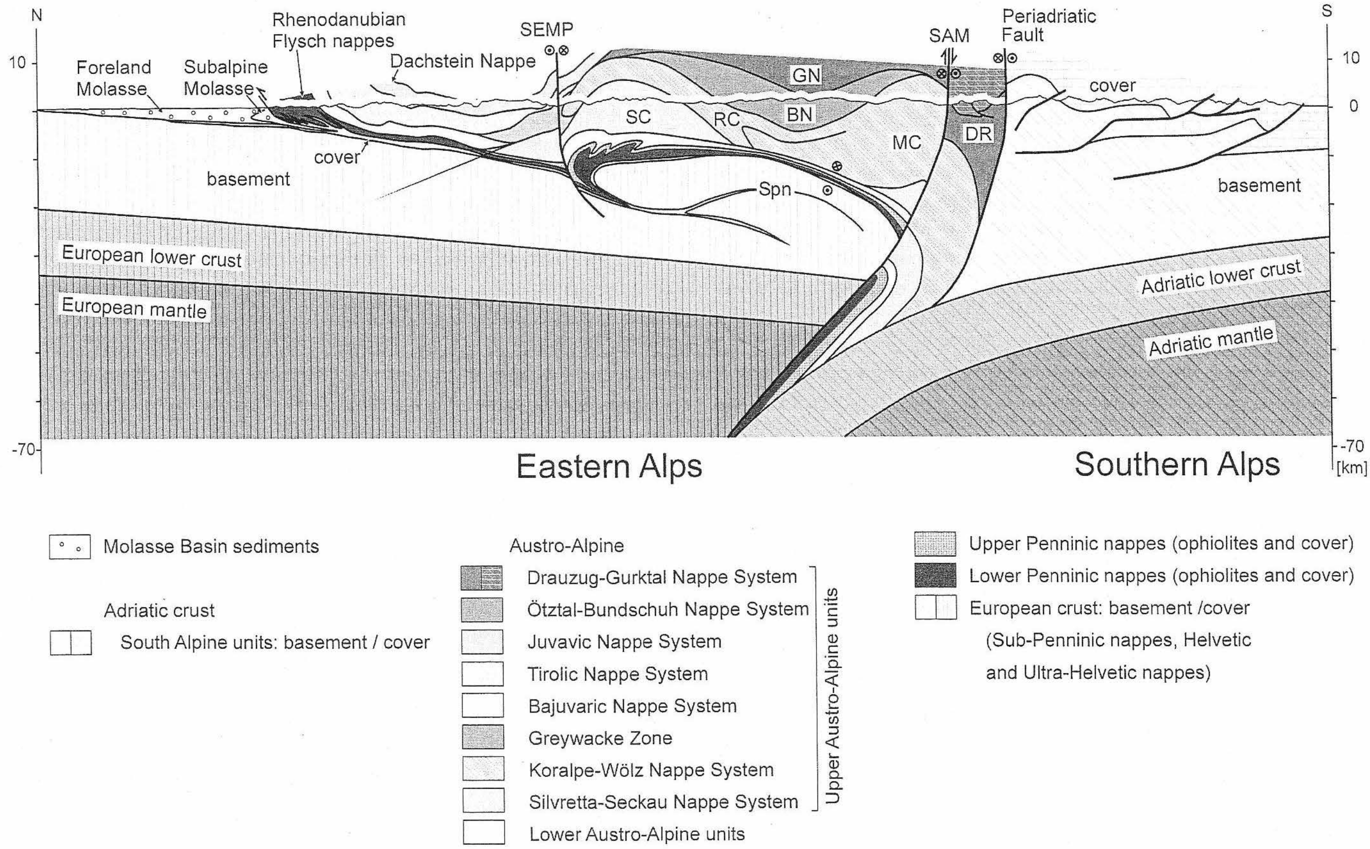



**Fig. 18.11.** Cross-section through the Eastern Alps. Abbreviations: BN, Bundschuh Nappe; DR, Drau Range; GN, Gurktal Nappe; MC, Millstatt Complex; RC, Radenthein Complex; SAM, Southern limit of Alpine metamorphism, coinciding with the Mölltal Fault in this cross-section; SC, Seckau Complex; SEMP, Salzach-Ennstal-Mariazell-Puchberg Fault; SPN, Sub-Penninic nappes. SEMP and Periadriatic Fault accommodate eastward extrusion of crustal units in the central part of the profile. Neogene switch in subduction polarity has led to north-dipping subduction zone at depth.

Rechnitz Window Group; Fig. 18.1), only the Tauern Window exposes Sub-Penninic nappes. These are represented by the Venediger Nappe System (Fig. 18.10 & 18.11; Frisch 1976), the lowermost unit exposed in the Tauern Window. It comprises continental crust from the European margin: Variscan granitoids (Zentralgneis) intruded into partly migmatized, pre-existing Palaeozoic sequences (e.g. Habach Complex) and possibly even older country rocks, and Permo-Mesozoic cover rocks. Within the nappe system a transition in the sediment facies and degree of allochthony can be recognized: in the palaeogeographically northern parts, Upper Jurassic carbonates represent a shelf environment and are still partly attached to thick basement. For example, the large gneiss nappes of the Tux and Ahorn massifs are covered by the Upper Jurassic Hochstegen Marble, and the Hochalm Massif in the eastern part of the Tauern Window is covered by the Silbereck Marble (Höfer & Tichy 2005). In contrast, successions derived from the slope towards the oceanic basin in the south have been sheared off completely, or together with thin basement lamellae, and displaced towards the north. One of these is the Wolfendorn Nappe. It was derived from the proximal slope and is characterized by a Triassic succession overlain by the Hochstegen Marble and Cretaceous clastic sediments including black shales, quartzites and meta-arkoses (Kaserer Formation). The cover sequences of the Seidlwinkl Nappe (Cornelius & Clar 1939) and the Rote Wand-Modereck Nappe (Kober 1922) were deposited on the distal slope. These include Permian siliciclastic rocks (Wustkogel Formation; lying on a Variscan basement lamella in the Rote Wand-Modereck Nappe), Triassic shallow-marine sedimentary rocks (Seidlwinkl Formation) and Jurassic to Cretaceous phyllites and carbonatic quartzites (Brennkogel Formation). The central part of the Venediger Nappe System (central with respect to the boundaries of the Tauern Window) was metamorphosed under amphibolite-facies conditions, whereas the outer parts reached greenschist-facies conditions during Tertiary metamorphism ('Tauernkristallisation') (Schuster *et al.* 2004). Typical cooling ages of the Venediger Nappe System range from 30 to 17 Ma (Oxburgh *et al.* 1966; Frank *et al.* 1987).

The Eclogite Zone (Fig. 18.10) is a south-dipping thrust sheet near the southern border of the central Tauern Window, intercalated between the 'Zentralgneis' and its Mesozoic cover to the north, and the Rote Wand-Modereck Nappe to the south. It is either a mixed or a transitional continental-oceanic unit. It comprises eclogite-facies metabasites, serpentinites and metamorphosed sedimentary rocks of probable Permian to Late Cretaceous age (quartzite, marble, calcareous schist). The age of the metabasite protoliths is not known. The eclogite metamorphism is probably Eocene in age (Zimmermann *et al.* 1994; Ratschbacher *et al.* 2005) and pre-dates the 'Tauernkristallisation'. Kurz *et al.* (1998) assumed that the Eclogite Zone was derived from the continent–ocean transition, thrust northward over the Sub-Penninic nappes, and then buried under the Rote Wand-Modereck Nappe by an out-of-sequence thrust.

The internal structure of the Venediger Nappe System reflects a polyphase evolution (Braumüller & Prey 1943; Kurz *et al.* 2001*b*). During the period of subduction in the Eocene, the structurally higher nappes of the system were formed from the distal margin and thrust over the more proximal parts, together with the units of the overlying Glockner Nappe System derived from the Penninic Ocean (see below). Parts of the Glockner and Venediger nappe systems were isoclinally interfolded during this process (e.g. Seidlwinkl Nappe and Glockner Nappe *sensu stricto*). In the Miocene the entire nappe pile was refolded by an east–west trending fold system with steeply dipping axial planes, and at the same time stretched east–west. Stretching became progressively localized in a west-dipping normal fault at the western end of the Tauern Window (Brenner Fault; Behrmann 1988; Selverstone 1988) and an east-dipping one at the eastern end (Katschberg Fault; Genser & Neubauer 1989). This period of deformation was related to the indentation of the South-Alpine block into the Eastern Alps (Rosenberg *et al.* 2004).

In the Engadine Window, the Sub-Penninic nappes are not exposed but occur at depth under the Lower Penninic (Valaisan) Bündnerschiefer, according to reflection-seismic interpretation (Pfiffner & Hitz 1997).

To summarize, the Sub-Penninic nappes represent the southern continental margin of the European plate which was subducted in the Eocene. Accordingly, Eocene high-pressure and ultrahigh-pressure metamorphic rocks occur in the originally most distal parts of the continental margin which were subducted to the greatest depths and now form the structurally highest nappes (from east to west): the Eclogite Zone of the Tauern Window, the Adula Nappe, the Monte Rosa Nappe, the Gran Paradiso Nappe, and the Dora-Maira Nappe. The Ultra-Helvetic nappes represent parts of the sedimentary cover of the Sub-Penninic nappes which were sheared off and frontally accreted to the orogenic wedge.

## *Lower Penninic (Valaisan) nappes*

### *Valaisan nappes in the Central Alps*

Sedimentary rocks and ophiolites from the Valais Ocean crop out in a narrow zone from Graubünden in the NE to Savoie in the SW. To the NE, these units plunge under the Austro-Alpine nappes and reappear in the Engadine Window. To the SW, the Valais units pinch out near the town of Moutiers in Savoie. In the Lepontine Dome area, the Valaisan rocks interdigitate with the Sub-Penninic nappes, due to the post-thrusting recumbent folding. The typical sediments of the Valais Ocean are calcareous schists of Cretaceous age ('Bündnerschiefer'). These pass upwards into calcareous to siliciclastic turbidites in the latest Cretaceous and Early Tertiary. The Valais Ocean floor (serpentinite, gabbro and basalt) is preserved as isolated slivers in the Bündnerschiefer and locally as coherent ophiolite bodies several kilometres in length.

The suture of the Valais Ocean, between the Middle Penninic (Briançonnais) nappes above and the Sub-Penninic nappes below, is exposed in two areas, to the west and east of the Lepontine area. In the east, it comprises the east-dipping Misox Zone (Fig. 18.2) above the Adula Nappe. At the same structural level, the Chiavenna ophiolites are exposed farther to the south (Fig. 18.2). Oceanic gabbros in the Chiavenna ophiolite were dated at *c.* 93 Ma (Liati *et al.* 2003). A metagabbro in the Misox Zone was dated at 161 Ma (Liati *et al.* 2005). Hence, both Jurassic and Cretaceous ocean floor existed in the Valais Ocean (Fig. 18.4), as predicted by Stampfli (1993). The Jurassic seafloor originally belonged to the Piemont-Ligurian Ocean and was captured in the Valais Ocean when the Briançonnais fragment rifted away from Europe during the Earliest Cretaceous and moved to the east. This implies renewed rifting of Jurassic oceanic lithosphere (Fig. 18.3; Liati *et al.* 2005).

To the west, the suture of the Valais Ocean is represented by the eclogite-facies (Colombi & Pfeifer 1986) Antrona Ophiolites. These are folded-in below the Monte Rosa Nappe, but were originally at a higher structural level (Fig. 18.9). The Valais suture can be traced from the Antrona Ophiolites through discontinuous slivers at the front of the Monte Rosa Nappe towards the SW, and over the top of the Monte Rosa Nappe towards the SE into the Balma Unit (Pleuger *et al.* 2005; Fig. 18.9). Protolith ages of metagabbros are between 155 and 163 Ma in the Antrona Ophiolites (Liati *et al.* 2005), and *c.* 93 Ma in the Balma Unit (Liati & Froitzheim 2006). Hence, in this part of the Valaisan suture both Jurassic and Late Cretaceous ocean floor occurs.

In the French part of the Valaisan, eclogite-facies metabasics occur in the Versoyen Complex. A Late Carboniferous age (*c.* 309 Ma) for a leucogabbro dyke in the Versoyen Complex (Schärer *et al.* 2000) was interpreted as indicating a Palaeozoic age for the ophiolite complex. This is in marked contrast to the Jurassic and Cretaceous ages from the Swiss-Italian Valais ophiolites (see above). In the Préalpes klippe, the Valaisan is represented by the Zone Submediane, a mélange zone below the floor thrust of Briançonnais-derived nappes, and by the underlying Niesen Nappe with thick Late Cretaceous to Eocene deep-marine clastics (Niesen Flysch; Fig. 18.7).

### *Valaisan nappes in the Engadine Window*

Rocks from the Valais Ocean are exposed in the central part of the Engadine Window. The lowermost unit in the window is the Pfunds Zone, a thick pile of probably Cretaceous calcareous schists (Bündnerschiefer) with intercalated ophiolites (e.g. at Piz Mundin; Oberhauser 1980). The Pfunds Zone may be subdivided into a lower part (Mundin Unit) containing blueschist-facies assemblages (Mg-carpholite and glaucophane) and a higher part (Arina Unit) which experienced lower greenschist-facies metamorphism (Bousquet *et al.* 1999). The overlying tectonic mélange including Bündnerschiefer, deep-marine clastics, and ophiolite slices is referred to as the Roz-Champatsch-Pezid Zone (Oberhauser 1980, 1998). Slivers of Mesozoic sedimentary rocks, in a typical Austro-Alpine facies, are tectonically intercalated at the boundary between the Pfunds and the Roz-Champatsch-Pezid zones (Stammerspitze). In the southwestern part of the Engadine Window, the Roz-Champatsch-Pezid Zone is overlain by the Prutz-Ramosch Zone which includes pillow basalts and fragments of serpentinized mantle lherzolite. These are, in turn, overlain by a sheet of granitic/gneissic basement with a cover of Triassic to Palaeogene-age sediments (Tasna Nappe, derived from the Briançonnais fragment). Florineth & Froitzheim (1994) suggested that Lower Cretaceous sediments of the Tasna Nappe onlap both the granitic/gneissic basement and a lherzolite body belonging to the Prutz-Ramosch Zone (Piz Nair serpentinite). Hence, this location preserves a former ocean–continent transition between the Briançonnais continental crust to the south and the Valais Ocean to the north (Florineth & Froitzheim 1994.). These relationships support the assumption of Early Cretaceous opening of the Valais Ocean kinematically linked with the opening of the Bay of Biscay and transtensional basin formation in the Pyrenees (Frisch 1979; Stampfli 1993; see Manatschal *et al.* 2006 for a different interpretation). Towards the NE, the Tasna Nappe is replaced by an Alpine tectonic mélange, the Fimber Zone (Oberhauser 1998).

### *Middle Penninic (Briançonnais-derived) nappes*

The Briançonnais fragment was a block of continental crust, wider in the SW and tapering out to the NE, that existed during the Cretaceous between the Piemont-Ligurian Ocean and the Valais Ocean (Fig. 18.3). Frisch (1979) and Stampfli (1993) interpreted the Briançonnais fragment as a prolongation of the Iberian continent into the Alpine realm, a hypothesis that is today widely accepted. The southward pinching-out of the Valais units in France may be explained by omission through Alpine faults (Fügenschuh *et al.* 1999) or, alternatively, may reflect the primary termination of the oceanic basin. The term Briançonnais, as used here, comprises not only the Briançonnais *sensu stricto* as exposed in the type area, but also the Sub-Briançonnais units derived from a Jurassic to Cretaceous basin on continental crust along the northwestern margin of the Briançonnais fragment. It also comprises the Pre-Piemontais units. These sedimentary successions were probably deposited on the SE margin of the Briançonnais fragment (although some of the Pre-Piemontais units may have been derived from a more internal continental fragment, Cervinia). The Middle Penninic nappes are partly sedimentary nappes, partly composite sediment/basement nappes. The sedimentary successions are characterized by Early to Middle Jurassic rift tectonics related to the opening of the Piemont-Ligurian Ocean, leading to the emergence of the Briançonnais *sensu stricto* as the rift shoulder and the formation of the Sub-Briançonnais rim basin NW of it (Stampfli *et al.* 1998). Rifting and opening of the Valais Ocean are also recorded in the sediments (e.g. Tithonian-age, rift-related Falknis Breccia in the Falknis Nappe of eastern Switzerland).

The Middle Penninic nappes escaped Alpine eclogite-facies metamorphism, except in the most internal Briançonnais units of the Western Alps (Acceglio Unit; Schwartz *et al.* 2000). Briançonnais nappes in internal positions show a blueschist-facies overprint (e.g. the Suretta and Tambo nappes in eastern Switzerland and the Vanoise and Ambin units in the western Alps).

*Middle Penninic nappes in the eastern Central and Eastern Alps*

The Tambo Nappe and the overlying Suretta Nappe (Fig. 18.6) are basement nappes with a thin overlying cover of Permo-Triassic clastic sediments which follow an important hiatus with equally thin Cretaceous (or Jurassic?) breccias and calcareous schists. This sedimentary cover is, in the case of both the Tambo Nappe and the Suretta Nappe, overlain by an allochthonous Triassic to Cretaceous sedimentary succession, again showing an important hiatus in the Middle to Upper Triassic and Liassic (Baudin *et al.* 1995; Stampfli *et al.* 1998). This allochthonous cover, the Starlera Nappe, was emplaced on the Tambo and Suretta nappes by thin-skinned thrusting before the Suretta Nappe was thrust towards the north over the Tambo Nappe, leading to the duplication of the Starlera Nappe. All of this occurred during the Palaeogene. The Starlera Nappe is derived from the southeastern part of the Briançonnais fragment (Baudin *et al.* 1995) and the Suretta and Tambo nappes from a palaeogeographic high in the central part of the fragment. However, the existence of the Starlera Nappe is controversial because, due to the lack of fossils, the interpretation of Baudin *et al.* (1995) is based on lithostratigraphy alone. Other authors assume that the succession which forms the Starlera Nappe, according to Baudin *et al.* (1995), belongs rather to the autochthonous sediment cover of the Suretta and Tambo nappes (e.g. Schmid *et al.* 1996).

The Schams Nappes (Fig. 18.6), which are extremely deformed sediment nappes with various facies evolutions during the Jurassic and Cretaceous, were probably derived from the northwestern margin of the Briançonnais fragment (Schmid *et al.* 1990). They are probably rooted below the Tambo Nappe, at the top of the Misox Zone, and were folded back around the front of the Tambo and Suretta nappes by a north-closing recumbent fold (the Niemet-Beverin fold, Fig. 18.6; Milnes & Schmutz 1978; Schmid *et al.* 1990) of late Eocene to Early Oligocene age (35 to 31 Ma). Above the Suretta Nappe, a south-closing recumbent fold brings the Schams Nappes back into their proper position in the nappe stack, between the Piemont-Ligurian units above and the Valaisan units below (Schmid *et al.* 1990; Schreurs 1993; Weh & Froitzheim 2001). In this area, the Schams Nappes are omitted due to a top-to-the-east low-angle normal fault of the same age (35 to 31 Ma), the Turba Fault (Fig. 18.6; Nievergelt *et al.* 1996). Further to the north, where the Turba Fault dies out, sedimentary nappes equivalent to the Schams Nappes reappear at the same structural position, in the form of the Falknis Nappe and the Sulzfluh Nappe (Figs 18.2 & 18.7).

The Late Jurassic to Cretaceous sediment succession of the Falknis Nappe was deposited in a basin similar to that of the Sub-Briançonnais. After deposition of the youngest sediments in the Palaeocene to possibly Early Eocene (Allemann 1957), the sedimentary succession was spectacularly folded. Two generations of folds may be distinguished. The Sulzfluh Nappe overlies the Falknis Nappe and is derived from a Late Jurassic carbonate platform south of the basin from which the Falknis Nappe was derived. It forms an antiformal stack of imbricated Upper Jurassic limestone sheets.

The Tasna Nappe in the Engadine Window (see above) had a sedimentary evolution similar to that of the Falknis Nappe. It was situated at the immediate transition from the Briançonnais continental fragment to the Valais Ocean (Florineth & Froitzheim 1994). No unequivocal Briançonnais elements occur further to the east, either in the Tauern Window or at the northern border of the Alps. Therefore it may be assumed that the Tasna Nappe marks the eastern end of the Briançonnais fragment where the Valais and Piemont-Ligurian basins merged into one Penninic Ocean.

*Middle Penninic nappes in the western Central Alps*

Briançonnais-derived nappes occur both in the internal part of the Alpine chain, in the Pennine Alps, and close to the front of the Alps in the Préalpes, a large outlier of the Penninic nappes, and minor klippen further to the east (e.g. the Mythen, Fig. 18.2). The main Briançonnais unit in the Préalpes is the Préalpes Médianes Nappe, also termed Klippen Nappe, comprising Triassic to Middle Eocene sediments. Its external part (Préalpes Médianes Plastiques) is deformed by large-scale thrust-related folds (Mosar *et al.* 1996; Wissing & Pfiffner 2002) whereas the internal part (Préalpes Médianes Rigides) is present as large fault-bounded blocks and slices (Fig. 18.7, profile C–C′). This reflects the different stratigraphies in each part. In the Préalpes Médianes Rigides, Jurassic platform carbonates rest with an erosional unconformity (due to the Middle Jurassic emersion) directly on Triassic ones. In the Préalpes Médianes Plastiques, in contrast, Lower and Middle Jurassic shale horizons exist and allow flexural flow. The Préalpes Médianes Plastiques are stratigraphically comparable to the Sub-Briançonnais in France, the Préalpes Médianes Rigides to the Briançonnais *sensu stricto* (Trümpy 1980). The Préalpes Médianes Nappe is unmetamorphosed in the NW changing to anchizonal in the area of the Préalpes Médianes Rigides. To the south, the Préalpes Médianes Rigides are overlain by the Breccia Nappe (Fig. 18.7, profile C–C′; Upper Triassic to Palaeocene) characterized by thick, rifting-related breccias of Jurassic age which may have been shed from fault scarps at the transition from the Briançonnais to the Piemont-Ligurian Ocean. The Breccia Nappe may be included in the Pre-Piemontais units.

In the Pennine Alps, the Briançonnais is represented by the St. Bernard Nappe (or Bernhard Nappe, Figs 18.8 & 18.9). This composite basement/sediment nappe has been subdivided into several sheets, including, from base to top and from originally external to internal (Escher *et al.* 1997; Sartori *et al.* 2006): (1) the Zone Houillière, a detached, coal-bearing Permo-Carboniferous graben fill, overlain by Triassic cover rocks; (2) the Ruitor Zone, a basement unit which is present only in the western part; (3) the Siviez-Mischabel Nappe, containing basement with a cover ranging from Late Carboniferous to Eocene; and (4) the Mont Fort Nappe, comprising basement and Permo-Carboniferous to Liassic cover rocks. The cover of the Siviez-Mischabel Nappe is best preserved in the Barrhorn area where the typical Briançonnais facies was recognized by Ellenberger (1953). The metamorphism of the St. Bernard Nappe is in the greenschist facies, increasing from base to top. Only the Mont Fort Nappe shows blueschist-facies metamorphism. Deformation and metamorphism commenced in the Eocene.

*Middle Penninic (Briançonnais and Sub-Briançonnais) nappes in the Western Alps*

The Briançonnais of the Western Alps comprises basement and cover nappes. The main Variscan basement complexes are (from north to south) the Ruitor, Vanoise and Ambin 'massifs' (they are in fact completely allochthonous and underlain by Lower Penninic nappes). These are formed by paraschists and gneisses with Variscan and older, greenschist- to amphibolite-facies metamorphism (Desmons 1977). Similar Briançonnais basement complexes crop out further SE in the Ligurian Alps. The Briançonnais cover is still partly attached to the basement and partly forms detached cover nappes.

Thick Upper Carboniferous to Permian clastic units comprise the Zone Houillière, the continuation of the same zone described above for the western Central Alps. Lower Triassic sandstones are overlain by Triassic platform carbonates with intercalated evaporites. During the Liassic to Middle Jurassic, the platform broke up into fault blocks along rift-related normal faults and was uplifted, partly above sea level, as evidenced by karst and bauxite formation (Faure & Mégard-Galli 1988; Goffé 1977). Large parts of the prerift cover were eroded. In the Bathonian, the area subsided again below sea level and during the Late Jurassic it was covered by pelagic limestones.

In the internal part of the Briançonnais, sedimentation continued until the Early Eocene; in the external parts, until the Late Eocene (Jaillard 1999). Breccia and olistolith deposition during the Late Cretaceous and earliest Tertiary, particularly in the internal part, provide evidence of tectonic activity at this time, but not yet leading to accretion of the nappes.

The most internal part of the Briançonnais in the Western Alps is the Acceglio Zone (= 'Ultra-Briançonnais'). The Alpine metamorphic grade is blueschist facies in the internal part (eclogite facies in the Acceglio Zone; Schwartz *et al.* 2000) to unmetamorphosed in the external part. Lemoine *et al.* (1986) interpreted the sedimentary nappes in the classic Briançon profile as assembled by simple east-over-west stacking and retrodeformed them accordingly. Claudel & Dumont (1999), however, found evidence for early north-directed thrusting and later west-directed out-of-sequence thrusting. Still later, the area was affected by dramatic top-to-the-east shearing and backfolding (Platt *et al.* 1989). The blueschist-metamorphic, Permo-Carboniferous metasediments of the Pinerolo Unit, underlying the Sub-Penninic Dora-Maira Nappe along its eastern contact, and similar rocks exposed in the Money Window below the Gran Paradiso Nappe, may also belong to the Briançonnais. As stated above, we assume a more external, European origin for the Gran Paradiso and Dora Maira nappes than for the Briançonnais. If the Pinerolo and Money units are indeed Briançonnais, they would have been overthrust by the Gran Paradiso and Dora-Maira nappes along out-of-sequence thrusts.

The Sub-Briançonnais underlies the Briançonnais *sensu stricto* along its western margin. This unit comprises unmetamorphosed Triassic and Jurassic carbonates and marls, detached along the Carnian evaporites. During the Early and Middle Jurassic, the Sub-Briançonnais represented a rim basin located to the NW of the Briançonnais rift shoulder on the northwestern margin of the Piemont-Liguria Ocean. In the Early Cretaceous it became part of the Iberia-Briançonnais microcontinent, separated from Europe by the Valais Ocean.

*Pre-Piemontais nappes*

The Pre-Piemontais nappes comprise several units that are generally found in two different structural positions: (1) between the Briançonnais (below) and Upper Penninic ophiolite nappes (above); and (2) intercalated between the Upper Penninic ophiolite nappes. The Breccia Nappe of the Préalpes is an example of the first situation, while an example of the second is the Cimes-Blanches Nappe (Fig. 18.9) at the boundary between the Zermatt-Saas Zone (below) and the Tsaté Nappe (above), two ophiolite-bearing Upper Penninic nappes (see below).

The Pre-Piemontais nappes are mostly formed by sedimentary successions originally deposited on, and detached from, continental basement. Some slivers of Variscan basement are also present, particularly in the Ligurian Alps, at the base of the Pre-Piemontais nappes (Seno *et al.* 2005). The Pre-Piemontais stratigraphy does not show the hiatus resulting from Early to Middle Jurassic uplift and erosion which is typical for the Briançonnais. Characteristic for the Pre-Piemontais are evaporites and dolomites in the Upper Triassic, Lower to Middle Jurassic rift breccias (e.g. in the Breccia Nappe), and Upper Jurassic radiolarites. The youngest sediments are early (to Middle?) Eocene turbidites (Seno *et al.* 2005).

It is generally assumed that the Pre-Piemontais units originated from the passive margin between the Briançonnais and the Piemont-Ligurian Ocean. However, this is probably not the case for all of the Pre-Piemontais units. In particular, the Cimes-Blanches Nappe was probably detached from Cervinia and emplaced by top-to-the-NW thrusting over the Zermatt-Saas Zone (Pleuger *et al.* 2007). It is, therefore, likely that the Pre-Piemontais is of heterogeneous origin, being partly derived from the internal Briançonnais margin and partly from Cervinia.

## Upper Penninic (Piemont-Ligurian) nappes

The Upper Penninic nappes include oceanic crust and sedimentary cover of the Piemont-Ligurian Ocean, also known as the Alpine Tethys. In addition, they include several continental nappes, the Sesia Nappe (= Sesia-Lanzo Zone), the Dent Blanche Nappe and the Margna Nappe (Upper Penninic continental nappes). These are derived from a continental fragment or microcontinent, Cervinia (Pleuger *et al.* 2007) or the Margna-Sesia fragment (Schmid *et al.* 2004*a*), within the Piemont-Ligurian Ocean. The Sesia and Dent Blanche nappes were originally treated as Penninic (Argand 1909) but later correlated with the Lower Austro-Alpine Err and Bernina nappes in the

Eastern Alps (Staub 1938) and, therefore, termed Austro-Alpine. More recent work has raised serious doubts about this correlation (e.g. Trümpy 1992; Froitzheim & Manatschal 1996; Froitzheim *et al.* 1996; Pleuger *et al.* 2007). Since the Penninic–Austro-Alpine boundary is defined as the top of the uppermost (palaeogeographically southernmost) Piemont–Ligurian ophiolites (Trümpy 1975), we adopt the view of Argand (1909) and treat the Sesia, Dent Blanche and Margna nappes as Upper Penninic continental nappes.

Oceanic Upper Penninic nappes are found all along the Alpine chain, from Liguria to eastern Austria. In the Central and Western Alps, radiometric ages of oceanic gabbros and their more acidic differentiates yielded ages of between $142 \pm 5$ and $166 \pm 1$ Ma, indicating oceanic spreading in the late Middle to Late Jurassic (Fig. 18.4; Liati *et al.* 2003, and references therein; Stucki *et al.* 2003). The Piemont-Liguria-derived nappes are of two types: ophiolite nappes and deep-marine clastic ('flysch') nappes. The characteristic sediment cover of the ophiolite nappes consists of radiolarites (Bathonian–Late Jurassic), Calpionella limestones (Tithonian–Berriasian) and Argille a Palombini Formation (shale with limestone layers; Neocomian), overlain by shales (often with ophiolite detritus), calcareous schists (Schistes lustrés) and various units of deep-marine clastics that are Late Cretaceous in age. Biostratigraphic ages from the radiolarite confirm that oceanic spreading began in the Bajocian (Bill *et al.* 2001).

The ophiolite nappes are subdivided in the Western Alps and western Central Alps into structurally deeper ones with eclogite-facies metamorphism (Voltri Ophiolites, Monte Viso Ophiolites, Lanzo Peridotite, Zermatt-Saas Zone) and higher ones with blueschist- to greenschist-facies metamorphism (Queyras area, Tsaté Nappe; Figs 18.2, 18.8 & 18.9). An important pressure gap occurs at the boundary of the two groups of nappes, between high pressure below and lower pressure above (Ballèvre & Merle 1993; Dal Piaz *et al.* 2001; Bousquet *et al.* 2004). (The Tsaté Nappe and the Cimes-Blanches Nappe are collectively termed the Combin Zone; see Fig. 18.8.) The uppermost ophiolite-bearing nappes in the Western Alps are unmetamorphosed (Montgenèvre or Chenaillet Ophiolite).

In the eastern Central Alps, Piemont-Ligurian ophiolites and their sediment cover form the greenschist-facies Platta Nappe. Towards the south, the Platta Nappe is prolongated by two layers of ophiolites, above and below the continental Upper Penninic Margna Nappe (see below). The structurally deeper ophiolite layer includes the Malenco Ultramafic Complex, comprising mostly mantle rocks exhumed by rifting (Fig. 18.6; Trommsdorff *et al.* 1993; Hermann *et al.* 1997). At a still deeper structural level, Piemont-Ligurian ophiolites also occur in the Avers Bündnerschiefer Nappe which underwent blueschist-facies metamorphism (Oberhänsli 1978). The Avers Bündnerschiefer Nappe is separated from the structurally higher Piemont-Ligurian ophiolite nappes by the Turba Fault, an east-dipping, Palaeogene normal fault (Nievergelt *et al.* 1996). Towards the north, the Platta Nappe continues into the Arosa Zone, a strongly dismembered unit along the base of the Austro-Alpine nappes which records a complicated history of Cretaceous and Tertiary subduction (Ring *et al.* 1988; Nagel 2006). The Arosa Zone contains Piemont-Ligurian ophiolites, their sedimentary cover, and slivers of basement and sediments of Austro-Alpine provenance. The Cenomanian–Turonian-age Verspala Flysch (part of the Arosa zone) contains abundant chromite, suggesting erosion of already accreted Piemont-Ligurian oceanic crust at that time (Burger 1978; Oberhauser 1983). The Arosa Zone is also exposed in the Engadine Window where it forms the uppermost structural unit. It comprises Jurassic-age ophiolites (e.g. Idalpe Ophiolite) with a cover of radiolarite, which experienced a blueschist-facies metamorphic imprint (Höck & Koller 1987).

The flysch nappes derived from the Piemont-Ligurian Ocean lack an ophiolite basement. They comprise the Helminthoid Flysch nappes, of Campanian to locally Middle Eocene age and including characteristic light-coloured, fine-grained calciturbidites ('Alberese'). Such nappes include the San Remo and Alassio nappes in the Ligurian Alps, the Parpaillon and Autapie nappes in the external part of the Western Alps, the Gurnigel Flysch Nappe in the frontal part of the Préalpes (Fig. 18.7), the Dranse Nappe on top of the Préalpes, and the Schlieren and Wägital Flysch nappes further to the east. (However, Trümpy (2006) proposed that the Gurnigel, Schlieren and Wägital nappes may be derived from the Valais Ocean.) Some of these nappes lie in a very external position, for example, the Parpaillon and Autapie nappes which travelled over the Briançonnais nappes transporting slivers of Briançonnais at their base. They were later overthrust by the main mass of the Briançonnais along a top-to-the-west out-of-sequence thrust. Some klippen of the Helminthoid Flysch nappes are also found in the Western Alps on the Briançonnais nappes themselves. The Gets Nappe, the uppermost unit of the Préalpes, consists of Cretaceous deep-marine clastics containing olistoliths of both oceanic (166 Ma gabbro; Bill *et al.* 1997) and continental provenance. The underlying Simme Nappe consists of chromite-bearing (ophiolite-fed) deep-marine clastics of Cretaceous age. Assigning certain depositional areas in the Piemont-Ligurian ocean basins to these tectonic units is still rather speculative. It is often assumed that the Helminthoid Flysch units were deposited close to the Apulian margin, since the Late Cretaceous calciturbidites required a carbonate shelf source area which can only have been the Apulian shelf at that time.

The Sesia and Dent Blanche nappes in the western Central Alps and the Margna Nappe in the Eastern Alps were derived from a continental fragment or microcontinent (i.e. Cervinia) between two branches of the Piemont-Ligurian Ocean. The Sesia Nappe and the Dent Blanche Nappe (Figs 18.8 & 18.9) represent the rear and frontal part, respectively, of one originally continuous thrust sheet. Both units lie on top of the Tsaté Nappe (Piemont-Ligurian ophiolites, see above). Both comprise three different types of basement: (1) Variscan basement with a Permian granulite-facies overprint (Valpelline Series in the Dent Blanche Nappe, Seconda Zona Dioritico-Kinzigitica or '2DK' in the Sesia Nappe); (2) Variscan upper-crustal basement; and (3) Permian gabbros. In addition, thin slivers of Mesozoic cover rocks are present. The Alpine-age overprint comprises eclogite-facies metamorphism in the Sesia Nappe, dated at *c.* 65 Ma (Rubatto *et al.* 1999), to blueschist-facies metamorphism in the Dent Blanche Nappe. Basement rocks with the typical characteristics of the Sesia and Dent Blanche nappes occur not only within these but also at deeper structural levels (Ballèvre *et al.* 1986), either immediately below the contact between the Zermatt-Saas Zone and the Tsaté Nappe (Fig. 18.8) or deeper within the Zermatt-Saas Zone. The slivers underwent eclogite-facies metamorphism, together with the Zermatt-Saas ophiolites, at 45 to 40 Ma (Dal Piaz *et al.* 2001).

Along the Periadriatic Fault, thin lenses of ophiolite occur between the Sesia Nappe to the NW and the Canavese Zone to the SE, e.g. at Levone (Fig. 18.8). The Canavese Zone represents the distal passive continental margin of the Apulian continent, characterized by rifting-related breccias in the Jurassic (Ferrando *et al.* 2004). This led several authors (Aubouin *et al.* 1977; Mattauer *et al.* 1987) to place the suture of the Piemont-Ligurian

Ocean between the South Alpine units and the Sesia Nappe, and not NW of Sesia. New structural work (Pleuger *et al.* 2007) suggests that the Tsaté Nappe represents an accretionary prism formed during the subduction of the internal basin of the Piemont-Ligurian Ocean, between Cervinia and the Apulian margin, and that the Zermatt-Saas Zone is a remnant of the external basin. Following closure of the internal basin, the Tsaté Nappe was thrust over the Sesia and Dent Blanche nappes and came to lie on the – already partly exhumed – Zermatt-Saas Zone. The Tsaté and Zermatt-Saas units were subsequently overthrust by the Sesia and Dent Blanche nappes along an out-of-sequence thrust.

The Margna Nappe at the western border of the Eastern Alps comprises the same lithological association as the Sesia and Dent Blanche Nappes (see above) but was only affected by elevated-pressure greenschist-facies metamorphism during the Alpine Orogeny (Liniger & Guntli 1988). It is sandwiched between Piemont-Ligurian ophiolites above (Platta Nappe) and below (Malenco Ultramafic Complex).

## *Penninic nappes in the Eastern Alps*

Since the Brianconnais probably ended in the Engadine Window, oceanic series found further east in the Tauern Window, the Rechnitz window group, and along the northern border of the Alps were derived from one large Penninic ocean. A subdivision into Valaisan and Piemont-Ligurian-derived units is, therefore, not necessary. Nevertheless, structurally deeper units in the Tauern Window (Glockner Nappe System) resemble the Valais units in the Central Alps, and structurally higher units (Matrei Zone) the Piemont-Ligurian units. In Figure 18.10, the units were correlated in this way.

*Glockner Nappe System in the Tauern Window*

The Glockner Nappe System (Fig. 18.10) overlies the Sub-Penninic Venediger Nappe System and comprises slices of ophiolites (mostly greenschists, prasinites to amphibolites, and some serpentinites) and a thick mass of calcareous schists, phyllites and mica-rich marbles (Bündnerschiefer) which were deposited on the ophiolites. Permo-Triassic sedimentary rocks directly beneath the Glockner Nappe System, partly with underlying continental basement lamellae, were previously interpreted as forming the pre-Jurassic part of the Glockner Nappe System (e.g. Tollmann 1977). However, it is now thought that they originated at the European continental margin (Kurz *et al.* 1998) and, thus, belong to the Sub-Penninic nappes (see above). In the Glockner Nappe System several nappes can be distinguished based on their facies and structural position. The lower nappes are rich in ophiolite fragments ('Glockner facies') whereas structurally higher nappes lack ophiolitic material ('Fusch facies').

The Glockner Nappe System underwent a subduction-related high-pressure metamorphism (reaching blueschist-facies conditions in some parts) and was overprinted by an Oligocene to Miocene greenschist- to amphibolite-facies metamorphism ('Tauernkristallisation') (Frank 1987; Schuster *et al.* 2004).

*Matrei Zone and equivalents in the Tauern Window*

In the Tauern Window the Glockner Nappe System is overlain by units representing a tectonic and sedimentary mélange with a characteristic lithological content. These comprise the Matrei Zone in the south and the 'Nordrahmenzone' (Northern frame zone) in the east, north and west. Slices of oceanic crust consisting of serpentinite, in contact with radiolarite and Aptychus limestone, are similar to the Piemont-Ligurian ophiolite nappes of the Central and Western Alps (Koller & Pestal 2003). Olistoliths of material from the Austro-Alpine continental margin are also found (Frisch *et al.* 1989). The matrix consists of Cretaceous to Jurassic calcareous schists or phyllites and mica-rich marbles (Bündnerschiefer). The Matrei Zone and its equivalents formed during the subduction of the Penninic Ocean below the Austro-Alpine continental margin in an accretionary wedge setting. The ophiolite components are derived from Jurassic oceanic crust in the southern part of that ocean. The Reckner Complex, overlying the adjoining Austro-Alpine units at the NW corner of the Tauern Window, also represents an ophiolite sequence with Jurassic radiolarites. Its present tectonic position is a result of complex tectonics in this region which are not yet fully understood.

The Reckner Complex and parts of the Matrei Zone underwent a blueschist-facies metamorphism with conditions of *c.* 1.0 GPa and 350°C, dated at *c.* 50 Ma (Dingeldey *et al.* 1997; Koller & Pestal 2003). Together with the 'Nordrahmenzone' they were overprinted under greenschist-facies conditions ('Tauern-Kristallisation').

*Penninic nappes of the Rechnitz Window Group*

At the eastern margin of the Eastern Alps, Penninic units occur in several tectonic windows (Fig. 18.1; e.g. Rechnitz, Eisenberg Window). They comprise slices of ophiolites and a pile of metasediments several kilometres thick, including calcareous mica schists, quartz phyllites, graphite phyllites, rare breccias and a few horizons of cargneule (Pahr 1980). Microfossils indicate a late Lower Cretaceous to Upper Cretaceous age for parts of the metasediments (Schönlaub 1973). During the Alpine metamorphic cycle, parts of the unit were metamorphosed under blueschist-facies conditions (330–370°C at a minimum pressure of 600–800 MPa), whereas the entire unit was affected by a greenschist-facies imprint with cooling ages ranging from 22 to 19 Ma (Koller 1985).

*Rhenodanubian Flysch Zone and Ybbsitz Klippen Zone*

The Rhenodanubian Flysch Zone (Figs 18.10, 18.11) along the northern border of the Eastern Alps is formed by predominantly deep-marine clastic sediments. Their age is Early Cretaceous to Maastrichtian (in the western part) or to Eocene (in the eastern part). In both parts, the Rhenodanbian Flysch Zone is subdivided into several nappes (Oberstdorf, Sigiswang and Üntschen nappes in the west, Laab, Greifenstein and Kahlenberg nappes in the east). These nappes were stacked by north- to NW-directed thrusts (Decker 1990). However, out-of-sequence thrusting led to the situation that the more northerly derived units (Oberstdorf Nappe in the west, Laab Nappe in the east) locally lie on top of more southerly derived units (Sigiswang Nappe in the west, Greifenstein and Kahlenberg nappes in the east; Fuchs 1985; Oberhauser 1995; Mattern 1999; Trautwein *et al.* 2001).

The Rhenodanubian Flysch was deposited at least partly on Jurassic oceanic crust which is preserved in the Ybbsitz Klippen Zone (Decker 1990; Schnabel 1992). This unit is structurally above the Rhenodanubian Flysch, directly at the base of the Austro-Alpine nappes. Similar to the Matrei Zone of the Tauern Window, the Ybbsitz Klippen Zone contains serpentinites in contact with Jurassic radiolarites and Aptychus limestones which correspond to the Piemont-Ligurian ophiolite nappes of the Central and Western Alps. The Ybbsitz Klippen Zone also includes Cretaceous deep-marine clastics which are developed in a similar facies to the Rhenodanubian Flysch nappes. Therefore, it probably represents a piece of the former basement of the Rhenodanubian Flysch nappes (Schnabel 1992).

Other parts of the basin floor may have been Cretaceous-age oceanic crust (see Fig. 18.3). The Rhenodanubian Flysch was detached from its oceanic basement during subduction and formed part of an accretionary wedge at the front of the Austro-Alpine nappes.

## *Austro-Alpine nappes*

The Austro-Alpine nappes represent a crustal fragment with a complex Phanerozoic history, documented by their (meta)sedimentary successions and the large variety of magmatic rocks. Due to imprints during the Variscan, Permo-Triassic, Eo-Alpine (Cretaceous) and Alpine (Tertiary) metamorphic events, large parts of the Austro-Alpine nappes consist of crystalline rocks. These rocks show different metamorphic histories, different relations to Permo-Mesozoic cover series, and occur in different tectonic positions within the nappe stack which formed during the Eo-Alpine and Alpine events. Other Austro-Alpine nappes are formed entirely from Permo-Mesozoic sedimentary rocks.

There has been a long-standing controversy about the tectonic correlation of the different nappes and their palaeogeographic restoration (Tollman 1959; Frank 1987; Neubauer *et al.* 2000). The subdivision provided below follows the one of Schmid *et al.* (2004*a*). The Austro-Alpine nappes are subdivided into Lower and Upper Austro-Alpine, and the Upper Austro-Alpine nappes themselves are grouped into several nappe systems (Figs 18.10, 18.11 & 18.12). This subdivision is an attempt to reflect not only the distribution of the sedimentary units of various ages and facies but also the internal tectonic style as well as the distribution and timing of metamorphism in the crystalline units.

### *Lower Austro-Alpine units*

The Lower Austro-Alpine units are defined as that part of the Austro-Alpine units which formed the northern margin of Apulia towards the Piemont-Ligurian Ocean in Jurassic to Early Tertiary times. For this reason they were affected by both the opening and closure of this oceanic realm. Preceding the opening in the Jurassic, the Lower Austro-Alpine units underwent extension, as demonstrated by the formation of tilted blocks, half-grabens and extensional detachment faults, as well as the deposition of breccias (Häusler 1987; Eberli 1988; Froitzheim & Eberli 1990; Manatschal & Nievergelt 1997). When the Penninic oceans closed in Late Cretaceous to Tertiary times, the Lower Austro-Alpine nappes were involved in the subduction-related deformation and underwent anchizonal to greenschist-facies metamorphism (Schuster *et al.* 2004).

Most of the Lower Austro-Alpine nappes contain a Variscan metamorphic basement and a Permo-Mesozoic cover series. Late Carboniferous sedimentary rocks occur between these in some nappes. The Permo-Mesozoic cover commences with Permian-age metaconglomerates and acidic metavolcanics, Lower Triassic quartzites (Semmering and Lantschfeld quarzites) and Middle Triassic shallow-marine carbonates (e.g. Wetterstein limestones and dolomites). A Keuper facies with gypsum and shales is characteristic of the Upper Triassic strata of the eastern part, whereas in the western part shales, sandstones and evaporites of Carnian age are followed by dolomite (Hauptdolomit) in the Norian. Fossil-rich marls (Kössen Formation) were deposited in the Rhaetian. Jurassic sediments include Liassic and early Middle Jurassic syntectonic breccias and Upper Jurassic radiolarites. The youngest sediments in the western Lower Austro-Alpine nappes are Late Cretaceous deep-marine clastics (Furrer 1985). The successions are tectonically truncated at different stratigraphic levels (Tollmann 1977).

The Lower Austro-Alpine units are widespread in eastern Switzerland, particularly along the western margin of the Austro-Alpine nappes. They comprise the structurally deeper Err Nappe System and the higher Bernina Nappe System (Froitzheim *et al.* 1994) in the southern part of that margin, as well as slivers of Lower Austro-Alpine nappes further north. As mentioned above, slivers of Austro-Alpine origin also occur within some Upper Penninic ocean-derived units, e.g. in the Engadine Window and the Matrei Zone and 'Nordrahmenzone' of the Tauern Window (Frisch *et al.* 1989). These may have been emplaced either as olistoliths, by tectonic mélange formation, or as extensional allochthons during rifting and continental breakup.

The Sadnig and Zaneberg complexes to the south of the Tauern Window, consisting of crystalline basement and Late Palaeozoic metasediments (Fuchs & Linner 2005), are interpreted to represent Lower Austro-Alpine units, as are the Radstadt Nappe System and the Katschberg Zone at the eastern margin of the Tauern Window, and the Reckner and Hippold nappes at the northwestern margin. In the easternmost part of the Alps, the Wechsel Nappe and the Semmering Nappe (with the exception of the Grobgneiss Nappe) can be correlated with the Lower Austro-Alpine units.

The Upper Austro-Alpine units comprise the remaining major part of the Austro-Alpine nappes. They represent a complex nappe stack formed by the Eo-Alpine tectonometamorphic events. During the Tertiary Alpine events, the nappe stack stayed in an upper-plate position and Tertiary deformation is mostly restricted to brittle faulting.

### *Bajuvaric, Tirolic, and Juvavic nappe systems (Northern Calcareous Alps)*

The Northern Calcareous Alps are composed of, from bottom to top, the Bajuvaric, Tirolic and Juvavic nappe systems (Figs 18.11 & 18.12). All three consist of Permian- to Palaeocene-age sediments. However, the Triassic facies evolution of the individual nappe systems shows significant differences and, furthermore, the tectonically controlled Jurassic and Cretaceous sediments are variable (Tollmann 1985). Additionally, these nappe systems were mobilized at different times during the Eo-Alpine orogeny. The nappe systems are described below in the context of the geodynamic evolution from Permian to Palaeocene times, closely following the more detailed descriptions in Mandl (2000) and Faupl & Wagreich (2000).

In Permian times, fluviatile red-beds (e.g. Prebichl Formation) and, locally, shallow-marine evaporitic sediments were deposited in the area of the three nappe systems. Subsequently, a shallow-marine basin developed, broadly silicilastic-filled with limestone layers intercalated in the upper part (Werfen schists, Alpiner Buntsandstein). In the Early Anisian, bituminous micritic carbonates were deposited under restricted shallow-water conditions (Gutenstein Formation). These pass laterally into Dasycladacean-bearing carbonates (Steinalm Formation). During the Middle Anisian rapid deepening and contemporaneous block faulting resulted in the formation of a seafloor relief and the deposition of nodular limestones (Reifling Formation). Most probably the tectonic activity was related to the rifting and breakup of the Meliata Ocean to the SE. Subsequently, extensive carbonate platforms (Wetterstein limestone and dolomite) developed in the proximal shelf area with lateral slope sediments (Raming limestone) deposited towards the intervening basins. These basins were filled by nodular limestones (Reifling Formation) and shales (Partnach Formation). On the deeper, outer shelf area closer to the Meliata Ocean, pelagic limestones were deposited (Hallstatt limestone). The transition from the pelagic limestones towards

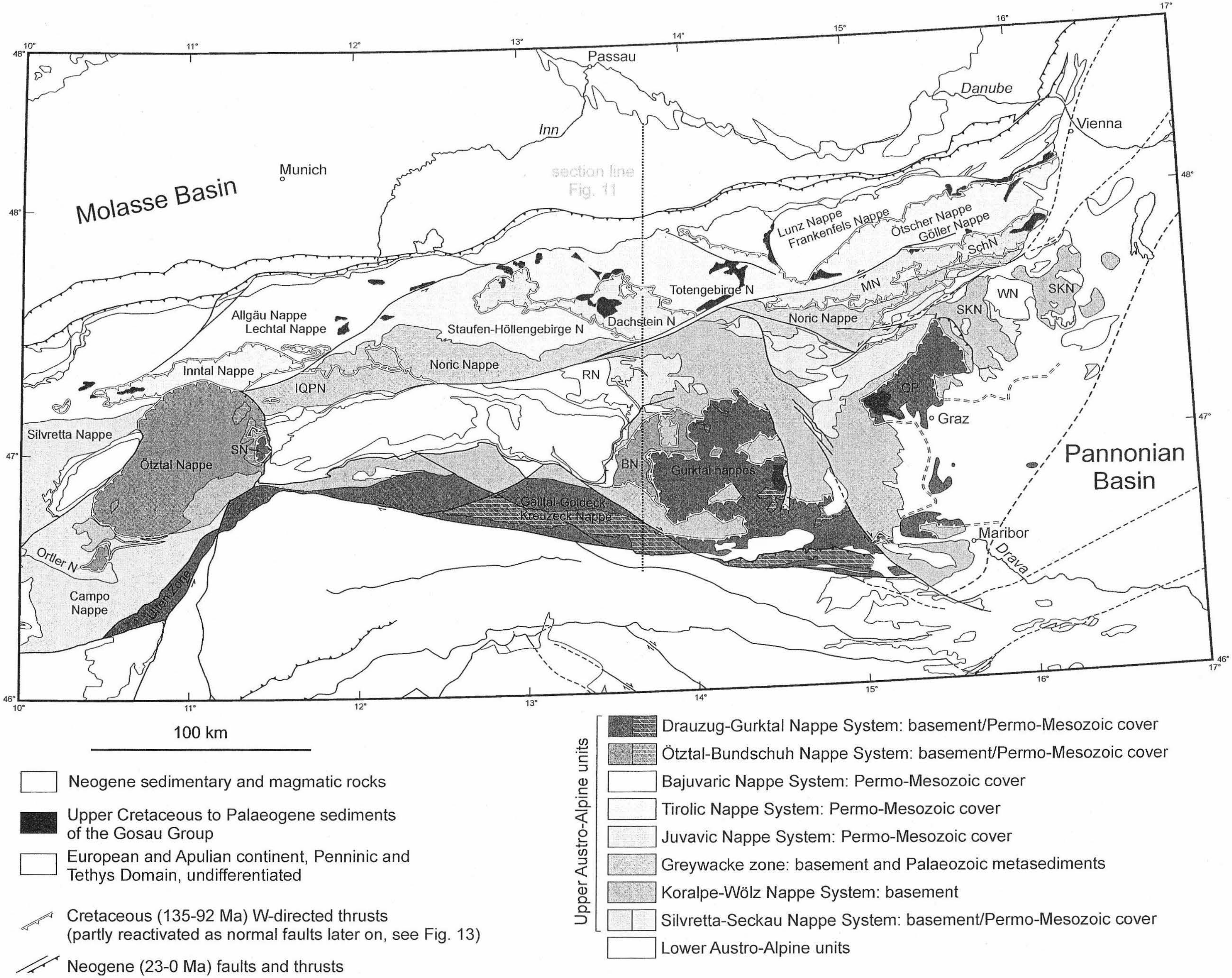



**Fig. 18.12.** Subdivision of the Austro-Alpine units into nappe systems. Abbreviations: see caption to Figure 18.10.

the radiolarites of the oceanic realm is not preserved in the Northern Calcareous Alps. In general, the Wetterstein carbonate platforms prograded onto the adjacent basinal sediments until earliest Carnian times. A subsequent sea-level lowstand resulted in a rapid decrease in carbonate production. The platforms became partly emergent and the remaining interplatform basins were completely filled by siliciclastic material from the European hinterland and intercalated carbonates (Raibl Group). Brackish-water sandstones with coals formed the Lunz Formation, whereas on the slopes and the deeper shelf the regression event can be recognized by the siliciclastic influx in the Reingraben schists.

In the Late Carnian, sea level began to rise and flooded the platform areas where both lagoonal carbonates (Waxeneck limestone and dolomite) and carbonates intercalated with gypsum (Opponitz limestone and dolomite) were now deposited. A transgressive pulse at the end of the Carnian resulted in the onlap of pelagic limestones onto the former platforms, and initial reef growth within the remaining shallow-marine areas. Subsequently a second extensive carbonate platform was established. In the central part of the Northern Calcareous Alps its reefs (Dachstein limestone) are situated on top of the Wetterstein reefs and are connected to the Hallstatt-facies basinal sediments (e.g. Hallstatt limestone, Pötschen limestone) of the deeper shelf by allodapic limestones (Gosausee limestone). In the eastern part of the Northern Calcareous Alps, in contrast, the reefs are located above the former platform interior, several kilometres behind the former Wetterstein reef front. Pelagic sediments were deposited on top of the latter until late Norian times (Aflenz Formation, Hallstatt limestone). A large lagoonal area extended behind (to the NW of) the Dachstein reefs. Bedded limestones (Dachstein limestone) were deposited directly behind the reefs, and dolomites (Hauptdolomit) more to the NW. In the most proximal (northwestern) part of the lagoonal area the dolomites contain evidence of siliciclastic influx derived from the siliciclastic shelf (Keuper facies) located further to the north. This influx increased in Rhaetian times and large parts of the lagoonal area were covered by dark marly sediments with small patch-reefs, indicating a restricted facies (Kössen Formation). Coevally, the marls of the Zlambach Formation were deposited in the deeper shelf areas. The total thickness of the Permo-Triassic succession in the Northern Calcareous Alps is up to *c.* 3 km.

At the beginning of the Jurassic, the Austro-Alpine shelf was drowned and synsedimentary faulting resulted in the formation of a complex seafloor topography. Strongly condensed nodular limestones with Fe/Mn hardgrounds and ammonoids (e.g. Adnet and Klaus formations) or crinoids (Hierlatz limestone) were deposited in elevated areas, whereas marly and cherty limestones (Allgäu Formation) were deposited in the intervening troughs. The water depths were greatest in the Callovian and Oxfordian, when radiolarites were deposited (Ruhpolding radiolarite). At this time, tectonically induced mass-transport complexes (i.e. breccias, olistoliths, and sliding blocks and nappes, several kilometres in size, in a radiolarite matrix) were deposited (e.g. Strubberg Formation). Initially, this material was derived from mobilized units of the Triassic deeper shelf (Hallstatt facies), and later from the Triassic platform margins which had been imbricated by a system of ramp faults (Gawlick & Suzuki 1999; Mandl 2000).

The material which became mobilized during the Jurassic tectonic event forms the Juvavic Nappe System. Ongoing sedimentation sealed the contacts along which the Juvavic nappes were emplaced on the units which later formed the Tirolic Nappe System. In the Kimmeridgian and Tithonian the new seafloor topography aided the growth of reefs (Plassen and Tressenstein limestones) on elevated parts of large sliding blocks and nappes of the Juvavic Nappe System, whereas pelagic limestones were deposited in deeper areas and to the north (Oberalm Formation). The Oberalm Formation extends up into the earliest Cretaceous, when deepening and increasing terrigeneous input led to a gradual transition into the Early Cretaceous marly Aptychus limestones of the Schrambach Formation.

In the Valanginian, shortening and nappe stacking propagated from SE to NW and resulted in the renewed deposition of mass-transport complexes. The nappes of the Tirolic Nappe System were detached from their basement. They represent southeastern parts of the Norian lagoonal area and carried the sealed Juvavic mass transport complexes on top. At their southern margins the Tirolic nappes remained in transgressive contact with their basement, which is formed by Palaeozoic-age metasediments of the Noric Nappe (belonging to the Greywacke Zone, see below). The Tirolic Nappe System was thrust onto the northwestern parts of the lagoonal area, which later became the Bajuvaric Nappe System.

During thrusting, large parts of the Tirolic and Juvavic nappe systems were elevated above sea level, but at the front of the individual Tirolic nappes the deep-marine clastics of the Rossfeld Formation were deposited. In the Aptian, the deposition of deformation-related mass-transport complexes shifted further to the NW. The deep-water marly Aptychus limestones passed up into a marl-rich succession with black shales (Tannheim Formation, Late Aptian to Albian) which was overlain by silty marls, turbidites and deep-marine conglomerates (Losenstein Formation, Albian to Cenomanian). The Tannheim-Losenstein Basin, located in the area of the Bajuvaric Nappe System, was filled with material from the south and from the north. According to Wagreich (2001*a*), local material of the arriving nappes was carried into the basin from the south whereas 'exotic' material was deposited from the north (e.g. metamorphic basement rocks, serpentinites, Jurassic and Cretaceous limestone of 'Urgonian' type).

In some areas of the Bajuvaric Nappe System, Cenomanian to Santonian marine sediments (including turbidites and shales) were deposited (Branderfleck Formation). They lie unconformably on older rocks. At this time detritus was still partly derived from the north and included phengites with Variscan cooling ages and glaucophane (von Eynatten & Gaupp 1999).

Large parts of the Northern Calcareous Alps became terrestrial in the Turonian. In the Late Turonian a new sedimentary cycle began with the deposition of the Gosau Group. The Gosau Group of the Northern Calcareous Alps can be subdivided into two subgroups (Faupl *et al.* 1987). The Lower Gosau Subgroup (Turonian to Campanian/Maastrichtian) consists of terrestrial, mainly fluviatile conglomerates which pass into a shallow-marine succession. Detritus from Permian quartz porphyries and from ophiolites is present, in addition to local detritus from the surrounding Mesozoic rocks. Sedimentation occurred in pull-apart-type basins in a predominantly transpressional regime (Wagreich & Faupl 1994).

By this time the entire Bajuvaric Nappe System had been detached from its basement. Typically the sedimentary profiles in the Bajuvaric nappes commence in the Middle or Upper Triassic. Permian-age strata are generally missing and may have remained attached to the basement.

The Upper Gosau Subgroup (Campanian–Palaeocene) is characterized by deep-water sediments such as marl-rich slope sediments with slumps (Nierental Formation) and breccias (Spitzenbach Formation, Zwieselalm Formation) which were deposited on a north-facing slope, partly below the carbonate compensation depth (CCD). The terrigenous material of the Upper Gosau Subgroup is characterized by metamorphic detritus eroded from an area to the south of the Northern Calcareous Alps. The sudden change in facies from the Lower to the Upper Gosau Subgroup took place diachronously, earlier in the NW of the Northern Calcareous Alps (Santonian) and progressively later in the SE (Palaeocene). In several localities the deep-water sediments unconformably rest upon deformed deposits of the Lower Gosau Subgroup (Faupl & Wagreich 2000).

During the Eo-Alpine Orogeny the Tirolic Nappe System underwent anchizonal to lowermost greenschist-facies metamorphism along its southern margin, whereas the rest of the Northern Calcareous Alps remained in the zone of diagenesis (Fig. 18.13; Kralik *et al.* 1987).

*Greywacke Zone*

The nappe system of the Greywacke Zone (Figs 18.11 18.12) underlies the Tirolic Nappe System and includes, from bottom to the top, the Veitsch, Silbersberg, Vöstenhof-Kaintaleck and Noric nappes (Neubauer *et al.* 1994*b*). The lowermost Veitsch Nappe mainly comprises Carboniferous clastic and carbonaceous metasediments with metamorphosed coal intercalations and magnesite. Permian clastic sediments are locally present. The overlying Silbersberg Nappe consists of phyllites with intercalations of chlorite schists and the Gloggnitz riebeckite gneiss. The latter represents alkaline metavolcanics, possibly Jurassic in age (Koller & Zemann 1990). The Vöstenhof-Kaintaleck Nappe is formed by a basement of paragneisses, mica schists and amphibolites with overlying post-Mid-Devonian clastic metasediments (e.g. Kalwang conglomerate). The crystalline rocks of this nappe show a Variscan amphibolite-facies imprint with remarkably old Variscan cooling ages of more than 350 Ma (Neubauer *et al.* 1994*b*).

The largest and uppermost Noric Nappe (including the western part of the Greywacke Zone) comprises a Lower Palaeozoic to Upper Carboniferous succession of phyllites, acidic and basic metavolcanics, quartzites and carbonates (marbles, dolomites, siderite and magnesite deposits). This is transgressively overlain by the Permian-age strata of the Tirolic Nappe System.

Within the nappe pile of the Greywacke Zone the Eo-Alpine metamorphic grade increases downward to upper greenschist facies in parts of the lowermost nappe. Only the uppermost Noric Nappe is in stratigraphic contact with the Tirolic Nappe System of the Northern Calcareous Alps; the other nappes are not.

*Silvretta-Seckau Nappe System*

This is the structurally lowermost Upper Austro-Alpine nappe system, directly overlying the Lower Austro-Alpine units. It includes in the west the Languard, Campo-Sesvenna and Silvretta nappes, to the south of the Tauern Window the Lasörling Complex, and to the east the Schladming, Seckau-Troiseck, Speik, Waldbach and Kulm complexes.

The Silvretta-Seckau Nappe System comprises biotite-plagioclase gneisses and mica schists, typical hornblende gneisses, layered amphibolites, and a wide spectrum of orthogneisses. In some units ultramafic complexes occur (Ulten peridotite, Speik Complex), as well as migmatites and eclogites (Ulten Zone, Silvretta Nappe, Speik Complex; Hauzenberger *et al.* 1996; Melcher *et al.* 2002).

According to Neubauer (2002), the magmatic rocks of the Silvretta-Seckau Nappe System reflect Precambrian to Ordovician collision, subduction and rifting processes. Orthogneisses from Variscan protoliths also occur. The eclogites formed during the Variscan orogenic cycle. Age determinations range from $397 \pm 8$ Ma in the Speik Complex (Faryad *et al.* 2002) to $351 \pm 22$ Ma in the Silvretta Nappe (Ladenhauf *et al.* 2001) and $336 \pm 4$ Ma in the Ulten Zone (Thöni 2006). The peak of Variscan metamorphism, which reached up to high-amphibolite-facies conditions and local anatexis, occurred at *c.* 330 Ma. Typical cooling ages are *c.* 310 Ma (Thöni 1999; Tumiati *et al.* 2003). Schlingen folds (regional-scale folds with steep axes) are a typical structural element of the Variscan deformation.

In the structurally lowermost parts of the Silvretta-Seckau Nappe System in the west (Languard, Campo-Sesvenna and Silvretta nappes), Permian pegmatites and gabbros are also present (Benciolini 1994). For these areas a Permo-Triassic thermal imprint up to amphibolite-facies conditions has been suggested (Schuster *et al.* 2001). The Eo-Alpine metamorphism reached uppermost-anchizonal to amphibolite-facies conditions (Fig. 18.13).

Most of the units show remnants of transgressive Permo-Mesozoic cover successions. The northwestern margin of the Silvretta Nappe is locally covered by Upper Carboniferous strata and transgressively overlain by Permo-Mesozoic sediments of the Lechtal Nappe, belonging to the Bajuvaric Nappe System (Amerom *et al.* 1982; Rockenschaub *et al.* 1983). Further to the south, transgressive Permo-Mesozoic sediments are preserved in the Landwasser and Ducan synclines on top of the Silvretta Nappe.

*Koralpe-Wölz Nappe System*

This comprises a series of basement nappes which lack Permo-Mesozoic cover. Within the nappe system several groups of units with special lithological compositions and a distinct metamorphic grade can be traced over long distances. The northernmost parts consist of lower-greenschist-facies units (e.g. Ennstal quartz phyllite). The southerly adjacent Wölz Complex and its equivalents are characterized by garnet mica schists with a dominant Eo-Alpine metamorphic imprint of upper-greenschist- to amphibolite-facies grade. Locally a Permo-Triassic greenschist-facies imprint has been documented (Schuster *et al.* 2001). The overlying Rappold Complex and its equivalents underwent Variscan, Permo-Triassic and Eo-Alpine amphibolite-facies metamorphism. Kyanite- and garnet-bearing two-mica gneisses with kyanite pseudomorphs after andalusite (so-called 'Disthenflasergneis') and Permian pegmatites are characteristic of this complex. Permian-age granites (e.g. Wolfsberg granite) occur in several places whereas pre-Permian orthogneisses have not been documented.

The southern part of the Koralpe-Wölz Nappe System is formed by units bearing Cretaceous-age eclogites. These units are also characterized by 'Disthenflasergneis' and Permian pegmatites. From west to east they include the Texel, Polinik-Prijakt, Millstatt, Saualpe-Koralpe-Pohorje and Sieggraben complexes. Most of the eclogites developed from amphibolites but some developed from Permian MORB-type gabbros and basalts (Miller & Thöni 1997). In the eclogite-bearing units some pre-Permian orthogneisses have been reported, although thus far not from the Saualpe-Koralpe-Pohorje Complex. An ultramafic body in the Pohorje area probably represents a fragment from the lithospheric mantle (Hinterlechner-Ravnik *et al.* 1991; Janák *et al.* 2006). The metasedimentary rocks of the Saualpe-Koralpe-Pohorje Complex record a Permo-Triassic amphibolite-facies imprint and an eo-Alpine eclogite-facies and subsequent amphibolite-facies overprint.

The Plankogel Complex overlies the eclogite-bearing units in the Saualpe-Koralpe area and is characterized by the presence of garnet-bearing mica schists which also show kyanite pseudo-

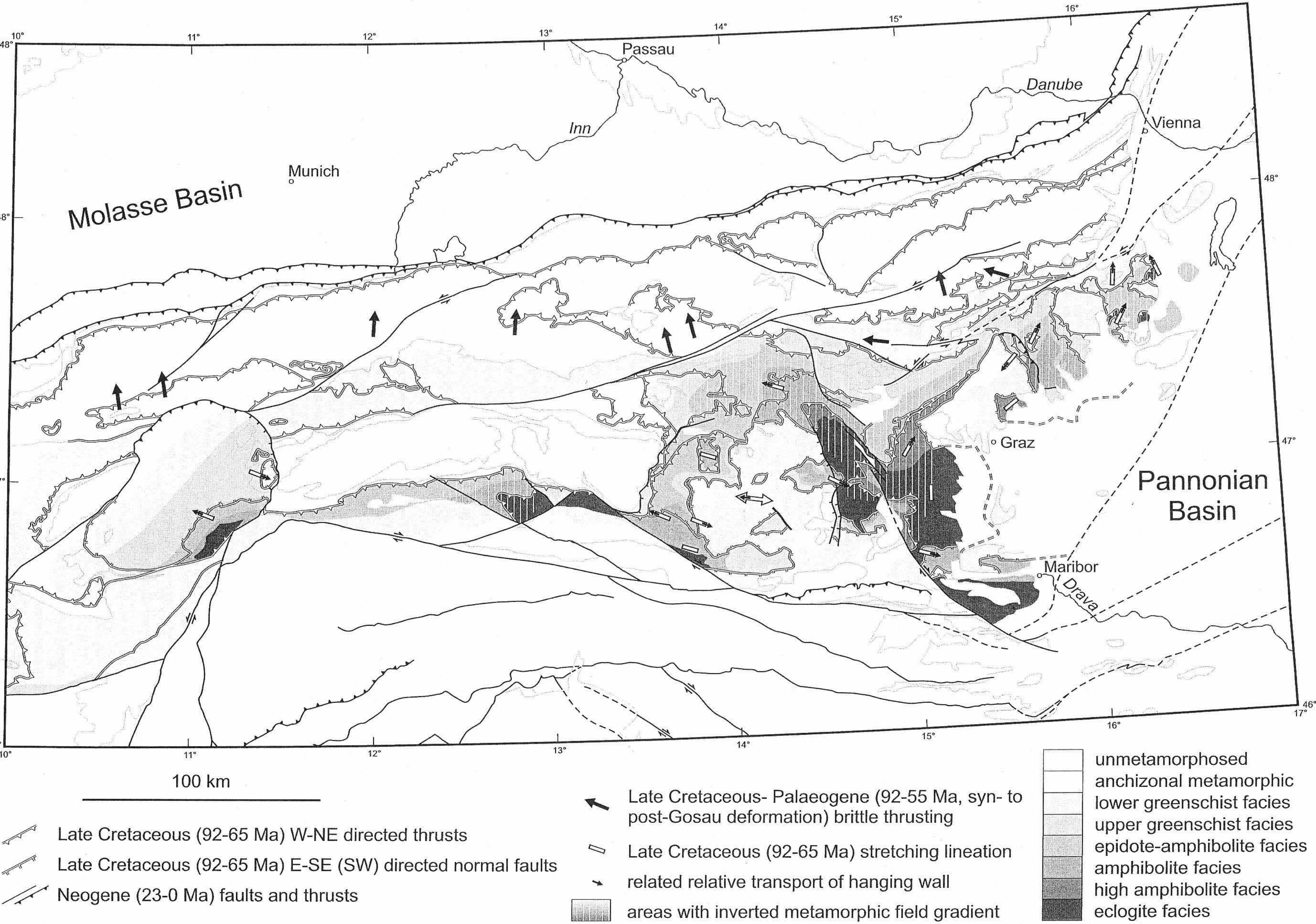



**Fig. 18.13.** Areal distribution of Cretaceous-age metamorphism in the Austro-Alpine units, age and character of major tectonic contacts, and directions of tectonic transport. Late Cretaceous top-to-the-SE low-angle normal faults have emplaced lower-grade on higher-grade metamorphic units.

morphs after andalusite. Intercalations of Mn-quartzites, serpentinites, amphibolites, marbles, and Permian pegmatites also occur. The Plankogel Complex is characterized by a Permo-Triassic imprint which reached up to amphibolite-facies conditions, and an Eo-Alpine amphibolite-facies overprint. The uppermost part of the Koralpe-Wölz Nappe System consists of several units rich in garnet-bearing mica schists. Some of these (e.g. Radenthein Complex) are very similar to the Wölz Complex. In all of them eo-Alpine upper-greenschist- to amphibolite-facies metamorphism is the predominant crystallization phase.

*Ötztal-Bundschuh Nappe System*

The Ötztal-Bundschuh Nappe System (Figs 18.11 & 18.12) overlies the Koralpe-Wölz Nappe System. It consists of the Ötztal Nappe including the 'Brenner Mesozoics' and the Bundschuh Nappe including the lower part of the 'Stangalm Mesozoics' (Stangnock Scholle; Tollmann 1977). Its lithological composition and pre-Alpine metamorphic history are quite similar to that of the Silvretta-Seckau Nappe System. The predominant lithologies include biotite-plagioclase gneisses, mica schists, amphibolites, and a wide range of orthogneisses. Migmatites and eclogites also occur. Remnants of transgressive Permo-Triassic cover sequences are present on top of the pre-Alpine basement and are characterized by successions quite similar to those of the western part of the Bajuvaric Nappe System, including the Norian Hauptdolomite.

Cambrian and Ordovician ages have been determined for some of the magmatic rocks as well as the migmatites in the Ötztal Nappe (Klötzli-Chowanetz *et al.* 1997; Thöni 1999). The eclogites developed during the Variscan tectonometamorphic cycle at *c.* 350 Ma (Miller & Thöni 1995) and the peak of Variscan amphibolite-facies metamorphism occurred at *c.* 340 Ma (Hoinkes *et al.* 1997). Schlingen folds are typical of the Variscan deformation. Cooling ages (Ar/Ar and Rb–Sr, muscovite and biotite) range between 315 and 300 Ma (Thöni 1999). According to Habler & Thöni (2005) the Matsch Unit in the southwestern part of the Ötztal Nappe belongs to the basal part of the nappe which was steepened up during the Eo-Alpine event. It shows an amphibolite-facies Permian metamorphic imprint and contains Permian pegmatites. The Eo-Alpine metamorphic imprint decreases upwards within the Ötztal Nappe from amphibolite- to greenschist-facies conditions (Schuster *et al.* 2004).

*Drauzug-Gurktal Nappe System*

This is structurally the highest of the nappe systems south of the Northern Calcareous Alps (Figs 18.11 & 18.12). It comprises units bordered by steeply dipping Alpine faults (tectonic blocks), located directly to the north of the Periadriatic Fault, as well as nappes overlying the Koralpe-Wölz and Ötztal-Bundschuh nappe systems.

The tectonic blocks consist of crystalline basement (e.g. Meran-Mauls basement, Defferegen, Strieden, Gaugen complexes), Palaeozoic metasediments (e.g. Thurntal quartz phyllite and Goldeck complexes), and Permo-Mesozoic sedimentary successions of the Drau Range (Lienz Dolomites, Gailtal Alps, Dobratsch, North Karawanken). The Permo-Mesozoic strata show a characteristic facies evolution similar to that of the western parts of the Southern Alps and the western parts of the Bajuvaric Nappe System (Bechstädt *et al.* 1976; Tollmann 1987; Lein *et al.* 1997), suggesting that all of these blocks were in a more westerly position during the Triassic than today. In the Jurassic and Early Cretaceous, they moved eastwards along a system of strike-slip faults, with respect to their former neighbouring units. The basement of these blocks consists of orthogneisses from pre-Variscan protoliths and metasedimentary rocks with a Variscan metamorphic imprint up to amphibolite-facies grade. A Permo-Triassic high-temperature/low-pressure overprint reached upper-amphibolite-facies conditions with local anatexis and formation of pegmatites in the lowermost parts (Schuster *et al.* 2001). This overprint becomes weaker towards higher levels of the basement. An Eo-Alpine overprint reached greenschist facies only at the base; the uppermost parts of the units stayed under conditions of diagenesis. Therefore, the pre-Alpine structures and metamorphic assemblages are well preserved.

The above-mentioned nappes comprise the Steinach Nappe, Gurktal Nappes, and the nappes of the Graz Palaeozoic. The Gurktal Nappes (including the Murau, Ackerl and Stolzalpen nappes) and the Steinach Nappe on the eastern and western sides, respectively, of the Tauern Window show many similarities indicating that they were parts of one continuous thrust sheet prior to the Tertiary exhumation of the Penninic rocks in the Tauern Window. The Murau, Stolzalpen and Steinach nappes are composed of Lower Palaeozoic rocks metamorphosed under greenschist-facies conditions. Additionally, the latter two include remnants of Carboniferous clastic sediments and coals of a Variscan intramontane basin (Krainer 1993). In contrast, the Ackerl Nappe comprises mica schists and paragneisses with a Variscan metamorphism (Neubauer 1980). Permo-Mesozoic sedimentary cover is present on top of the Gurktal Nappes. As in the case of the Mesozoic cover of the tectonic blocks described above, the facies of these cover rocks indicates a westerly position during the Triassic and later eastward translation relative to the Bajuvaric Nappe System and the Southern Alps. The nappes of the Graz Palaeozoic (Schöckel, Hochlantsch and Rannach nappes) are formed by Lower Palaeozoic to lowermost Upper Carboniferous sequences with various patterns of facies evolution (Flügel & Hubmann 2000). Upward-decreasing Eo-Alpine metamorphic conditions, from greenschist facies at the base to diagenesis at the top, affected the Gurktal and Graz Palaeozoic nappes (Hoinkes *et al.* 1999). Sediments of the Gosau Group (Late Cretacous) rest unconformably on top of the Gurktal Nappes and the Graz Palaeozoic.

## *South Alpine units*

The South Alpine units comprise the part of the Alps located to the south of the Periadriatic Fault. They are characterized by the absence of (or a weak) Alpine metamorphic overprint. The South Alpine units may be subdivided, from a tectonic point of view, into the Lombardic-Giudicaric fold-and-thrust belt to the west, and the eastern South Alpine units (Dolomites, Carnio and Julian Alps) to the east.

The Lombardic-Giudicaric fold-and-thrust belt (Milano Belt; Fig. 18.5) is a south-vergent thrust belt affecting basement and sedimentary cover rocks. It evolved from Late Oligocene to Late Miocene times. It affected an area which had previously been stretched in an east–west direction during Liassic rifting preceding the opening of the Piemont-Ligurian Ocean. The resultant north–south trending normal faults were reactivated in Tertiary times as transverse zones with a transpressional character. Thrusts pre-dating the Adamello intrusion (i.e. older than 43 Ma) are observed in the northern part of the thrust belt. These are also directed towards the south. The various parts of the fold-and-thrust belt are briefly described below.

*Canavese Zone*

This is a narrow belt of crystalline basement and Permo-Mesozoic sedimentary rocks directly to the east of the westernmost part of the Periadriatic Fault (the Canavese Fault; Fig. 18.8). The Mesozoic sedimentary succession is very similar to that in the most distal parts of the Lower Austro-Alpine Jurassic-age passive continental margin, in particular the Err Nappe in Graubünden, and is characterized by the presence of rift-related polymict breccias in the Early and Middle Jurassic (Ferrando *et al.* 2004). In the Jurassic, the Canavese Zone and the Err Nappe were adjacent areas on the southeastern passive margin of the Piemont-Ligurian Ocean. The Canavese Zone is separated from the Ivrea Zone to the east by an Alpine fault contact (Internal Canavese Line; Biino & Compagnoni 1989).

*Ivrea Zone*

The Ivrea Zone (or Ivrea-Verbano Zone; Fig. 18.5) represents Variscan basement, comprising mainly paragneiss, metabasic rocks and some marbles, with a Permian-age high-temperature metamorphic overprint of granulite facies in the northwestern part, to amphibolite facies in the SE. The Permian-age (*c.* 285 Ma; Boriani & Villa 1997) metamorphism was associated with voluminous gabbro intrusions (magmatic underplating; Voshage *et al.* 1990).

The Ivrea Zone forms an upright, strongly asymmetric antiform (Proman antiform), with the hinge very close to the northwestern boundary. Peridotite bodies are found in the core of this antiform (Schmid *et al.* 1987) at Finero, Balmuccia and Baldissero. These represent former subcontinental mantle. In most of the Ivrea Zone the layering is steep, forming the southeastern limb of the antiform. Schmid *et al.* (1987) interpreted the Proman antiform as an Alpine structure deforming the entire (but already strongly thinned) crust, so that the top of the mantle is exposed in the core of the antiform, and is flanked by the steepened former lower crust to the SE. The crust, according to these authors, had been thinned by Early Jurassic rift extension. Other authors, however, assume that much of the crustal thinning and exhumation of the Ivrea Zone is Permian in age and attribute less importance to Jurassic and Alpine deformation (e.g. Brodie *et al.* 1989; Boriani & Villa 1997).

The Ivrea Zone is underlain by the Ivrea geophysical body, a shallow-seated mass of mantle material resulting in a strong positive gravity anomaly. The emplacement of this body at shallow depth is explained by a combination of extensional exhumation in the Permian and/or Mesozoic, Alpine forethrusting, and Late Alpine backthrusting (Schmid *et al.* 1987).

The contact between the Ivrea Zone and the southeasterly adjacent Strona-Ceneri Zone is formed by two different tectonic features: the Pogallo Shear Zone to the NE, and the Cossato-Mergozzo-Brissago Shear Zone to the SW. Both of these shear zones are steeply orientated. The Cossato-Mergozzo-Brissago shear zone was interpreted by Handy *et al.* (1999) as an Early Permian sinistrally transtensional shear zone (285 to 275 Ma; Mulch *et al.* 2004). According to Hodges & Fountain (1984) and Schmid *et al.* (1987), the Pogallo Shear Zone, which is younger and offsets the Cossato-Mergozzo-Brissago Shear Zone, was originally active as an east-dipping, extensional, rift-related shear zone during the Early Jurassic, exhuming the Ivrea Zone relative to the Strona-Cenri Zone. It was later steepened by Alpine shortening. In contrast to this, Boriani & Villa (1997), using $^{39}Ar/^{40}Ar$ dating (mostly on amphibole), suggested that there was no significant differential exhumation between the Ivrea and Strona-Ceneri zones after 270 Ma.

*Strona-Ceneri Zone*

The Strona-Ceneri Zone, the Val Colla Zone, and the Orobic basement (Fig. 18.5) represent upper crustal basement units of the Lombardic-Giudicaric fold-and-thrust belt. The Strona-Ceneri Zone was formed by amphibolite-facies metasediments and amphibolites intruded by Ordovician granitoids. The peak of the Variscan amphibolite-facies metamorphism occurred at *c.* 340 Ma (Boriani & Villa 1997) or 320 Ma (Handy *et al.* 1999). Schlingen folds (regional-scale folds with steep axes) formed at *c.* 320 Ma (Zurbriggen *et al.* 1998) or at *c.* 290 Ma (Boriani & Villa 1997). There is also evidence for earlier (Ordovician) metamorphism. In the Early Permian, the basement was intruded by granitoids (Baveno granite) approximately coeval with, and genetically related to, granulite-facies metamorphism and gabbroic underplating in the Ivrea Zone.

*Val Colla Zone*

The Val Colla Zone is separated from the Strona-Ceneri Zone by a mylonite belt, the Val Colla Shear Zone (Fig. 18.5), which is of Carboniferous age according to Handy *et al.* (1999). The Val Colla Zone comprises mica schists, phyllites, and characteristic leucocratic granitoid gneisses with mylonitic and cataclastic deformation features (Gneiss Chiari). Towards the SE, the Val Colla Zone is separated from the sedimentary cover rocks of the Generoso Basin by the Monte Grona Fault, a steeply dipping, east–west striking fault and shear zone which forms the northeastern extension of the Early Jurassic, east-dipping, rift-related Lugano Fault (Bertotti 1990; Fig. 18.5). This portion of the fault was originally shallowly inclined but was steepened and exhumed during Tertiary compression. To the east, the fault may be traced into the Orobic Basement as the Valgrande Mylonite Zone.

*Orobic Basement*

East of Lake Como, the basement of the Southern Alps is exposed as a west–east striking belt, the Orobic Basement, between the Periadriatic (Tonale) Fault to the north and the Permo-Mesozoic cover to the south. In addition, basement rocks are exposed in the cores of Alpine ramp anticlines further south, including, from west to east, the Orobic (Fig. 18.6), Trabuchello-Cabianca, Cedegolo and Camuna anticlines. Most of the Orobic basement underwent Variscan amphibolite-facies metamorphism but the easternmost part reached only greenschist facies and contains Ordovician–Silurian palynomorphs (Colombo & Tunesi 1999). The basement was locally intruded by Early Permian granitoids, e.g. the Val Biandino Pluton in the western part of the Orobic Anticline (285 ± 20 Ma; Thöni *et al.* 1992). In addition to polyphase Variscan deformation, the Orobic basement also underwent Permian normal faulting related to the formation of the Collio Basin (Sciunnach 2003; Frotzheim *et al.* 2008). Furthermore, the basement was affected by Mesozoic extensional shearing, e.g. along the Valgrande line, the intrabasement continuation of the Lugano-Monte Grona normal fault (see above) east of Lago di Como.

*Post-Variscan cover of the Lombardic-Giudicaric belt*

Deposition of the post-Variscan succession of the western Southern Alps locally commenced with Upper Carboniferous clastic sediments unconformably overlying the basement. These were followed by the mixed volcanic/sedimentary Collio Formation of Early Permian age (Sakmarian to Artinskian), with a thickness of up to 1250 m in the area of the Orobic Anticline (Sciunnach 2001). These rocks were deposited in rift grabens, the most important being the Collio Basin between Lake Como and the Eocene- to Oligocene-age Adamello intrusion. Deposition was contemporaneous with granitoid intrusion in the upper crustal basement and gabbroic underplating in the lower crust (Ivrea

Zone). The Collio Formation is overlain with an angular unconformity by Late Permian fluvial clastics of the Verrucano Lombardo, followed by a Triassic shelf succession similar to the one in the Austro-Alpine nappes. The westernmost portions are similar to the successions in the Lower Austro-Alpine nappes (relatively thin and marginal-marine), while the more easterly successions are thicker and more open-marine.

Rifting related to the formation of the Piemont-Ligurian Ocean began in the latest Triassic and was associated with the development of significant facies and thickness variations within the Rhaetian sediments. Rifting was most pronounced during the Early Jurassic. Prominent east-dipping normal faults active at this stage include the Lago Maggiore Fault at the western border of the Monte Nudo Basin and the Lugano Fault at the western border of the Generoso Basin (Bernoulli 1964). The basins were filled with up to 3 km of hemipelagic limestones and marls of Liassic age (Moltrasio limestone) whereas the coeval sediments on the graben shoulders are extremely thin. The former northern extension of the Monte Nudo Basin is found in the Ela Nappe (Lower Austro-Alpine nappes, Graubünden), while the Generoso Basin can be extended into the Ortler Nappe (Upper Austro-Alpine nappes, Graubünden) (Bernoulli *et al.* 1990), indicating that these areas were directly adjacent and were not separated by an ocean basin (Froitzheim *et al.* 1996).

The Late Jurassic–Early Cretaceous succession is mostly pelagic. In Late Cenomanian to Campanian times, siliciclastic deep-marine sediments were derived from source areas to the north (later Austro-Alpine nappes). Pelagic sedimentation was re-established in the Maastrichtian and continued until the Middle Eocene, interfingering with bioclastic turbidites (Schumacher *et al.* 1997). These sediments were unconformably overlain by the Late Oligocene to Middle Miocene deep-water clastics of the Gonfolite Lombarda Group. These represent the onset of the infilling of the South Alpine foreland basin.

The eastern part of the Lombardic-Giudicaric fold-and-thrust belt was intruded by the Adamello granitoids, of Eocene to Oligocene age. Intrusions are oldest in the southern part (43 Ma; Del Moro *et al.* 1983) and youngest in the north (34 to 32 Ma; Stipp *et al.* 2004).

*Dolomites*

The area of the Dolomites lies to the east of the Lombardic-Giudicaric fold-and-thrust belt and is characterized by Middle to Late Triassic carbonate edifices forming spectacular mountains. The basement is formed by Cambrian- to Silurian-age sedimentary and volcanic rocks deformed and metamorphosed to greenschist facies during the Variscan Orogeny. This is unconformably overlain by thin Lower Permian conglomerates and Lower Permian, predominantly rhyolitic volcanic rocks (Bozen quartz porphyry). The latter are up to 2000 m thick in the caldera structure near Bozen. Plutonic equivalents of this complex are found within the basement (Brixen granitoids close to the Periadriatic Fault, Cima d'Asta granitoids further south). The cover succession continues with red-beds (Gardena/Gröden sandstone, Upper Permian), evaporites and carbonates (Bellerophon Formation, Upper Permian) and shallow-water carbonates and terrigenous deposits (Werfen Formation, Lower Triassic). Early Middle Triassic (Pelsonian) block faulting and local erosion, followed by strong subsidence, were contemporaneous with the main rifting phase of the Meliata Ocean (Kozur 1991). This rifting event initiated the formation of carbonate platforms with pronounced lateral facies heterogeneities during the Middle Triassic (e.g. Ladinian Sciliar/Schlern dolomite). In the Ladinian, important volcanism is represented by shoshonitic basalt dykes cutting the older rocks and by pillow lavas and hyaloclastites filling the basins between the reef platforms. The Predazzo and Monzoni diorite, monzonite, monzodiorite and monzogabbro bodies intruded Permian and Triassic carbonates and represent plutonic equivalents of the volcanics. In the Late Ladinian, following the volcanic event, carbonate platform growth resumed. In the late Carnian, terrigeneous, evaporite and carbonate sediments of the Raibl Group were deposited. Subsequently, the area became part of the extensive Dolomia Principale/Hauptdolomit platform of Norian age. In the Rhaetian, Jurassic and Cretaceous, the Dolomites formed part of the relatively stable Venetian platform. Limestones were deposited during the Rhaetian and Jurassic, and marls during the Cretaceous.

The Lower Miocene, marine Monte Parei conglomerate overlies folded and eroded Jurassic rocks. These had been deformed in the course of SW-directed, Palaeogene thrusting. The conglomerates were themselves deformed by the second phase of Alpine deformation in the Dolomites, is Neogene NNW–SSE shortening leading to the formation of bivergent thrusts (Doglioni 1987).

*Carnic Alps and South Karawanken Mountains*

The belt of Variscan basement at the northern border of the Southern Alps continues eastward into the Carnic Alps. Further east, it is hidden beneath Permo-Mesozoic sediments, to be exposed again in the South Karawanken. Continuing the trend of the eastward-decreasing grade of Variscan metamorphism in the Southern Alps, the Variscan part of the Carnic Alps and the South Karawanken are formed by unmetamorphosed to low-grade metamorphic successions of Late Ordovician to Early Carboniferous sediments, affected by Carboniferous, south-directed nappe stacking. The Variscan unconformity lies in the Westphalian C to D and is overlain by continental to shallow-marine clastic sediments of Late Carboniferous age (Auernig Group). Marine shelf sedimentation began in the Early Permian (e.g. Trogkofel limestone) and was interrupted by a phase of uplift and erosion, in the Early Permian, after which a new transgressive series commenced with the Gröden Formation (red-beds) and the Bellerophon Formation (dolomites and evaporites) in the Late Permian, followed by a similar Mesozoic succession as described above for the Dolomites (Schönlaub & Histon 2000*b*; see also McCann *et al.* 2008*b*). In the Carnic Alps, metamorphism increases in grade from anchizonal to epizonal, in a direction from east to west but also from south to north towards the Periadriatic Fault. Variscan and Alpine metamorphism were approximately of the same grade (Rantitsch 1997). The fact that levels with epizonal Alpine metamorphism reach the surface can be explained by the action of Tertiary-age dextral transpression along the Periadriatic Fault, leading to the development of a pop-up structure and erosional exhumation. The metamorphic zonation is offset by NW-trending dextral and NE-trending sinistral strike-slip faults interpreted as synthetic and antithetic Riedel shears, related to the motion along the Periadriatic Fault (Rantitsch 1997).

## *Remnants of the Meliata Ocean*

In the Alps, remnants of basement and cover of the Meliata Ocean are present only within some small tectonic slivers. Similar units are more extensively exposed in the Western Carpathians (Meliaticum, see below). The Meliata Ocean existed from the Middle Triassic to Jurassic or possibly into Cretaceous times. It was bordered by the Western Carpathian and Austro-Alpine domains to the north and NW, by the South Alpine Domain to the west, and the Dinaride domain to the SW. The exact geometry of this ocean and its relations to other sub-basins of Tethys are the subject of discussion (Channell & Kozur 1997; Stampfli & Borel 2004).

Remnants of the Meliata Ocean occur in the eastern part of the Northern Calcareous Alps, in a strongly sheared zone below the base of the Schneeberg Nappe (Juvavic Nappe System), partly overlying the Permian cover of the Noric Nappe (Greywacke Zone), and partly the Göller Nappe (Tirolic Nappe System) (Kozur & Mostler 1992; Mandl & Ondrejickova 1993; Fig. 18.10). Lithologically, these units include serpentinites, basic volcanic rocks, and Jurassic metasediments. The latter comprise upper Callovian mass-flow deposits with components of Triassic radiolarites and limestones, radiolarian-bearing cherty shales, and dark shales and sandstones at the top. These rocks can be correlated with the Meliaticum in the Western Carpathians. They show an upper anchizonal metamorphic imprint but no indications of subduction-related high-pressure/low-temperature metamorphism.

Basic volcanic rocks and serpentinites also occur in mélanges with a matrix of Permian-age evaporites at the base of the Juvavic Nappe System. They were interpreted as dismembered Triassic ophiolites from the Meliata Ocean tectonically introduced into the mélanges (Kozur & Mostler 1992). Detritus of serpentinite and chrome spinel with a presumed origin from the Meliata Ocean occurs in many Cretaceous synorogenic sediments from the Berriasian onwards (Faupl & Wagreich 2000). Additionally, pebbles of Triassic radiolarite, together with basic volcanic rocks, greenschists and partly garnet-bearing amphibolites occur in some Late Cretaceous sediments of the Lower Gosau Subgroup (Gruber *et al.* 1992). In the garnet-bearing amphibolites, $^{40}Ar/^{39}Ar$ hornblende dating yielded Early Jurassic ages (Schuster *et al.* 2003). This material may have been eroded from the metamorphic sole of obducted ophiolites from the Meliata-Vardar oceanic realm.

The tectonic evolution of the Alps is described in chronological order in the following sections.

## Permian and Triassic tectonics in the Alps

Graben structures and sedimentation of post-Variscan continental sediments in the area of the future Alps indicate thinning of the thickened Variscan crust by orogenic collapse and erosion during Late Carboniferous to Early Permian time (Wopfner 1984; Ziegler 1993; Bonin *et al.* 1993). Extension is also recorded by magmatic rocks and by high-temperature/low-pressure metamorphism. These features are particularly widespread in the South Alpine and Austro-Alpine units, but similar processes also affected the Penninic and Helvetic units. In the Mid-Permian, marine transgressions in parts of the Austro-Alpine and South Alpine units were accompanied by the deposition of fine-grained clastic, evaporitic and carbonate sediments, whereas fluvial sediments were deposited in continental areas. The distribution and facies of the sediments argue for a relatively flat topography, a low altitude and a normal crustal thickness.

In the Early Permian, lithospheric extension resulted in the formation of basaltic melts in the mantle. The melts mainly underplated the crust while some rose into the crust, crystallizing as gabbros with ages of *c.* 290 Ma (Quick *et al.* 1992; Thöni 1999). The resultant heating was responsible for high-temperature/low-pressure metamorphism and melting in the lower crust.

In the Ivrea Zone (Southern Alps), where the Mesozoic mantle–crust boundary was exhumed during the Alpine event, gabbros are interlayered with kinzigites which are restites from the molten lower crust. Due to high melting rates, the melts generated from the lower crust have calc-alkaline signatures. They reached the surface at *c.* 275–285 Ma and are present mostly as rhyolitic volcanic rocks, e.g. the Bozen quartz porphyry in the area of the Dolomites (Klötzli *et al.* 2003), volcanites of the Collio Basin in the western Southern Alps (Schaltegger & Brack 1999), rhyolites within the Permian succession of the Helvetic nappes (Glarus Nappe) and Penninic nappes (e.g. St. Bernard Nappe; Sartori *et al.* 2006), and as layers and pebbles in the Lower Permian clastic sediments of the Austro-Alpine nappes.

Thinning due to extension and additional advective heat from the mantle and lower crustal melts led to a high-temperature/low-pressure metamorphic event in the middle and upper crust and the development of anatectic granitic melts and pegmatoids. Based on andalusite- and sillimanite-bearing assemblages, the geothermal gradient in the middle crust was typically >40°C/km (Diella *et al.* 1992; Habler & Thöni 2001). Pegmatoides which are partly spodumen-bearing are frequent in granulite- and upper amphibolite facies metamorphic rocks, whereas they have not been found in the overlying lower amphibolite and greenschist facies rock series (Schuster *et al.* 2001). Sm–Nd dating of garnets from the pegmatoides as well as from metapelitic rocks indicates that the thermal peak occurred at 270 to 245 Ma (Thöni & Miller 2000; Schuster *et al.* 2001). These ages correspond to those of the Permian gabbroic and granitic intrusions in the Koralpe-Wölz Nappe System (Miller & Thöni 1997). The Permian sediments show evidence of a diastathermal metamorphic imprint due to the high heat flow near the surface (Ferreiro Mählmann 1995).

Following thermal subsidence in the Early Triassic, a further rifting event took place in the Early Middle Triassic, leading to the opening of the Meliata Ocean. Crustal extension recommenced in the Late Triassic and continued into the Jurassic (Bertotti *et al.* 1993). Through these processes, the metamorphic rocks were covered by Triassic sedimentary sequences up to 3 km thick. This resulted in slow cooling from the high geothermal gradient to a gradient of about 25°C/km at *c.* 200 Ma. This thermal history can be studied in the Gailtal-Goldeck-Kreuzeck Nappe (Drauzug-Gurktal Nappe System) and the Silvretta Nappe (Silvretta-Seckau Nappe System). There the $^{40}Ar/^{39}Ar$ muscovite ages decrease continuously in a downward direction from typical Variscan cooling ages of c. 310 Ma close to the top of the Variscan basement, to about 210 Ma in the andalusite-bearing, and 190 Ma in the sillimanite-bearing rocks (Schuster *et al.* 2001).

The slow cooling and the episodic extension of the lithosphere during the Triassic led to subsidence and the accumulation of shallow-marine sediments. In the Lower Triassic almost the entire Alpine area was transgressed. The change from restricted to open-marine conditions in the Ladinian and the accompanying facies differentiation into lagoonal, reef (Wetterstein dolomite and limestone) and deep-water sediments indicates the existence of a nearby ocean at that time. This Triassic-age ocean was the Meliata Ocean which may have been a backarc ocean related to subduction of the Palaeotethys Ocean (Stampfli & Borel 2004). Following a phase of regression in the Carnian, due to a global sea-level fall (Brandner 1984), a second stage of carbonate platform growth produced the classic facies succession with the proximal Hauptdolomit, Dachsteinkalk reef, and deep-water Hallstatt facies developed along the passive continental margin (Tollmann 1977; Mandl 2000).

## Jurassic tectonics

### *Rifting and break-up of the Piemont-Liguria Ocean*

A distinct change in tectonic regime and sedimentation can be recognized at the beginning of the Jurassic. The extensive shallow-water shelf area NW of the Meliata Ocean was affected by rifting. The Triassic carbonate platforms were drowned and the input of tectonically controlled mass-flow deposits became increasingly important. Rifting eventually led to the opening of the Piemont-Ligurian Ocean in the Middle Jurassic, cutting obliquely across the Triassic-age shelf. Also during Middle Jurassic times, compressional tectonics affected the southeasternmost part of the Austro-Alpine units.

Fault systems related to the breakup of the Piemont-Ligurian Ocean are well-preserved in the Austro-Alpine nappes in Graubünden (Froitzheim & Manatschal 1996), in the western Southern Alps (Bernoulli *et al.* 1990), and in the Briançonnais and Dauphinois zones of the Western Alps (Lemoine *et al.* 1986). Rifting was most intense in two phases, early Liassic and late Liassic–Middle Jurassic. During the first phase, the upper crust was fragmented by normal faults which dipped predominantly to the east, but the overall geometry of lithospheric stretching was probably close to symmetric. During the second phase, a west-dipping normal fault cut through the already strongly thinned crust and into the remaining mantle lithosphere, and developed into a sequence of top-to-the-west low-angle detachment faults that tectonically denuded the upper mantle in the Austro-Alpine–Penninic transition zone (Froitzheim & Manatschal 1996; Manatschal & Nievergelt 1997; Manatschal 2004). On the Austro-Alpine–South Alpine side of the Piemont-Ligurian Ocean, rifting produced a basement high with thin sediment cover in the distal part of the passive continental margin. This high can be followed from the western Southern Alps, where it was located immediately east of the Canavese zone (Ferrando *et al.* 2004), to the Lower Austro-Alpine Bernina Nappe in Graubünden. A similar position is occupied by the Tatric/Inovec basement nappe in the Western Carpathians (Plašienka 1995*a*).

In asymmteric rifting models (e.g. Lemoine *et al.* 1987; Froitzheim & Manatschal 1996; Manatschal 2004), the Austro-Alpine and South Alpine units were interpreted as representing the lower-plate margin, and the Briançonnais the upper-plate margin with respect to a detachment fault cutting through the entire lithosphere. This hypothesis is supported by the observed top-to-the-west detachment faults, and it explains the Middle Jurassic emergence of the Briançonnais units in terms of an isostatic response to non-uniform stretching, where the thickness of the mantle lithosphere was markedly reduced beneath the Briançonnais. The asymmetric hypothesis was recently challenged by a numerical modelling study (Nagel & Buck 2004) where two conjugate continental margins with the characteristics of the Austro-Alpine–Penninic boundary (mantle exhumation, top-to-the ocean detachment fault, blocks tilted towards the continent) were produced by a symmetric rifting process. If this model were applied to the opening of the Piemont-Ligurian Ocean, a different explanation for the Briançonnais emersion would be required.

During the process of continental breakup, the Cervinia Terrane was isolated from the Austro-Alpine–South Alpine continental margin. According to Froitzheim & Manatschal (1996), this terrane represented an extensional allochthon close to the Austro-Alpine–South Alpine margin, that is, a fault block of continental crust that was left behind during extensional unroofing of the mantle. Alternatively, the Cervinia Terrane may represent a microcontinent between two branches of the Piemont-Ligurian Ocean (Fig. 18.3).

In contrast to the approximately orthogonal, east–west to SE–NW directed stretching in the Central and Western Alps, Jurassic rifting occurred in a framework of sinistral transtension further east in the Eastern Alps. In this regime, half-grabens along the northern margin of the Austro-Alpine units (Lower Austro-Alpine domain) were filled with marine, breccia-rich successions (Häusler 1987).

*Compressional tectonics in the internal Austro-Alpine units*

Beginning in Mid-Jurassic times, the southeastern part of the Austro-Alpine units, close to the Meliata Ocean, was affected by compressional deformation. In general, this deformation commenced in the SE and propagated northwestward, as documented by the presence of mass-transport complexes in the upper Jurassic deep-water sediments of the Juvavic and Tirolic nappe systems (Gawlick *et al.* 1999; Mandl 2000). In the initial stage, ophiolites of the Meliata-Vardar oceanic realm were obducted onto the Austro-Alpine continental margin, probably resulting from the collision of this margin with an east- or south-dipping, intra-oceanic subduction zone. In contrast to the Dinarides where the obducted ophiolite sheets are exposed over large areas (Pamić 2002), the obduction is poorly documented in the Eastern Alps. It is, however, suggested by the presence of redeposited detritus of ophiolitic material and Triassic radiolarites within syntectonic sediments (Faupl & Wagreich 2000). Any obducted ophiolite sheets which may have existed on top of the Austro-Alpine nappes have since been completely eroded.

The oldest sediments containing redeposited Triassic radiolarites are Callovian mass-transport complexes in the Meliata-derived slivers of the Northern Calcareous Alps (Mandl & Ondrejickova 1993). Callovian to Early/Middle Oxfordian mass-transport complexes in the Strubberg Formation of the Tirolic Nappe System (Gawlik & Suzuki 1999) record the initial detachment of the most distal Triassic shelf sediments from their basement and their stacking as thrust sheets. These complexes contain up to kilometre-sized olistoliths of Triassic sediments developed in the pelagic Hallstatt facies. In the Oxfordian, northwestward-propagating thrusting affected the Triassic-age carbonate platforms, creating the large Juvavic nappes (Dachstein Nappe and Mürzalpen Nappe). The Dachstein Nappe was thrust over Jurassic mass-transport complexes containing olistoliths of pelagic Hallstatt limestones. On the other hand, it carried similar mass-transport complexes on its back. The basal thrust of the Dachstein Nappe, of Oxfordian age, was reactivated during the Tertiary.

Rare fragments of metamorphosed and ductilely deformed Triassic-age limestones occur in the Jurassic-age, unmetamorphosed to anchizonal metamorphic mass-transport complexes. They indicate a thermal imprint during Early to Mid-Jurassic times (Gawlick *et al.* 1999) which may be related to ophiolite obduction.

These tectonic processes coeval with sedimentation of chert-rich deep-water deposits in the Callovian to Oxfordian. Following this phase of intensive deformation, a period of reduced tectonic activity continued until the end of the Jurassic. The tectonic contacts were sealed by Tithonian-age basinal (Oberalm Formation) and carbonate platform sediments (Plassen Formation).

The regional framework of Jurassic tectonics in the internal Austro-Alpine units is still under discussion. Obduction of Meliata oceanic crust over the Austro-Alpine continental margin units is suggested by, among other lines of evidence, the occurrence of Early Jurassic amphibolites, probably derived from the metamorphic sole of such ophiolites, as clasts in Upper Cretaceous-age Gosau sediments (Schuster *et al.* 2003). Some authors (e.g. Gawlick & Höpfer 1999) have suggested that the Jurassic tectonism resulted from the collision of the Austro-Alpine margin with another continental terrane. In this collision, the Austro-Alpine margin represented the lower plate which was partly subducted towards the south or SE. Gawlick & Höpfer (1999) found evidence for high-pressure/low-temperature metamorphism in the Pailwand Complex, a part of the Juvavic Nappe System, and dated this metamorphism as Jurassic. This appeared analogous to the blueschist-facies metamorphism of ophiolites in the Meliaticum of the Western Carpathians, which proves Jurassic subduction of the Meliata Ocean there (Maluski *et al.* 1993; Dallmeyer *et al.* 1996; Faryad & Henjes-Kunst 1997). According to Gawlick *et al.* (1999), the rocks in the Pailwand Complex were metamorphosed when the Austro-Alpine continental margin entered the south-dipping subduction zone following the closure of the Meliata Ocean and was subducted below a continental element. There is, however, no agreement in the literature as to which continental fragment formed the overriding plate. In some articles it is unspecified (Gawlick & Höpfner 1999), others favour Tisia (Tizia), a terrane largely hidden under the Pannonian Basin (Froitzheim *et al.* 1996; Channell & Kozur 1997), or a continental fragment comprising the South Alpine units and the upper Juvavic nappes (Schweigl & Neubauer 1997; Neubauer *et al.* 2000). This leads to different positions for a proposed Meliata suture, e.g. between the Austro-Alpine and the South Alpine units, or within the Austro-Alpine units between the Tirolic Nappe System and the upper Juvavic nappes (Schweigl & Neubauer 1997).

Several lines of evidence, however, make a Jurassic continent collision unlikely. The proposed oceanic Meliata suture has not been found in the Alps. The slices of the Meliata zone occur in a structurally high position and there is no evidence of continuation at deeper levels. The proposed Jurassic high-pressure/low-temperature metamorphism in the Juvavic nappes is not generally accepted, because no pressure data exist for metamorphic olistoliths in the Jurassic mass-transport complexes, and, on the other hand, the age of the elevated-pressure metamorphic imprint in the Pailwand Complex is not Jurassic but Cretaceous, according to Frank & Schlager (2005).

Alternatively, it may be assumed that the Austro-Alpine margin did not collide with another continent but entered an intra-oceanic subduction zone, leading to ophiolite obduction onto the continental margin in a similar way as in the Dinarides (e.g. Pamić *et al.* 2002; Stampfli & Borel 2004). After that a system of east–west striking, sinistral strike-slip faults may have developed (Schmid *et al.* 2004*a*; Frank & Schlager 2005). The Juvavic and Tirolic nappe systems, which came from the distal part of the Austro-Alpine margin and carried the Meliata remains, were on the northern side of this fault system. They were transferred towards the west and emplaced to the north of other Austro-Alpine units which represented more proximal parts of the Triassic continental margin. Such elements include the Drauzug-Gurktal Nappe System (Bechstädt *et al.* 1976; Kázmér & Kovać 1985; Tollmann 1987), the Ötztal-Bundschuh Nappe System, the Pelson Unit (Transdanubian Range) and the Southern Alps. The activity of this fault system would have post-dated the Triassic and pre-dated the onset of Mid-Cretaceous thrust tectonics. This is a viable hypothesis explaining the lack of a Meliata suture in the Alps. It is, however, only based on indirect evidence such as the distribution of Triassic facies. Jurassic strike-slip faults have not been identified thus far (in contrast, for example, to Jurassic rift-related detachment faults in the Lower Austro-Alpine units).

## *Cretaceous tectonics*

### *Austro-Alpine units*

Following a period of relative quiescence, tectonic shortening in the Austro-Alpine units recommenced in the Valanginian (*c.* 137 Ma). This was the beginning of the Cretaceous orogenic cycle, often referred to as Eo-Alpine. As noted above, there are various opinions concerning the arrangement of the tectonic units at the time when Eo-Alpine tectonics began. The model of Schmid *et al.* (2004*a*) and Schuster (2004) suggests that a SE- to east-dipping, intracontinental subduction zone developed along the hypothetical Jurassic strike-slip fault, that is, to the north of the future Drauzug-Gurktal and Ötztal-Bundschuh nappe systems. These units formed the upper plate with respect to the subduction zone, whereas the lower plate to the NW can be subdivided into a number of units; Lower Austro-Alpine units represent the most external part. An intermediate area comprised the rocks of the future Bajuvaric Nappe System, underlain by the rocks that later formed the lower nappes of the Greywacke Zone (Vöstenhof-Kaintaleck, Silbersberg and Veitsch nappes). The Silvretta-Seckau Nappe System followed to the SE. Close to the initial position of the subduction boundary was an area that comprised the future Tirolic Nappe System, lying with a stratigraphic contact on Palaeozoic rocks of the Noric Nappe, and already carrying the Juvavic Nappe System on top, which had been emplaced during the Jurassic. The middle and lower crust of this area later formed the Koralpe-Wölz Nappe System.

During early stages of convergence, the Ötztal-Bundschuh and Drauzug-Gurktal nappe systems formed from the upper crust of the subduction zone's upper plate. This thick nappe pile (more than 10 km in the southeastern part) moved westward (Ratschbacher *et al.* 1989) relative to the units of the lower plate. The upper part of the lower plate (Noric Nappe, Tirolic Nappe System, Juvavic Nappe System) was detached from the deeper basement (future Koralpe-Wölz Nappe System). The latter was subducted towards the SE. The detached units formed an orogenic wedge whose early evolution is recorded by synorogenic sediments of the Rossfeld Formation (Hauterivian–Aptian, 120–135 Ma), which were deposited in basins in front and on top of the Tirolic nappes. In the Albian, the southern (upper) nappes of the Bajuvaric Nappe System (Upper Bajuvaric Nappes) and, probably, also the lower nappes of the Greywacke Zone were stripped off from their basement and incorporated into the wedge. At this stage, the Tannheim-Losenstein Basin formed in the northern part of the future Lower Bajuvaric nappes in front of the wedge (Upper Aptian–Lower Cenomanian, 97–105 Ma). The Tannheim-Losenstein basin was partly filled with material from a ridge to the north, where metamorphic basement rocks and serpentinites were exposed ('Rumunic Ridge'; Faupl 1979). Wagreich (2001*a*) suggested that the fomation of this ridge along the northern margin of the Austro-Alpine units against the Penninic Ocean resulted from oblique subduction and the related dextrally transpressive deformation, and that the onset of subduction of the Penninic oceans below the Austro-Alpine units began at that time.

Geochronology suggests that the Eo-Alpine eclogites formed around 92 Ma, i.e. in Turonian times (Thöni 2006). Middle and lower crustal material of the Austro-Alpine units, which had been subducted since Valanginian times, reached a maximum depth at that time. The pressure and temperature data on the Eo-Alpine eclogites show a distinct distribution. The greatest depths were reached in the Saualpe-Koralpe-Pohorje Complex. Up to 2 GPa at 700°C (Miller 1990; Thöni & Miller 1996; Habler 1999) were reached in its northern part, whereas ultrahigh-pressure conditions of 3 GPa at 800°C have been reported by Janàk *et al.* (2004) from the Porhorje mountains in the south. The eclogites formed from Permian-age, N-MORB gabbroic rocks and from metabasalts. Garnet peridotite associated with the eclogites in the

Pohorje mountains yielded up to 4 GPa at 900°C (Hinterlechner-Ravnik *et al.* 1991; Janák *et al.* 2006). To the east and west of the Saualpe-Koralpe-Pohorje Complex the metamorphic peak conditions decrease (Hoinkes *et al.* 1999; Schuster *et al.* 2004) and here the eclogites are derived from metabasite interlayers within pre-Alpine metamorphic complexes.

A major tectonic change can be recognized in the Turonian (*c.* 92 Ma). The eclogite-bearing parts of the Austro-Alpine units were rapidly exhumed to middle crustal levels and formed the central part of the Koralpe-Wölz Nappe System. Cooling of the rocks is dated by Rb–Sr, K–Ar and Ar/Ar ages measured on hornblendes, muscovites and biotites. These are generally in the range of 90 to 60 Ma (Thöni 1999), whereas those of the southern part of the Saualpe-Koralpe-Pohorje Complex in the Pohorje Mountains are Miocene (see below). Exhumation is assumed to have occurred in a tectonic wedge regime (Fig. 18.10) with thrusting in the lower part and normal faulting in the upper part of the wedge (Schuster *et al.* 2004; Schmid *et al.* 2004*a*; Wiesinger *et al.* 2006). In the eastern part of the Eastern Alps a pro-wedge geometry with respect to a SE-directed subduction was established. NW- to north-directed thrusting is documented in the lower part of the wedge, below and within the eclogite-bearing complexes (Krohe 1987), and SE-directed normal faulting in the upper part of the wedge (Fig. 18.13). Several ductile deformed Austro-Alpine nappes of the extrusion wedge exhibit an east- to ESE-trending extensional stretching lineation which may indicate a component of extension oblique to the general extrusion direction. In the southwestern part of the Austro-Alpine units (Texel Mountains and southern Nockberge area), on the other hand, SE-directed extrusion has been suggested (Sölva *et al.* 2001).

Whereas most authors assume exhumation of the Austro-Alpine eclogite-bearing units by buoancy-driven wedge extrusion (Sölva et al 2001; Wiesinger *et al.* 2006), Janák *et al.* (2004, 2006) suggested that slab extraction (Froitzheim *et al.* 2003) was the dominant exhumation mechanism.

The units of the tectonic upper plate, positioned on top of the extruding wedge, were affected by normal faulting. Extensional basins, partly associated with extensional detachment faults (Neubauer *et al.* 1994*a*; Rantitsch *et al.* 2005), developed on top of the Gurktal Nappes and the Graz Palaeozoic Nappes (Krappfeld, St. Paul and Kainach Basins) and were filled with sediments of the Gosau Group (Santonian to Eocene). The clastic material is mainly derived from the surroundings. In addition to detritus of Austro-Alpine provenance, detritus of South Alpine provenance also occurs in the Kainach Basin. This material indicates a close spatial relationship of this basin to the South Alpine units or, alternatively, the existence of klippen with that kind of material on top of the Austro-Alpine nappes during the time of sedimentation (Gollner *et al.* 1987).

At the northern front of the orogen, the basal detachment jumped down to deeper structural levels. The Lower Bajuvaric nappes and the lower nappes of the Greywacke Zone were now incorporated into the orogenic wedge, and this wedge overrode the 'Rumunic Ridge'. In the Turonian (90 Ma), large parts of the Northern Calcareous Alps emerged, resulting in erosion and the formation of bauxite. Subsequently, the Gosau basins developed and were filled with conglomerates and clastic shallow-marine sediments derived from the local surroundings (Lower Gosau Subgroup, Upper Turonian to Campanian).

The Gosau basins of the Northern Calcareous Alps have been interpreted as extensional (Ortner 1994), but most authors assume that they formed in compressional, transpressional or strike-slip regimes (Wagreich 1995; Ortner 2001), in contrast to the extensional Gosau basins on the Drauzug-Gurktal Nappe System. Following a short phase of deformation and erosion, the sedimentation of slope and deep-water clastics of the Upper Gosau Subgroup (Upper Santonian to Eocene) commenced. This was governed by the rapid subsidence of the Bajuvaric and Tirolic nappe systems into bathyal and abyssal depths (Faupl & Wagreich 1996). This facies change took place diachronously from the NW (Santonian) to the SE (Palaeocene). The subsidence pulse has been attributed to an event of tectonic erosion triggered by the subduction of a topographic high of the Penninic oceanic plate (ocean ridge) below the Austro-Alpine margin (Wagreich 1995). An alternative interpretation would be that the frontal part of the Austro-Alpine tectonic wedge passed over the Austro-Alpine–Penninic continental margin and moved onto the accretionary wedge in the oceanic basin to the north. A further alternative suggests that the subsidence resulted from westward retreat (roll-back) of the Piemont-Ligurian subduction zone (Froitzheim *et al.* 1997; Stampfli & Borel 2004).

As mentioned above, most authors assume exhumation of the eclogite-bearing Austro-Alpine units by wedge extrusion following the model of Chemenda *et al.* (1995). This model implies voluminous erosion of metamorphic rocks during exhumation. Larger amounts of Eo-Alpine metamorphic detritus, however, are only documented in Cenomanian- to Turonian-age deep-marine clastic sediments of the Reiselsberg Formation (Rhenodanubian Flysch Zone; Neubauer *et al.* 2007), whereas such detritus is absent in the Late Cretaceous sediments of the Austro-Alpine and South Alpine units prior to the Maastrichtian. In the Maastrichtian and Palaeogene, sandstones rich in detrital white mica, with Eo-Alpine cooling ages, occur in the Upper Gosau Subgroup of several localities (Schuster *et al.* 2003). The late input of this Eo-Alpine upper-greenschist- to eclogite-facies material may be explained as follows. The P-T-t paths of the Eo-Alpine eclogite-bearing units were characterized by rapid (in some cases isothermal: Hoinkes *et al.* 1999) exhumation from eclogite facies to amphibolite facies and a subsequent slow cooling (Thöni 1999). This may indicate that the rocks were exhumed rapidly by 'blind' wedge extrusion in the Late Cretaceous and then stored in an upper crustal position without reaching the surface. The later exhumation to the surface would have been controlled by other processes, including erosion and block tilting (see below). The slab extraction model as proposed by Janák *et al.* (2004, 2006), on the other hand, does not necessitate erosion of metamorphic rocks during exhumation.

As noted above, subduction of the Penninic Ocean beneath the Austro-Alpine units probably commenced in Albian time (Wagreich 2001*a*). Between the downgoing Penninic slab and the Upper Austro-Alpine nappes, the Lower Austro-Alpine nappes represent slices of continental crust which entered the subduction zone at different times and reached different depths. A decrease in metamorphic grade from lower greenschist facies at the top to anchizonal conditions at the base can be recognized in the Radstadt Nappe System. The formation ages of white micas that formed during the metamorphic event decrease from Cretaceous to Miocene in the same direction (Slapansky & Frank 1987; Liu *et al.* 2001). At the eastern border of the Alps an inverted metamorphic field gradient can also be established. In the lowermost Wechsel Nappe (Lower Austro-Alpine), only newly formed white micas within shear bands yield Cretaceous ages, whereas in the overlying Semmering Nappe (Lower Austro-Alpine) at least partly rejuvenated Ar/Ar ages are found in pre-Alpine metamorphic rocks, and muscovite and biotite from greenschist facies metamorphic Mesozoic rocks yield formation ages of about 82 Ma (Müller *et al.* 1999; Dallmeyer *et al.* 1998).

In this area, the Upper Austro-Alpine nappes were emplaced on the Lower Austro-Alpine Wechsel Nappe by north-directed out-of-sequence thrusting (Willingshofer & Neubauer 2002).

In the westernmost part of the Austro-Alpine nappes (Silvretta Nappe and units to the south of it), Cretaceous tectonics commenced with west-directed thrusting (*c.* 100 to 90 Ma), followed by sinistral transpressive shearing in an east–west striking corridor, and by top-to-the-ESE extensional faulting (*c.* 80–67 Ma) (Froitzheim 1992; Froitzheim *et al.* 1994; Handy 1996). This deformation affected not only the Austro-Alpine units, but also the underlying Upper Penninic nappes (Platta Nappe and Margna Nappe). Thrusting may have been diachronous to some extent, older in the structurally higher units and younger in the deeper ones. The Late Cretaceous episode of extensional faulting affected the entire nappe stack from the Ötztal Nappe at the top to the Margna Nappe at the base. Some of the top-to-the-ESE extensional faults are detachment faults with large offsets, e.g. the Schlinig Fault at the western border of the Ötztal Nappe (Froitzheim *et al.* 1997) which had previously been regarded as a thrust (Schmid & Haas 1989). At deep levels of the nappe stack, recumbent folds formed during Late Cretaceous extension as a result of the collapse of initially steeply orientated layers (Froitzheim 1992; Handy 1996).

*South Alpine units*

South-directed, thick-skinned thrusting and folding in the internal part of the Lombardic-Giudicaric thrust belt pre-dates the oldest part of the Adamello intrusion and must therefore be older than *c.* 43 Ma (De Sitter & De Sitter Koomans 1949; Brack 1981). The 64 Ma K–Ar amphibole/groundmass age of a dyke cutting one of the thrusts (Zanchi *et al.* 1990) suggests a Cretaceous age, older than 64 Ma, for the thrusting event. However, in view of the many K–Ar ages published in the Alpine literature which have turned out to be geologically meaningless, thrusting prior to the intrusion of the Adamello pluton could also be Palaeocene to Early Eocene and could thus be correlated with the Dinaric thrusting observed in the Dolomites. Hence, there is no unequivocal evidence for Cretaceous thrusting in the Southern Alps. There is some evidence from sediment facies distribution, however, that narrow elongate horsts and grabens formed in a transpressional setting in the area of the Giudicarie Fault (Doglioni & Bosellini 1987).

*Penninic nappes*

It has long been assumed that the high-pressure metamorphism in the Penninic units of the Alps was Cretaceous in age and, consequently, that subduction-related thrusting affected these units during the Cretaceous. This belief was mostly based on K–Ar and $^{40}Ar/^{39}Ar$ data affected by excess argon. The application of U–Pb, Sm–Nd and Lu–Hf dating resulted in Tertiary ages (e.g. Tilton *et al.* 1991; Becker 1993; Gebauer 1996; Duchêne *et al.* 1997). It is now generally accepted that the high- and ultrahigh-pressure metamorphism in the Penninic nappes is Tertiary in age. The only exception to the Tertiary age is the Cervinia Terrane, where eclogite-facies metamorphism occurred at *c.* 65 Ma and prograde metamorphism preceded the eclogite facies at c. 76 Ma (Sesia Nappe; Rubatto *et al.* 1999). Hence, the Cervinia fragment entered the Piemont-Ligurian subduction zone at *c.* 76 Ma. The accretionary prism of the Tsaté Nappe would thus have formed between *c.* 100 Ma, when subduction of the Piemont-Ligurian Ocean began, and 76 Ma, when the Cervinia Terrane arrived at the trench. However, there is still no geochronological proof for this assumption.

In the Valais Basin, oceanic spreading took place during the Cenomanian and Turonian, as evidenced by metagabbros with an age of *c.* 93 Ma in the Chiavenna Ophiolite (eastern Central Alps; Liati *et al.* 2003) and in the Balma Unit (western Central Alps; Liati & Froitzheim 2006). Spreading may also have occurred earlier in the Cretaecous, as Lower Cretaceous (Barremian or older) sediments seal the transition between thinned continental crust and exhumed mantle material in the Tasna Nappe derived from the southern margin of the Valais Ocean (Florineth & Froitzheim 1994), indicating that continental break-up took place there in the Early Cretaceous. Late Cretaceous oceanic spreading was strongly transtensional, related to sinistral displacement of the Iberia-Briançonnais microcontinent relative to Europe.

It is not totally clear when the subduction of the Valais Ocean began. In the Western and Central Alps, the Valais Ocean was consumed during the Palaeogene in its own subduction zone which dipped southeastward, under the Briançonnais fragment. In Eocene times, the Iberia-Briançonnais microcontinent was an independent tectonic plate and the northeastern spur of this plate (the Briançonnais fragment) was bordered by SE-dipping subduction zones on both sides (Stampfli & Borel 2004). The Valais subduction zone was connected to the west across the Pyrenees with the subduction zone that consumed the southern part of the Bay of Biscay oceanic crust beneath Iberia. Subduction of the Valais Ocean may have commenced in the Late Cretaceous, if it is assumed that the Upper Cretaceous Niesen Flysch, in the external part of the Valaisan, was deposited in a subduction-related trench. However, the Valais Ocean cannot have been completely consumed in the Cretaceous, because marine sedimentation in the Valais Basin extends into the Eocene (Ziegler 1956; Bagnoud *et al.* 1998). Thus the question of Cretaceous subduction of the Valais Ocean is still open. We prefer to assume that subduction began in the Palaeogene, because of the lack of direct evidence for Cretaceous subduction, that is, radiometric ages for subduction-related metamorphism. Radiometric evidence exists only for Eocene subduction of the Valais Ocean (Liati *et al.* 2005; Liati & Froitzheim 2006).

Stratigraphic evidence from the Briançonnais and Valais units indicates tectonic activity in the Late Cretaceous. This was the case for the internal Briançonnais in the western Alps, where Late Cretaceous breccias and olistoliths occur (Jaillard 1999), as well as for the Tasna Nappe in the Engadine Window. In the latter area, the Cretaceous succession from the Tristel Formation (Barremian-Lower Aptian) through the 'Gault' sandstones and shales up to the Late Cretaceous–Palaeogene 'Couches Rouges' received coarse-grained clastic input, including both basement and sediment clasts, from a nearby escarpment that was located to the south. Olistoliths, several tens of metres in size, were shed down the escarpment in Late Cretaceous–Palaeogene times. This escarpment was probably related to a strike-slip fault, since a thrust fault would have led to the closure of the sedimentary basin after a certain time. Breccias of probable Cretaceous age have also been reported from the Schams Nappes (Schmid *et al.* 1990), but their exact age is unclear. Similar lithological associations are observed in the Valaisan of the Western Alps. There, the Upper Cretaceous sediments contain subaqueous mass-flow deposits (Brèche de Tarentaise) shed from a rapidly eroding source area to the west (Lomas 1992).

*European margin*

During the Cretaceous, a series of folds trending approximately east–west developed in the Diois, Baronnies and Dévoluy areas of the Chaînes Subalpines, SW of the Pelvoux Massif. The folds are overlain by Late Cretaceous sediments along a spectacular unconformity which was itself later deformed by folds and thrusts reflecting east–west shortening in the external part of the Western Alpine arc. The age of folding has been suggested to be Turonian (Mercier 1958) or Coniacian (Flandrin 1966). The north–south orientated contraction is associated with NE–SW striking, sinistral strike-slip faults (Arnaud 1973). The folding preceded the Pyrenean-Provençal deformation resulting from the collision of Iberia with Europe.

In conclusion, Cretaceous tectonics in the Penninic units and more external zones was characterized by oceanic subduction in the southeastern sub-basin of the Piemont-Ligurian Ocean, probably beginning at *c.* 100 Ma (Albian), by the collision of the Cervinia Terrane with the upper plate of this subduction zone at *c.* 76 Ma (Campanian), by strike-slip faulting in the Briançonnais, by sinistrally transtensive oceanic accretion in parts of the Valais Ocean until 93 Ma (Cenomanian–Turonian), and by north–south shortening in the southwestern part of the European margin at *c.* 90 to 85 Ma.

## *Tertiary tectonics*

The Tertiary period was characterized by the consumption of the remaining oceanic spaces and continental collision in the Alps. Following collision, extensional faulting and strike-slip faulting, alternating or coeval with crustal shortening, took place during ongoing continent convergence.

### *Central Alps*

The southeastern sub-basin of the Piemont-Ligurian Ocean (Tsaté sub-basin, Fig. 18.3) had, at least partly, been consumed in the Cretaceous. Subduction in the Zermatt-Saas sub-basin occurred during the Palaeocene to Eocene and is constrained by radiometric ages for the eclogite-facies and locally ultrahigh-pressure metamorphism in the Zermatt-Saas Zone ranging between 49 and 40 Ma (Ypresian-Lutetian; Rubatto *et al.* 1998; Lapen *et al.* 2003). The Zermatt-Saas sub-basin was subducted towards the SE beneath the upper plate, which included the Cervinia Terrane and Apulia.

As noted above, it is not clear when the Valais Ocean began to be subducted. At 40 to 37 Ma, ophiolites of this ocean had reached eclogite-facies depth. Ages in this range were determined for eclogite-facies metamorphism of Valaisan ophiolites in the Central Alps (Fig. 18.4; Liati *et al.* 2003, 2005; Liati & Froitzheim 2006). The subduction zone dipped below the Briançonnais fragment (Fig. 18.14). Rocks of the distal European margin must have arrived at the subduction zone a few million years before 40 Ma, in order to leave sufficient time for distal European margin units such as the Adula Nappe and the Monte Rosa Nappe to reach eclogite-facies conditions at *c.* 40 Ma (Figs 18.4 & 18.14). There is some evidence that the subduction zone was not located directly at the boundary between the Valais Ocean and the Briançonnais fragment, but was rather intra-oceanic, within the Valais Ocean (Fig. 18.14). This evidence comes from two sources. Firstly, the ocean–continent transition between the Briançonnais fragment and the Valais Ocean preserved in the Tasna Nappe was not disrupted by subduction but preserved within this nappe until the present day, and subduction-related shearing and metamorphism are only found at a deeper level in the Bündnerschiefer of the Engadine Window (Bousquet *et al.* 1999). Secondly, Bagnoud *et al.* (1998) reported late Middle Eocene fossils in sediments from the transition area between the Briançonnais fragment and the Valais Ocean in western Switzerland, indicating that this area was still open at *c.* 40 Ma when the distal European margin had already entered the subduction zone.

Tertiary deformation in the Penninic units was dominated by thrusting until about 35 Ma. This thrusting was initially thin-skinned, when sedimentary cover sheets were detached from their oceanic or continental basement, and subsequently thick-skinned, when the continental basement was shortened. Thrusting, both thin- and thick-skinned, was generally directed towards the European foreland. Overprinting of early thin-skinned by thick-skinned, out-of-sequence thrusting led to an earliest phase of nappe refolding. Examples of this include the folding of the Valaisan Bündnerschiefer, representing the detached sediment cover of the Valais Ocean, around the frontal fold of the Adula Nappe in the eastern Central Alps (Schmid *et al.* 1996), and the analogous folding of the Valaisan Antrona Ophiolites around the front of the Monte Rosa Nappe in the western Central Alps (Fig. 18.9; Froitzheim 2001). According to a new interpretation (Froitzheim *et al.* 2006; Pleuger *et al.* 2007), the basal thrust of the Sesia and Dent Blanche nappes over the Tsaté Nappe is also out of sequence, and the Tsaté Nappe was originally in a higher structural position than the Sesia and Dent Blanche nappes.

At about 35 Ma, the kinematics of the Penninic nappes changed from foreland-directed thrusting to extension. The extension direction was initially parallel to the orogen, as in the case of top-to-the-east normal shearing along the Turba Normal Fault (Figs 18.5 & 18.15) at the eastern border of the Lepontine Dome (Nievergelt *et al.* 1996). Orogen-parallel extension is also reflected by top-to-the-SW shearing in the area of the Monte Rosa Nappe (Steck 1990; Pleuger *et al.* 2007) and other parts of the Penninic nappe stack to the west of the Lepontine Dome. These two large-scale shear zones with opposite dips and shear senses accommodated early exhumation of the Lepontine Dome between them. The Lepontine nappes themselves were also affected by this orogen-parallel stretching. This was mainly expressed by the formation or modification of large-scale isoclinal recumbent folds with hinges parallel to the stretching lineation. Examples are the Antigorio Nappe (D2 of Grujic & Mancktelow 1996) and the modification of the frontal fold of the Adula Nappe during pronounced ENE–WSW orientated stretching in the 'Leis phase' (Löw 1987).

During this process of extensional shearing, a slight change in the extension direction occurred in the Swiss-Italian part of the Penninic nappes (Fig. 18.15). The direction of extension became top-to-the-SE, oblique to the strike of the chain. In the Monte Rosa region, orogen-parallel top-to-the-SW shearing was followed by top-to-the-SE shearing (Pleuger *et al.* 2007). The latter was most pronounced along the Gressoney Shear Zone (Reddy *et al.* 1999; Wheeler *et al.* 2001) where it reactivated an earlier top-to-the-NW thrust which had emplaced the Tsaté Nappe on the Zermatt-Saas Zone. The shear zone dips to the SE beneath the Sesia Nappe; its geometry is extensional. Towards the east, the shear zone followed the southern border of the Lepontine Dome whose southern parts were penetratively affected by top-to-the-SE shearing (e.g. the southern parts of the Adula and Simano nappes; Nagel *et al.* 2002). The shear zone thus acted as an oblique normal fault in a south-dipping orientation. Hence, part of the exhumation of the Lepontine Dome was not accommodated by backthrusting but by normal faulting. Later, beginning at *c.* 30 Ma, the normal fault was rotated into the orientation of a backthrust and pervasively overprinted by ongoing mylonitic shearing until *c.* 26 Ma (Schärer *et al.* 1996). Further east, the 'Insubric normal fault' curved away from the Periadriatic Fault towards the north (33 to 30 Ma Preda Rossa Shear Zone; Berger *et al.* 1996) and was probably connected

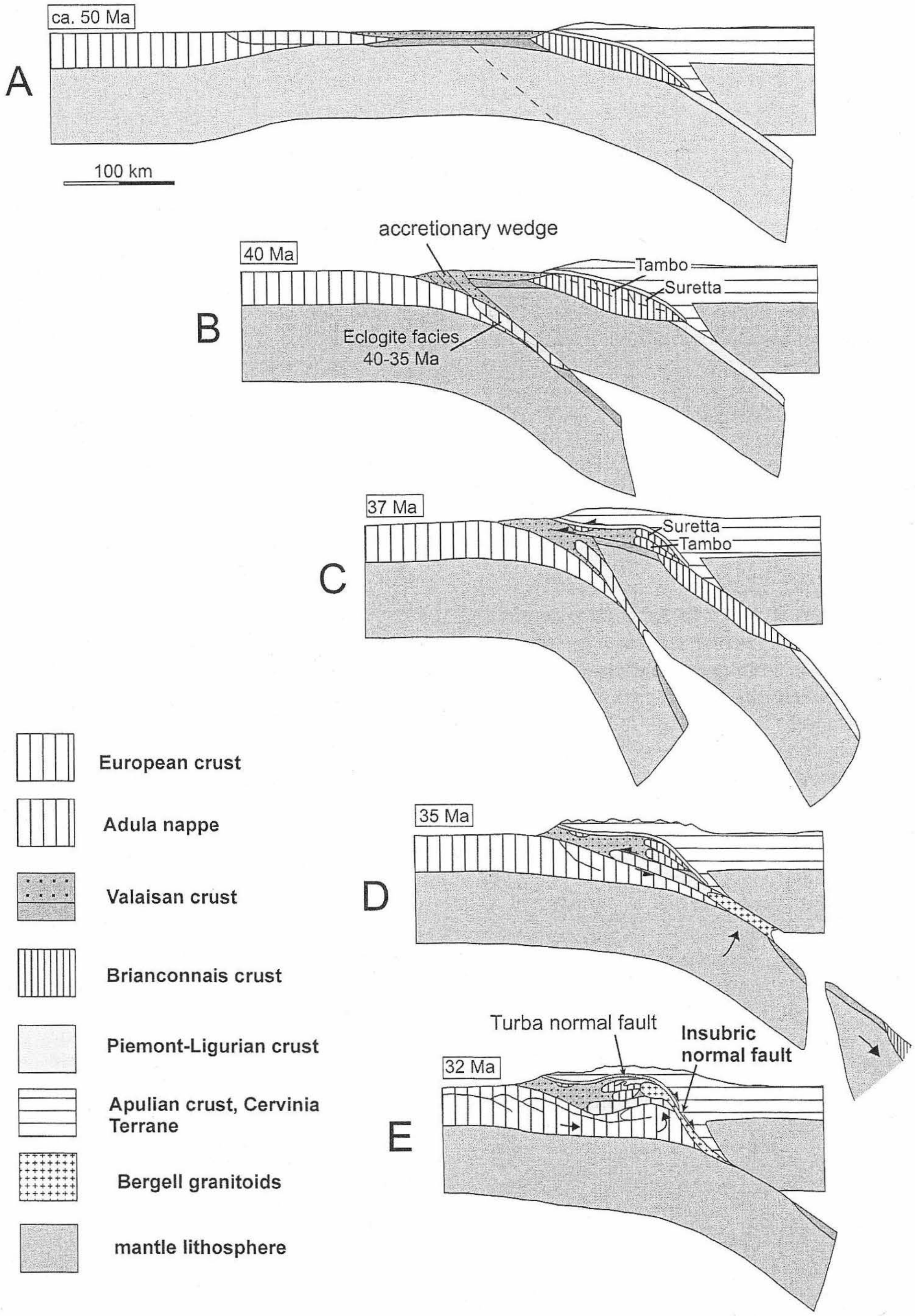



**Fig. 18.14.** Reconstructed kinematic evolution of the Central Alps along the cross-section of Figure 18.6. Exhumation of the Adula nappe high-pressure rocks was achieved by subduction of the overlying microplate including the Briançonnais Terrane (slab extraction) between steps B and D.

with the Turba Normal Fault. This connection was interrupted by the intrusion of the 30 Ma Bergell granodiorite which truncates the Turba Normal Fault (Nievergelt *et al.* 1996) and, thereby, sets a lower age limit to the top-to-the-SE extensional shearing.

Although the geometry of top-to-the-SE shearing is extensional in the Penninic nappes, it does not reflect plate divergence across the Alps but rather it occurred contemporaneously with shortening and foreland-directed thrusting in the more external units (Helvetic units; Milnes & Pfiffner 1980; Weh & Froitzheim 2001). Therefore, it may reflect corner flow during plate convergence. Important backfolds formed during the extensional event, most notably the Niemet-Beverin Fold (Fig. 18.6) in the eastern, and the Mischabel Fold in the western Central Alps. These are not related to compressional 'backthrusting' along the Insubric line but to extensional 'backshearing' since the Insubric Fault was not yet a backthrust at that time but rather a south-dipping normal fault (Figs 18.6, 18.10 & 18.14; Nagel *et al.* 2002).

The Helvetic units of the Central Alps evolved in a similar fashion to the Penninic units, with an early phase of thin-skinned

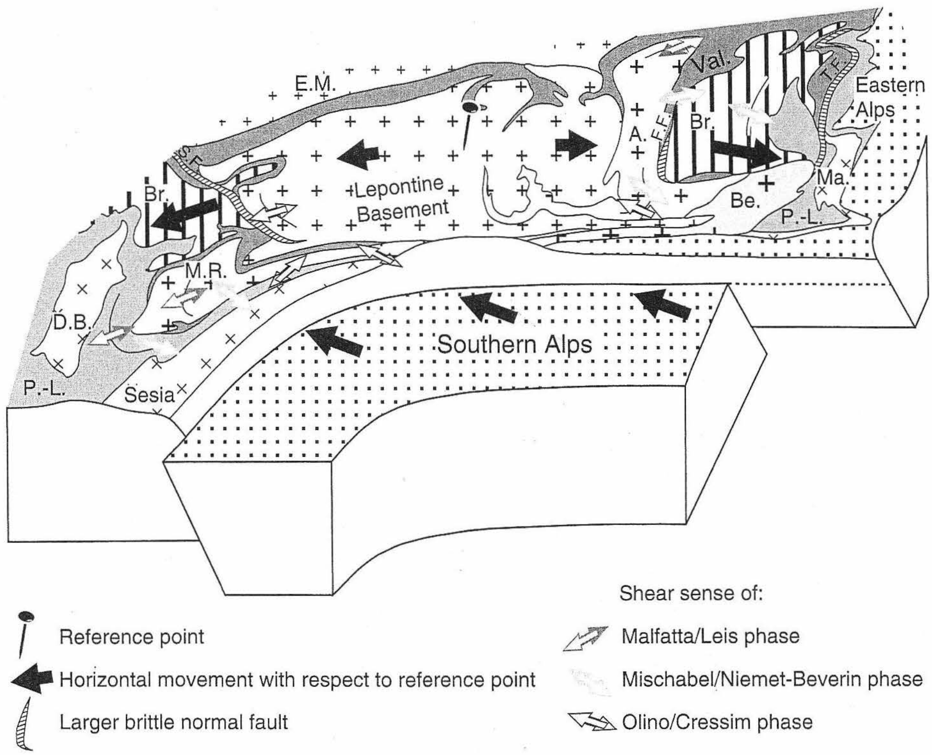



**Fig. 18.15.** Schematic block diagram illustrating kinematics of the Central Alps and the Insubric Fault during Oligocene to Miocene times. Black arrows indicate overall motions with respect to a reference point in the Lepontine Dome. Double arrows indicate shear sense during Oligocene to Miocene phases of extensional deformation in the Penninic units: orogen-parallel extension during the Malfatta/Leis phase (*c.* 35–33 Ma), SE- to east-directed extension during the Mischabel/Niemet-Beverin phase (*c.* 33–30 Ma), again orogen-parallel extension during the Olino/Cressim phase (after 30 Ma), progressively localized in normal faults (Simplon and Forcola faults). Extensional deformation of the Penninic units occurred during convergence of Europe and Adria and accommodated exhumation of the Penninic units relative to the Southern Alps. Abbreviations: A., Adula Nappe; Be., Bergell intrusion; Br., Briançonnais units; D.B., Dent Blanche Nappe; E.M., External Massifs; F.F., Forcola Fault; Ma., Margna Nappe; M.R., Monte Rosa Nappe; T.F., Turba Fault; P.-L., units derived from Piemont-Ligurian Ocean; Val., units derived from Valais Ocean. After Pleuger *et al.* (2008).

thrusting followed by later phases of thick-skinned thrusting. However, the entire sequence began later and lasted longer. Synorogenic extension, as observed in the Penninic nappes, is either absent or at least weaker. In the eastern Central Alps, thrusting in the Helvetic units began in the latest Eocene to earliest Oligocene (*c.* 35 Ma) with the emplacement of the Ultra-Helvetic and South Helvetic units (Sardona Flysch, Blattengrat Flysch) on top of the units that would later form the Helvetic nappes and the Infra Helvetic Complex. This emplacement was coeval with extension in the Penninic nappes. The contemporaneity may be interpreted as indicating gravitational spreading at the scale of the orogen as suggested by Milnes & Pfiffner (1980).

In the Penninic units, the extensional phase lasted from *c.* 35 Ma to 30 Ma (Schreurs 1993; Schmid *et al.* 1996). It was followed by NW–SE to north–south compression associated with orogen-parallel extension, creating folds with steep axial planes like the Vanzone Antiform (east of Monte Rosa) and the Cressim Antiform (southern Adula Nappe) (Nagel *et al.* 2002). The formation of these folds was related to the development of the Southern Steep Belt adjacent to the Periadriatic Fault (Milnes 1974). Dextral shearing along the Periadriatic Fault (Schmid *et al.* 1987) began at *c.* 30 Ma in the Central Alps.

At this time, the thrusts that define the Helvetic nappes, such as the Glarus Thrust, also formed. North-directed thrusting along the Glarus Thrust must have begun after 30 Ma because the 32 to 29 Ma Taveyannaz sandstone (Boyet *et al.* 2001) is present in the footwall of this thrust. To the north, the thrust continues into the frontal thrust of the Helvetic nappes which buried Miocene sediments of the Molasse Basin in its footwall. The activity of the Glarus Thrust must therefore have continued during the Miocene. Further south, the Glarus Thrust buried the earlier-emplaced Blattengrat and Sardona Flysch units in its footwall; hence it is an out-of-sequence thrust. It is rooted to the south in the boundary zone between the Aare Massif and the Gotthard Nappe. During the Miocene, the Glarus Thrust was deformed into its present antiformal shape by the updoming of the Aare Massif. This updoming was the result of underthrusting from the north. The Aare Massif can be viewed as an antiformal stack of basement units imbricated during this stage.

In Miocene times, a second phase of extensional tectonics affected the internal part of the Central Alps, south of the Helvetic units. This led to the development of normal faults with slip direction parallel to the orogen. Typical examples include the Simplon Fault (Fig. 18.5; Mancktelow 1985), the Forcola Fault (Meyre *et al.* 1998; Ciancaleoni & Marquer 2006), and the Engadine Fault (Schmid & Froitzheim 1993). These faults are related to the exhumation of Penninic metamorphic domes: the Lepontine Dome in the case of the Simplon and Forcola faults, and the Engadine Window in the case of the Engadine Fault.

The Simplon Fault is a large-scale normal fault with a wide zone of greenschist-facies mylonitic shearing in the footwall, evolving upwards into cataclasites and finally into a discrete fault. Top-to-the-SW normal faulting along the Simplon Fault was contemporaneous with NW–SE shortening. This is demonstrated from slightly older, but already Simplon-Fault-related mylonites in the footwall that were deformed into SW–NE striking folds and truncated by younger mylonites of the Simplon Fault (Mancktelow 1992). The main extensional movements along the Simplon Fault occurred at *c.* 18 to 15 Ma but displacement went on at a lower rate until 3 Ma (Grasemann & Mancktelow 1993). The extension direction of the Simplon Fault is parallel to the direction of Early Oligocene orogen-parallel extension (see above). Top-to-the-SW greenschist-facies mylonites found in the hanging wall of the Simplon Fault, which are related to the Early Oligocene orogen-parallel extension, were therefore interpreted by Steck (1990) to be part of the wider 'Simplon Shear Zone', assuming continuity between Early Oligocene and Miocene extensional faulting.

Although it seems logical from map view that the Simplon Fault is connected with the Insubric Fault through the Centovalli, a west–east striking zone of cataclasite, this is in fact not the case. Instead, the Simplon Fault becomes a ductile shear zone towards the east, then turns to the NE into the Lepontine Dome and becomes indistinct (Mancktelow 1985). The Simplon and Insubric faults are not directly linked (Mancktelow 1985; Schmid *et al.* 1987). Towards the west, the Simplon Fault is connected with a dextral strike-slip fault in the Rhone Valley. This fault is assumed to continue into the Western Alps as a shear zone between the Aiguilles Rouges and Mont Blanc massifs (Hubbard & Mancktelow 1992).

The east-dipping Forcola Fault (Meyre *et al.* 1998) is the symmetrical counterpart of the Simplon Fault on the eastern side of the Lepontine Dome but the amount of displacement is smaller. At its southern termination, Forcola-Fault-related shearing affected the 24 Ma old (Liati *et al.* 2000) Novate leucogranite intrusion, probably soon after its intrusion (Ciancaleoni & Marquer 2006). Together, the Simplon and Forcola faults accommodated Miocene exhumation of the Lepontine Dome by east–west extension.

The SW–NE striking Engadine Fault (Fig. 18.5) shows a combination of strike-slip and normal-fault motion (Schmid & Froitzheim 1993). In the Lower Engadine, towards the NE, it has a SE-dipping normal-fault orientation and accommodated relative uplift and exhumation of the Penninic nappes of the Engadine Window in its footwall. Its vertical component in this area amounts to at least 4 km. Towards the SW in the Upper Engadine, it becomes vertical and acted as a sinistral strike-slip fault with a vertical component opposite to the one in the Lower Engadine. Here, it is the southeastern block which was uplifted relative to the northwestern one. Activity of the Engadine Fault post-dates the emplacement of the Bergell granodiorite at 30 Ma (Von Blanckenburg 1992), since it truncates the contact metamorphic aureole of the Bergell intrusion. The continuations of the Engadine Fault towards the NE and SW are still unclear. Towards the NE, it may continue into the fault that separates the Ötztal Nappe to the south from the Northern Calcareous Alps to the north.

During the Miocene, shortening of the European crust prograded towards the north. The basement of the Aare Massif was progressively shortened by ductile to brittle thrust faults; the lowermost of these is probably connected with the frontal thrust fault of the Alps, which is the thrust of the Subalpine Molasse over the Foreland Molasse. The Subalpine Molasse forms a stack of imbricates dipping towards the south under the Helvetic nappes. Movement of the frontal thrust lasted at least until the end of the Burdigalian (*c.* 16 Ma) since the Burdigalian Upper Marine Molasse is affected by thrust-related deformation. Seismic and geological field studies suggest the existence of a south-directed backthrust at depth to the north of the frontal thrust and cropping out a short distance to the north of the frontal thrust. Together, this backthrust and the frontal thrust form a triangle structure (Fig. 18.6; Stäuble & Pfiffner 1991; Berge & Veal 2005). The backthrust was active at *c.* 15 to 13 Ma (Kempf *et al.* 1999).

The Jura Mountains are kinematically linked to the Alps and represent the youngest part of the north-directed Alpine thrust belt. The age of folding and thrusting in the Jura Mountains is Late Miocene to Pliocene. Younger, Pleistocene to recent shortening has been observed in the foreland of the Jura frontal thrust to the west of Basel. At its southwestern end, the Jura folds and thrusts continue into the most external folds and thrusts of the Western Alps. At its eastern termination, the Jura shortening dies out and the Jura Mountains end without a connection to the Alps, at least at the surface. The Jura folds represent classic detachment folds; they were detached along Middle and Upper Triassic evaporites. Although it is clear that shortening in the Jura Mountains is part of the Alpine Orogeny, there is a long-standing controversy about whether the basal thrust is rooted in the basement below the Jura Mountains or runs within the Mesozoic-age sediments below the fill of the Molasse Basin and is rooted below the Alps ('Fernschubhypothese', e.g. Laubscher 1961). Based on seismic sections of the Molasse Basin, Pfiffner *et al.* (1997) suggested that the Jura basal thrust is indeed rooted in the basement under the Jura Mountains, and continues southward at a shallow level within the basement beneath the Molasse Basin. Thrusting in the Jura Mountains also involved inversion of Late Carboniferous to Permian basins.

*Eastern Alps*

The tectonic evolution of the Eastern Alps during Tertiary times was governed by four major processes: closure of remaining parts of the Penninic Ocean by subduction (until *c.* 47–40 Ma; Kurz *et al.* 2001*a*); thrusting of Austro-Alpine and Penninic units over the European continental margin and imbrication of the latter (from *c.* 47 to *c.* 17 Ma); Oligocene synintrusive shearing along the Periadriatic Fault (*c.* 35 to 28 Ma); east–west extension and eastward extrusion of Austro-Alpine crustal fragments, coeval with north–south shortening (from 25–20 Ma to *c.* 9 Ma; Decker & Peresson 1996) (Fig. 18.16).

Subduction of the Penninic Ocean continued from the Late Cretaceous into the Palaeogene, and parts of this ocean remained open until the middle Eocene when the youngest sediments were deposited. Sedimentation ended in the Penninic units of the Engadine Window in the western part of the Eastern Alps at *c.* 49–48 Ma (Ypresian–Lutetian boundary; Bertle 2002), and in the Laab Nappe, one of the Rhenodanubian Flysch nappes in the eastern part of the Eastern Alps, at *c.* 40 Ma (end of Lutetian; Piller *et al.* 2004). This indicates eastward-propagating closure of the Penninic Ocean and accretion of the oceanic units under the Austro-Alpine nappes. Subduction of the distal European margin may have begun earlier, since $^{40}Ar/^{39}Ar$ geochronology on high-pressure rocks from the Eclogite Zone (Sub-Penninic) of the Tauern Window yielded middle Eocene ages (*c.* 42 Ma) for early exhumation after the pressure peak (Ratschbacher *et al.* 2005). However, a much younger age has also been proposed for the high-pressure metamorphism in the Eclogite Zone (*c.* 31.5 Ma using Rb/Sr; Glodny *et al.* 2005).

The Austro-Alpine units in the upper plate of the subduction

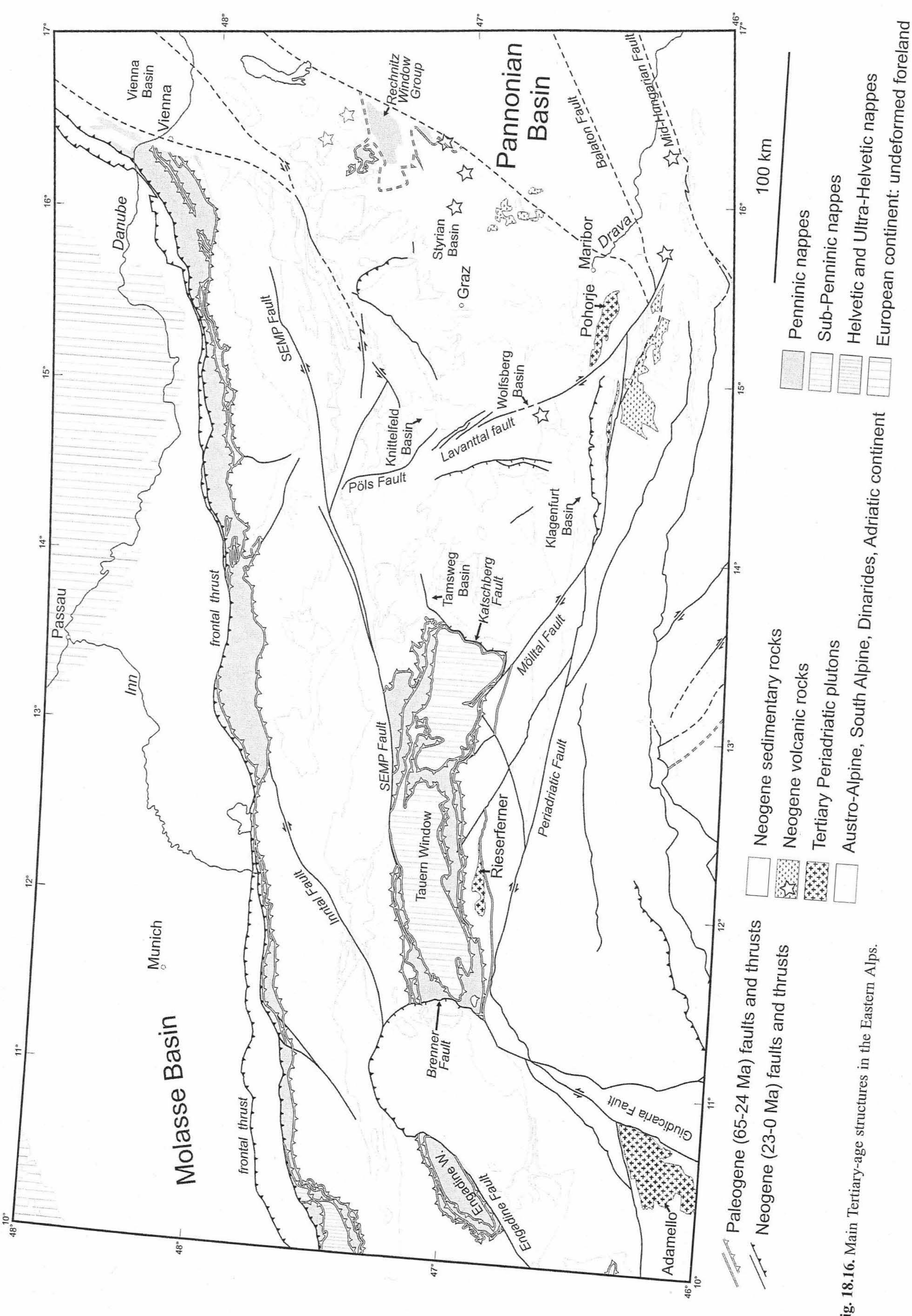



**Fig. 18.16.** Main Tertiary-age structures in the Eastern Alps.

zone represented a flooded (and, in the southern part, extended) mountain belt where the highest peaks formed belts of islands. Between these islands and on the northern slope towards the subduction zone, sedimentation of the Gosau Group continued until the Early Eocene (Wagreich 2001*b*). The collision with the European continental margin resulted in uplift of the upper plate and the termination of marine sedimentation of the Gosau Group. As noted above, Eo-Alpine metamorphic crystalline detritus can be observed in mica-rich sandstones of the Gosau Group from the uppermost Cretaceous onward (Schuster *et al.* 2003). The oldest apatite fission track data from the Austro-Alpine basement units are over 60 Ma (Dunkl 1992; Hejl 1992; Fügenschuh *et al.* 2000). Hence, the metamorphic cycle and pervasive deformation of most of the Austro-Alpine units was completed at that time, with the exception of some parts located to the south, which were significantly exhumed during the Tertiary. Examples for this include the Austro-Alpine basement hosting the Rieserferner Pluton between the Defereggen-Antholz-Vals Fault and the southern border of the Tauern Window (Steenken *et al.* 2002) and the basement in the Pohorje Mountains (Fodor *et al.* 2003).

The sedimentary succession of the Ultra-Helvetic Liebenstein Nappe, representing the distal European continental margin in the western part of the Eastern Alps, ends at *c.* 47 Ma (Piller *et al.* 2004). This datum approximately marks the beginning of continental collision in the Eastern Alps. After that time, the Ultra-Helvetic units were accreted to the nappe stack consisting of Rhenodanubian Flysch nappes and Austro-Alpine nappes. In the eastern part of the Eastern Alps, sedimentation in the Helvetic/Ultra-Helvetic units continued until *c.* 38 Ma, which reflects oblique, eastward-propagating collision (Piller *et al.* 2004). When the Helvetic units were detached from their basement, beginning at *c.* 38 Ma in the western part, thrust-related folds in the Helvetic units deformed the older basal thrusts of the Ultra-Helvetic and Rhenodanubian Flysch nappes. Finally, thrusting propagated into the southern part of the Molasse Basin (Folded Molasse) beginning at *c.* 24 Ma (Oberhauser 1998). The Folded Molasse is formed by northward-overthrust synclines; the intervening anticlines are cut by the thrust faults. This structure represents closely spaced fold-propagation folds. At the southern rim of the Foreland Molasse, i.e. the part of the Molasse Basin located north of the Folded Molasse, the strata dip to the north and are floored by a north-dipping backthrust. Together with the southward-dipping basal thrust of the Folded Molasse, this backthrust forms a triangle structure (Berge & Veal 2005) which can be followed over 350 km from eastern Switzerland through Bavaria into Austria. Thrusting along the northern rim of the Eastern Alps continued until *c.* 17 Ma, the age of the youngest overthrust sediments (Decker & Peresson 1996). Blind thrusts at depth may still have been active after this time.

In the eastern part of the Alps, several tonalitic, granodioritic and granitic intrusions are found in the vicinity of, or directly at, the Periadriatic Fault including: Adamello Pluton, Rensen Pluton, Rieserferner Pluton, Karawanken Pluton, Pohorje Pluton and several smaller fault-parallel tonalite lamellae intruded into the middle and upper crust. Subvolcanic dykes and extrusive volcanics with tonalitic and basaltic compositions reached the uppermost crust and the surface. The basaltic rocks show alkaline and calc-alkaline signatures (Deutsch 1984; Müller *et al.* 1992). Except for the southern part of the Adamello Pluton (45 Ma) and the Pohorje Pluton, all granitoid intrusions occurred in a rather short timespan between 34 and 28 Ma (Rosenberg 2004). The intrusion of the Pohorje Pluton was dated by K–Ar at 18 to 16.5 Ma (Marton *et al.* 2002); U–Pb or other data from high-retentivity systems are not yet available for this pluton.

There is obviously a close relationship between the Periadriatic Fault and the intrusions (this is also the case for the Periadriatic intrusions in the Western and Central Alps). Either there was a linear magma source (von Blanckenburg & Davies 1995), or the magma rising from a wider source area was channelled into the fault zone (Rosenberg 2004). Von Blanckenburg & Davies (1995) suggested that the Periadriatic magmatism resulted from break-off of the subducted Penninic Ocean slab at 45 Ma, leading to melting of supra-subduction-zone lithosphere. Although many authors have adopted the slab break-off hypothesis, there is no consensus about the timing. Slab break-off events have been proposed to have taken place in the Alps at 100 Ma and 45 Ma (Von Blanckenburg & Davies 1995, 1996), 35 to 30 Ma (Schmid *et al.* 1996), 25 Ma (Michon *et al.* 2003), and 5 Ma (Schmid *et al.* 2004*b*). This inflation of events sheds serious doubt on the testability of the slab break-off hypothesis in the case of the Alps. On the other hand, slab break-off is still an elegant model to explain short-time, intense magmatism in a long and narrow zone. Froitzheim *et al.* (2003) suggested slab extraction as a cause of the Periadriatic magmatism, which would be quite similar to slab break-off regarding the thermal consequences.

Along the Periadriatic Fault System in the Eastern Alps, important movements took place in the Oligocene, contemporaneously with Periadriatic magmatism. At the northern end of the Adamello Pluton, dextral strike-slip shearing occurred at 35 to 32 Ma, coeval with the intrusion, and probably continued until *c.* 20 Ma (Stipp *et al.* 2004). Along the northern part of the SW–NE striking Giudicarie Fault, which formed a restraining bend in the dextral Periadriatic Fault, Austro-Alpine units were backthrust over South Alpine ones around 32 Ma according to Viola *et al.* (2001). This is in contrast to other interpretations (e.g. Frisch *et al.* 1998; Fig. 18.17) which assume that the Periadriatic Fault was initially straight and only offset by the northern Giudicarie Fault during the Miocene.

In the area south of the Tauern Window, displacement along steeply orientated shear zones north of the Periadriatic Fault, such as the Defereggen-Antholz-Vals Fault and the southern border of the Tauern Window, was sinistrally transtensive in Early Oligocene times, with a north-side-up vertical component (Mancktelow *et al.* 2001). This is similar to the top-to-the-SE extensional faulting along the Periadriatic (Insubric) Fault in the Central Alps during the same time interval. A change to dextrally transpressive kinematics occurred at *c.* 30 Ma, and dextral movement along the Periadriatic Fault continued at least until 13 Ma (Mancktelow *et al.* 2001), in the framework of east-directed extrusion of the Eastern Alps (see below). Although the Periadriatic Fault is a first-order tectonic boundary in this area, it was not imaged by the TRANSALP reflection seismic experiment (Lüschen *et al.* 2004).

According to Frisch *et al.* (1998) the western parts of the Eastern Alps formed a mountainous region after the Oligocene, whereas the eastern part was characterized by hilly to peneplane topography with a north-directed drainage system. On the low-lying northeastern part (Northern Calcareous Alps) sands and conglomerates, rich in quartz pebbles eroded from a greenschist-facies metamorphosed hinterland, were deposited by braided rivers (Augenstein Formation of Late Oligocene to Early Miocene age), probably passing into marine sediments of the Molasse Basin to the north. The difference in the topography is also expressed in the Oligocene sediments of the Molasse Basin, where coarse-grained material, including detritus from Periadriatic volcanoes and plutons, is frequent in the western part.

During the Miocene, continued north–south shortening of the Eastern Alps ceased to be primarily accommodated by thrust

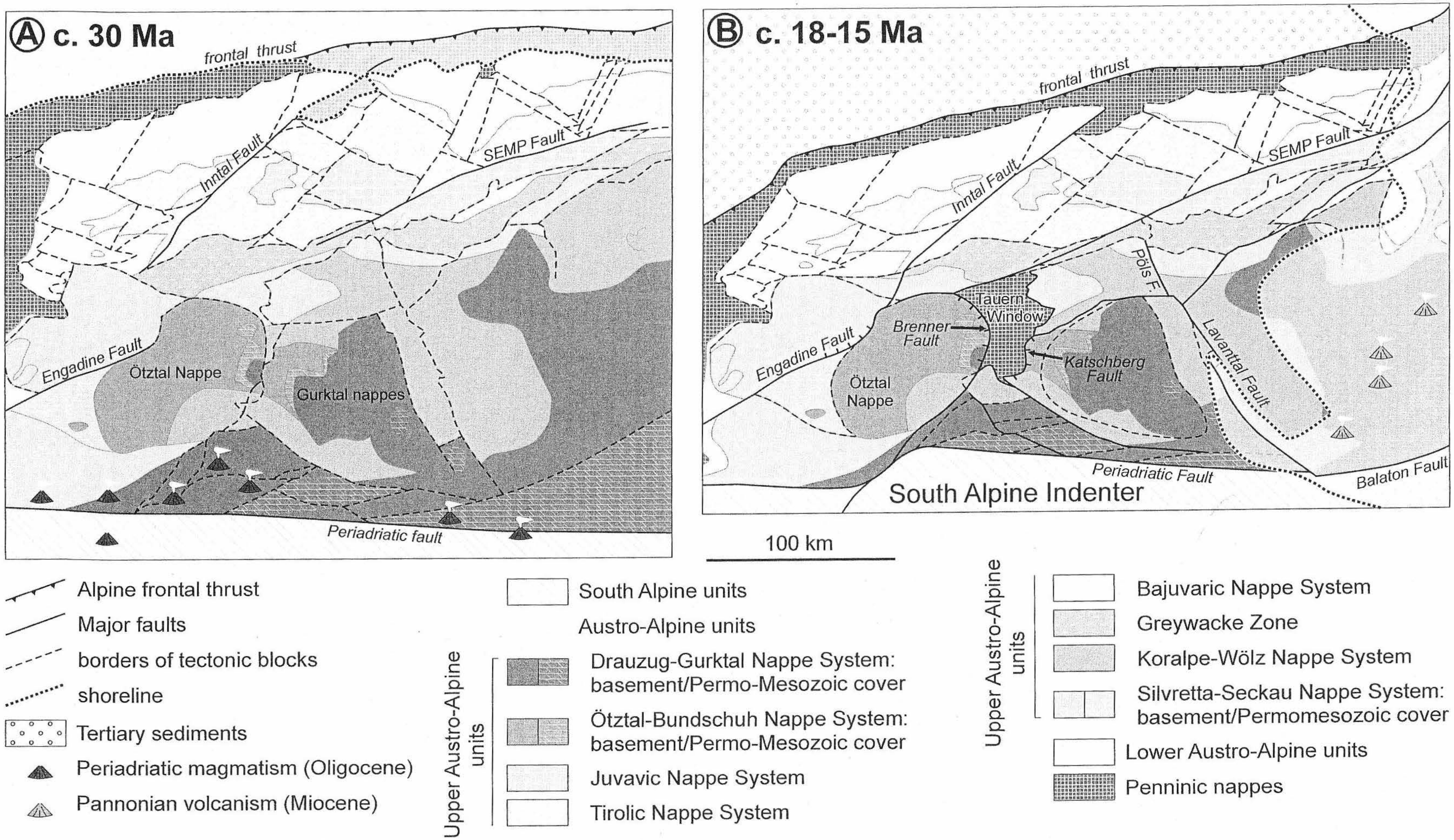



**Fig. 18.17.** Reconstructions of the Eastern Alps during (**A**) Oligocene (*c.* 30 Ma) and (**B**) Miocene times (*c.* 18 to 15 Ma). Miocene reconstruction shows beginning of east–west extension and eastward extrusion of lozenge-shaped blocks. Tauern Window begins to form by east–west extension and north–south shortening in front of the northward-advancing South Alpine indenter. After Frisch *et al.* (1998), modified.

imbrication of the European margin, but increasingly by eastward extrusion of the units located between the front of the Alps and the Periadriatic Fault (Ratschbacher *et al.* 1989; Decker & Peresson 1996; Linzer *et al.* 2002). This eastward extrusion was responsible for, or at least accentuated, the elongate shape of the Eastern Alps. It led to disintegration of the Austro-Alpine nappes above the ductilely deformed Penninic and Sub-Penninic units, cropping out in several windows (Engadine Window, Tauern Window and Rechnitz Window Group), by a system of strike-slip and normal faults. Sinistral, WSW–ENE striking faults developed in the northern part (e.g. Inntal-Salzburg-Amstetten and Salzach-Ennstal-Mariazell-Puchberg faults) whereas WNW–ESE striking dextral faults dominated in the southern part (Periadriatic, Isel, Mölltal and Lavanttal faults). Major normal faults border the western and eastern end of the Tauern Window (Brenner and Katschberg faults) and the Rechnitz Window Group (Behrmann 1988; Selverstone 1980; Genser & Neubauer 1989; Grasemann & Dunkl 1996; Fügenschuh *et al.* 1997). The new fault system had a strong influence on the topography and the drainage pattern which has been east-directed in the eastern part of the Eastern Alps since that time (Frisch *et al.* 2000). Along the faults Miocene basins developed mostly in a pull-apart regime (e.g. Tamsweg, Knittelfeld, and Vienna basins). In the case of the Vienna Basin, the evolution began with the formation of a piggyback basin on top of the Austro-Alpine nappes (Strauss *et al.* 2006). To the SE the Pannonian Basin and the related Styrian Basin formed in the interior of the Alpine-Carpathian arc. This process was associated with Miocene volcanism.

The Klagenfurt Basin formed in the Late Miocene to Pliocene as a flexural basin in front of the NW-directed overthrust of the Karawanken Mountains (Nemes *et al.* 1997). This overthrust was probably related to dextral shearing along the Periadriatic Fault (Polinski & Eisbacher 1992).

The extension and eastward extrusion of the Eastern Alps during the Miocene may be explained in different ways. Ratschbacher *et al.* (1991) suggested that it was driven by a combination of forces applied to the boundaries of wedge-shaped blocks (tectonic escape) and gravitational spreading away from a topographic high in the area of the Tauern Window (extensional collapse). This topographic high resulted from the northward motion of the South Alpine indenter bounded by the Giudicarie Fault to the NW and the Pustertal Fault to the NE, which caused strong north–south shortening, antiformal stacking and uplift of the Sub-Penninic basement nappes in the Tauern Window (Lammerer & Weger 1998). Rosenberg *et al.* (2004, 2007) emphasized the tectonic escape part and assumed extensional collapse to be less important. In addition to these 'intra-Alpine' forces, subduction retreat (roll-back) in the Eastern Carpathians, where oceanic lithosphere was probably subducted in a west-dipping subduction zone, exerted a pull force on the upper-plate lithosphere in the Pannonian Basin and Eastern Alps which contributed to the extension and extrusion (Peresson & Decker 1997).

The extension was terminated by an east–west compressional event between 9 Ma and 5.3 Ma which may have resulted from the entrance of thick European continental crust into the Eastern Carpathians subduction zone and the ensuing termination of subduction (Peresson & Decker 1997). The extrusion, however, appears to have continued until the present. GPS data suggest that eastward extrusion of the Eastern Alps is presently taking place within the framework of north–south convergence between the Adria microplate and Europe (Grenerczy *et al.* 2005; Vrabec *et al.* 2006).

*Southern Alps*

In the Lombardic-Giudicaric thrust belt, i.e. the western part of the Southern Alps, thrusting evolved in two major phases: before the Adamello intrusion and after it. Important south-vergent basement thrusts (Gallinera, Porcile, Orobic thrusts; Schönborn 1992) and related folds are truncated by the oldest part of the Adamello intrusion and are, therefore, older than 43 Ma (Brack 1981). As discussed above, these 'pre-Adamello' structures may be Late Cretaceous or Early Tertiary (Palaeocene–Eocene) in age. These structures were later passively transported southward in the hanging walls of Miocene-age thrusts. The direction of shortening in the Palaeogene (or Late Cretaceous) was north–south to NE–SW. This implies that the Alps were already a bivergent orogen in the Palaeogene (if not already in the Late Cretaceous).

Thrusting after the Adamello intrusion began in the Chattian and lasted, possibly with interruptions, until the Tortonian. Over this time, a thrust system evolved whose frontal, southern part is hidden beneath sediments of the Po Basin but is known from drilling (Pieri & Groppi 1981). According to Picotti *et al.* (1995), post-Adamello thrusting culminated in three phases: Chattian-Burdigalian, Serravallian-Tortonian, and (?)Late Tortonian. The thrusts in the subsurface were sealed by Messinian sediments. In the most external foreland area, Alpine compression came to an end after the Early Messinian. In contrast, east of the Lombardic-Giudicaric fold-and-thrust belt, Alpine deformation lasted until the Pleistocene (Fantoni *et al.* 2004).

The thrust front of the Lombardic-Giudicaric fold-and-thrust belt strikes east–west but towards the eastern termination it swings around into a SSW–NNE strike parallel to the Giudicarie Fault. This fault lies inside the belt, to the NW of the thrust front. The Giudicarie Fault approximately coincides with the eastern border of the Lombardian rift basin (of Jurassic age) against the Venetian Platform, so that the western part of the Venetian Platform was also affected by thrusting (Picotti *et al.* 1995). The southern part of the Giudicarie Fault (south of the junction with the Tonale Fault) separates uplifted Adamello granitoids, Variscan basement, and Permo-Triassic rocks of the northwestern block from mainly Mesozoic rocks of the southeastern block. It represented a sinistrally transpressive fault active after 20 to 18 Ma (Viola *et al.* 2001).

In the central and eastern Dolomites, Palaeogene, NE–SW orientated compression produced a SW-vergent thin-skinned thrust belt (Doglioni & Bosellini 1987). The thrusts typically form ramps on the northeastern slopes of the Triassic carbonate platforms and flats on top. Spectacular west-facing folds occur in the Jurassic sediments above the flats ('Gipfelfaltung': summit folding). This thrust belt is the northwestern extension of the Eocene thrust system in the Dinarides. The thrust front (Dinaric Front in Fig. 18.1) is displaced towards the west relative to its position further SE in the Dinarides; this offset results from younger, Neogene, southward thrusting along the east–west striking Bassano Thrust (see below) and similar thrusts in the eastward continuation of the Bassano Thrust.

In the Neogene, a new, thick-skinned, south- to SE-vergent thrust belt formed. The thrusts are almost at right angles to the Palaeogene ones. Two main, basement-involving, south-directed thrust zones occur: the Serravallian-Tortonian Valsugana Thrust to the north, and the Messinian to Recent Bassano and Montello thrusts to the south (Fig. 18.1; Schönborn 1999; Castellarin & Cantelli 2000). The Valsugana Thrust forms the southern limit of the Dolomites and the Bassano Thrust runs close to the northern border of the Po Plain, whereas the blind Montello Thrust is kinematically linked to the Bassano Thrust and represents the frontal element of the Neogene thrust system. Quaternary terraces have been deformed by a ramp anticline related to the Montello Thrust (Benedetti *et al.* 2000), showing that this thrust is presently active. Antithetic, north-directed thrusts also occur. The Dolomites are in the structural position of a wide, open synform between the Periadriatic Fault to the north and the Valsugana Thrust to the south. The basement crops out on both sides in the vicinity of these faults.

*Western Alps*

Subduction of oceanic units in the Western Alps probably lasted until the Early Eocene and was followed by continental subduction of the European margin leading to high- to ultrahigh-pressure metamorphism of European margin units in the Late Eocene (Dora-Maira Nappe, Gran Paradiso Nappe; Lardeaux *et al.* 2006). Ongoing continental collision in the Oligocene resulted in the thrusting of the internal zones (Penninic) of the Western Alps on the external zones (External Massifs, Dauphinois Zone) along the Penninic Frontal Thrust, coeval with rapid exhumation of the high-pressure units in the internal zone (*c.* 35 to 32 Ma; Rubatto & Hermann 2001). During the Neogene and up to the present day, the strain is partitioned into widespread extension in the internal part of the Western Alps, with the extension direction initially parallel and later perpendicular to the strike of the orogen, and shortening in the western, external part. The boundary between external and internal zones, the Penninic frontal thrust, was reactivated as an east-dipping extensional detachment fault during the Neogene (Sue & Tricart 1999, 2003). Miocene-age, orogen-parallel extension of the inner Western Alps has been interpreted as reflecting extrusion of the inner part of the Western Alps towards the south (Champagnac *et al.* 2006).

Based on palaeomagnetic evidence, the internal part of the Western Alps (Penninic units) experienced strong counterclockwise vertical-axis rotation after the Oligocene (Collombet *et al.* 2002). The amount of this rotation increases dramatically from north to south (from *c.* 25° NW of Torino to *c.* 117° in Liguria). This increase in rotation angle approximately parallels the change in the strike of the orogen around the arc of the Western Alps, suggesting that the arcuate shape of the internal zones of the Western Alps resulted from the bending of an initially straighter orogen during its post-collisional interaction with Europe. The counterclockwise rotation of the internal zones is probably closely related to the Miocene and younger southwestward thrusting of the Dauphinois Zone along the Digne Thrust (Fig. 18.5).

Other authors (Schmid & Kissling 2000) suggested that the formation of the Western Alpine arc resulted from the two-stage motion of the Adriatic microplate relative to Europe: 195 km northward motion of Adria between 50 and 35 Ma, leading to sinistral transpression in the already north–south striking Western Alps, followed by 124 km northwestward motion with 18° anticlockwise rotation between 35 Ma and the present, leading to outward-directed thrusting. This model predicts little or no rotation of the internal zones of the western Alps.

### *Present-day kinematics of the Alps*

GPS data suggest that the Adriatic microplate is presently rotating anticlockwise with respect to stable Europe around a pole in the Po Basin near Milan at an angular rate of 0.52°/Ma (Calais *et al.* 2002). This movement implies that there is an eastward-increasing north–south convergence between Adria and Europe along the northern boundary of Adria, reaching 2–3 mm/a in the eastern Alpine region, and this is taken up by contraction in the Eastern Alps and concomitant eastward extrusion of the Eastern Alps into the Pannonian Basin (Grenerczy *et al.* 2005). The southern boundary of the eastward-extruding block is the Periadriatic Fault System where dextral displacements of *c.* 1 mm/a exist on individual faults (Vrabec *et al.* 2006).

High-resolution teleseismic tomography suggests that a NE-dipping, Adriatic, lithospheric slab is present under the Eastern Alps, in contrast to the Central and Western Alps where only a European slab dipping under the Adriatic microplate is imaged (Lippitsch *et al.* 2003). This suggests that the present north–south convergence in the Eastern Alps is accommodated at a lithospheric scale by the northward subduction of Adria. Hence, the polarity of subduction appears to have changed in the Neogene, from the southward subduction of Europe to the northward subduction of Adria (Fig. 18.11). This northward subduction may be kinematically linked to the presently active southward thrusting at the front of the Southern Alps in their eastern part (Montello thrust; Fig. 18.1).

Towards the west, the rate of north–south convergence across the Alps diminishes due to the rotation of Adria. The northwestern corner of the Adriatic microplate, west of the rotation pole, even moves southward with respect to Europe (Calais *et al.* 2002), which leads to north–south compression on the south side of this corner, i.e. in the Ligurian Alps (Calais *et al.* 2002).

In the Western and Central Alps, the present-day strain field is modified by stresses resulting from topography and crustal thickness. These have a much larger effect on the local strains than the relative motions of the foreland plates. The analysis of earthquake focal mechanisms from the Western and Central Alps revealed a continuous area of extensional strain which closely follows the topographic crest line of the Alps and the axis of thickest crust. In contrast, a thrusting regime is observed locally near the border of the Alpine chain. In addition, strike-slip faults occur in both the internal and external zones (Delacou *et al.* 2004). In the Western Alps, the directions of both $\sigma_3$ in the internal, extending zone and $\sigma_1$ in the external, shortening zone are perpendicular to the orogen, which results in a radial arrangement of these principal stresses. At present, the extension direction in the inner Western Alps is orogen-perpendicular and the Western Alps are in a state of gravitational spreading, in contrast to the Eastern Alps where lateral (eastward) extrusion caused by convergence of the bounding plates is more important.

## The Western Carpathians (D.P.)

### *Geographic outline*

The Western Carpathians are the northernmost segment of the European Alpides. They form a northward-convex arc *c.* 500 km long in the west–east direction and 300 km across (Fig. 18.18). Taking into consideration the pre-Tertiary units, the Western Carpathian area includes not only the mountainous regions mostly in its northern part, but also the subsurface of the wide flat lowlands of the Pannonian Basin to the south. The western limit of the Western Carpathians next to the Eastern Alps is conventionally located in the basement of the Vienna and Danube basins. On the narrow horst dividing these two basins, the boundary is located in the so-called Carnuntum Gate between the Leitha Gebirge (Mountains) and the Malé Karpaty Mountains. North of Vienna this boundary roughly corresponds to the

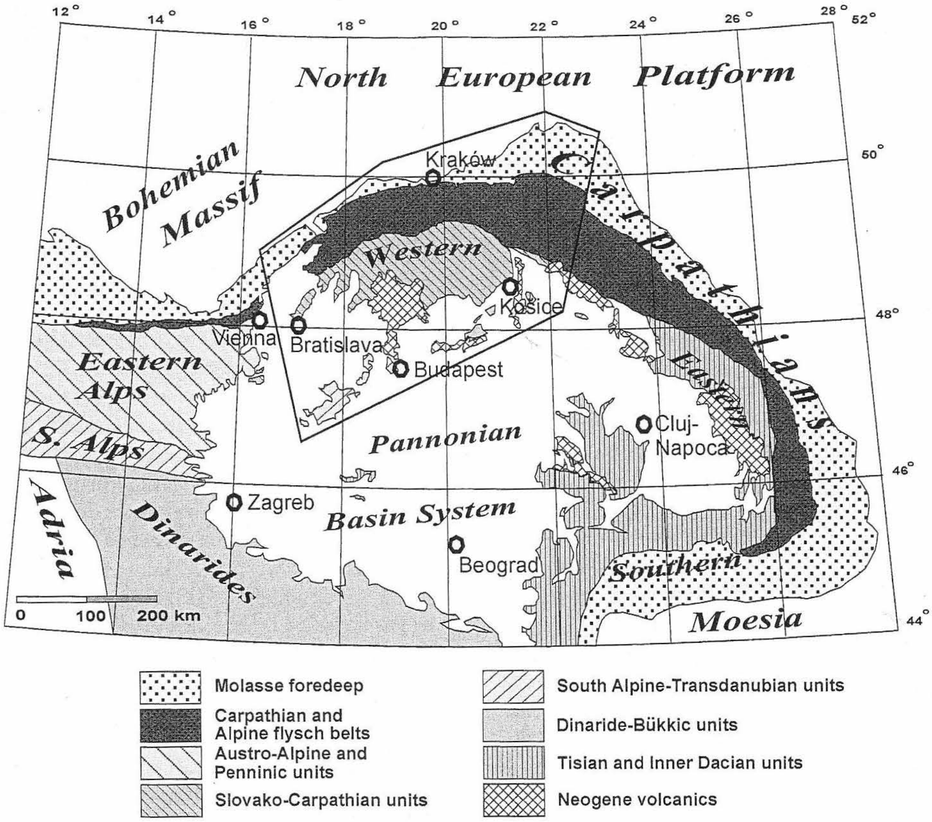



**Fig. 18.18.** Outline of the Alpine-Carpathian-Pannonian-Dinaride region. Outlined area delineates the extent of the Western Carpathians.

Rhenodanubian Flysch/Waschberg Zone division. The arcuate northern limit of the Western Carpathians follows the boundary between the most external Western Carpathian units (including the foredeep) and the foreland North European Platform composed of various pre-Alpine units and their Meso-Cenozoic sedimentary cover.

The eastern boundary of the Western against the Eastern Carpathians is the most problematic of the various boundaries, since it is not distinct either from a geographical or a geological point of view. Conventionally the boundary approximately follows the north–south trending Polish and Slovakian border with the Ukraine. However, the official geographic boundary of the Western and Eastern Carpathians is located further to the west. The extensive Rhenodanubian-Magura Superunit of the Alpine-Carpathian Flysch Belt and the Austro-Alpine–Slovako-Carpathian system wedge out in this area. In contrast, the outermost, typically Eastern Carpathian Stebnik, Sambor-Rozniatov and Borislav-Pokuty units first appear in this area and these widen to the east of the 'Przemyśl sigmoid'. However, some other large Western Carpathian units (Skole-Skiba-Tarcău, Silesian-Chornohora and Dukla units, Pieniny Klippen Belt) continue eastwards without any considerable changes. The conventional SE boundary of the pre-Tertiary Western Carpathian units follows the SW–NE trending, broad Mid-Hungarian Fault Zone (Zagreb–Zemplín Fault System). This complex Tertiary fault zone juxtaposes the southernmost Western Carpathian elements against the largely subsurface Tisza (Tisia) block in the SE part of the Pannonian Basin. However, all along this line the contact is buried beneath a thick Neogene sedimentary cover.

### *Geographic subdivision of the Western Carpathians*

The Western Carpathian area is geomorphologically divided into two principal parts: the mountainous Western Carpathians proper and the lowlands of the NW part of the Pannonian backarc basin system, i.e. the Vienna and Danube (Kisalföld) basins and the NW part of the Great Hungarian Plain (Alföld). The mountainous Western Carpathians are subdivided into the Outer Western Carpathians (mainly the Tertiary Flysch Belt) and the Inner Western Carpathians (mainly formed by pre-Tertiary complexes and Tertiary volcanic edifices), but their geographic boundary only partly corresponds to the major tectonic divide in the Western Carpathians, the Pieniny Klippen Belt. The Inner Western Carpathians are split into numerous horst blocks (so-called core mountains) separated by small Late Tertiary intramontane basins and embayments of the Pannonian Basin.

### *Tectonic subdivision*

With the exception of the unconformable Tertiary formations, the Western Carpathians can be longitudinally subdivided into three major zones, namely, the External, Central and Internal Western Carpathians (Plašienka 1999*a*; Figs 18.19, & 18.20).

The External Western Carpathians cover the territories of northeasternmost Austria, SE Moravia (Czech Republic), NW and NE Slovakia and SE Poland. They include the Carpathian foredeep, which is the eastern prolongation of the Alpine Molasse Basin, and the Carpathian Flysch Belt. Both continue along the entire Carpathian arc. The Carpathian Flysch Belt represents the Tertiary accretionary complex and consists of two

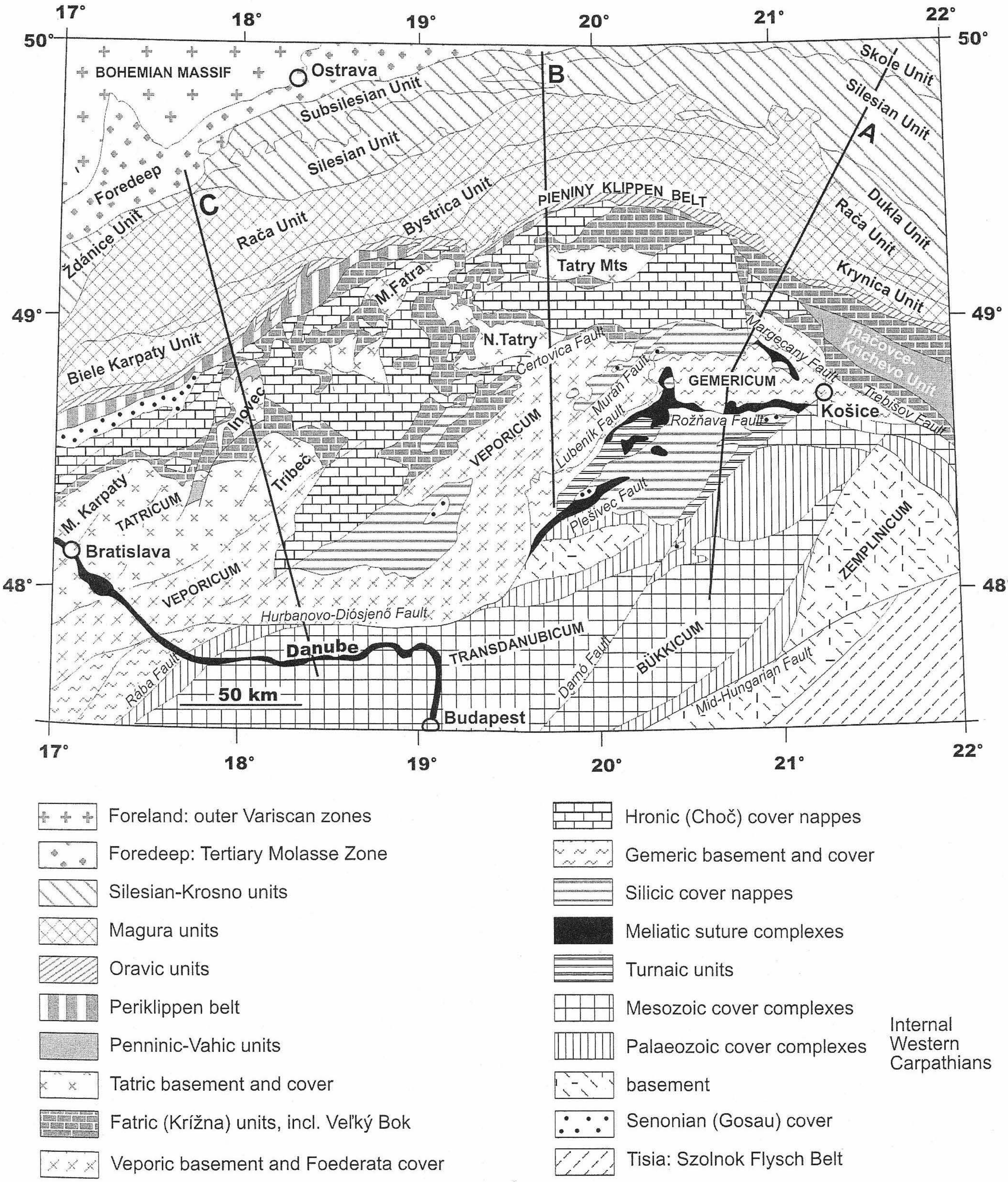



**Fig. 18.19.** Tectonic sketch of the Western Carpathians stripped off the Tertiary overstep complexes. Cross-sections A, B and C are shown in Figure 18.20.

groups of nappes (Figs 18.19 & 18.20). (1) The outer Moldavide tectonic system (Silesian-Krosno units) has no connections westwards into the Alps, but includes a substantial part of the Eastern Carpathian Flysch Belt with the exception of the Outer Dacides (Săndulescu 1988). (2) The inner Magura Superunit is considered to be a direct prolongation of the Rhenodanubian Flysch of the Eastern Alps (e.g. Schnabel 1992). It wedges out at the Western/Eastern Carpathian boundary. Palaeogeographically, the Magura Superunit represents the (northern) Penninic realm.

The Central Western Carpathians, covering much of Slovakia,

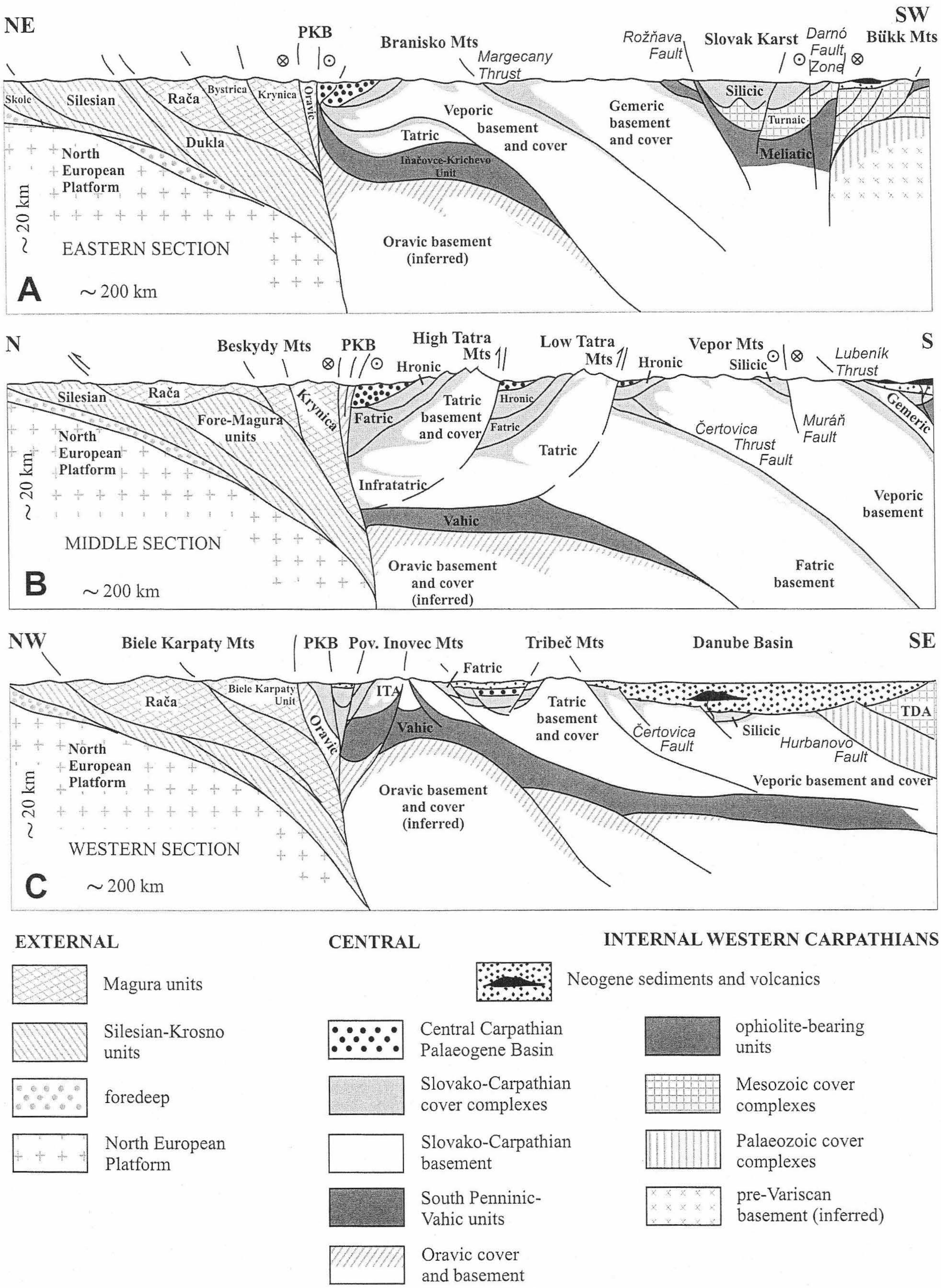



**Fig. 18.20.** Conceptual cross-sections through the Western Carpathians. Sections are vertically exaggerated; their approximate positions are shown in Figure 18.19. Abbreviations: PKB, Pieniny Klippen Belt; ITA, Infratatric Inovec nappe; TDA, Transdanubian Superunit.

include various pre-Tertiary units and the unconformable Cenozoic sedimentary and volcanic complexes. The former are distributed in several longitudinal morphotectonic belts (Plašienka *et al.* 1997). The narrow Pieniny Klippen Belt represents the boundary between the External and Central Western Carpathians. It includes units of (Middle) Penninic provenance (Oravic Superunit) and units probably derived from the Austro-Alpine–Slovako-Carpathian tectonic system (Plašienka 1995*a*). However, the Oravic Continental Fragment, from which the Oravic Superunit was derived, is only geometrically analogous to the Middle Penninic Briançonnais High of the Western Alps, since there is no connection between these units in the Eastern Alps (Trümpy 1988; Froitzheim *et al.* 1996). Areas to the south of the Pieniny Klippen Belt are composed of numerous Cretaceous nappe units which represent a continuation of the Austro-Alpine units of the central Eastern Alps and are collectively termed the Slovako-Carpathian tectonic system (Fig. 18.20). The Tatra-Fatra belt of core mountains includes the lowermost Tatric Superunit composed of (1) pre-Alpine basement and its Mesozoic sedimentary cover and (2) detached cover nappes of the Fatric and Hronic thrust systems. The former (1) is partially related to the Lower Austro-Alpine (*sensu* Tollmann 1977), while the latter (2) mainly correspond to the Upper Austro-Alpine nappes of the Northern Calcareous Alps (e.g. Häusler *et al.* 1993). The Vepor-Gemer Belt to the south corresponds to the central Eastern Alpine (Middle Austro-Alpine *sensu* Tollmann 1977) stack of basement-dominated nappes (Fig. 18.21). It is formed by the north-verging thick-skinned Veporic and Gemeric basement/cover thrust sheets and the Silicic cover nappes. The Vepor-Gemer Mountains gradually disappear towards the south under the Tertiary sediments and volcanics, making it difficult to follow the boundary between the Central and Internal Western Carpathians.

The Internal Western Carpathians are represented by the so-called Pelso Megaunit exposed mostly in isolated mountains (inselbergs) in northern Hungary (Transdanubian Range, the Bükk and surrounding mountains and the Aggtelek-Rudabánya Mountains; cf. Kovács *et al.* 2000). The supposed Central–Internal Western Carpathian boundary is represented by the oceanic complexes of the Meliata Unit, but this suture is mostly obliterated by superimposed nappe units and unconformable cover rocks. The Internal Western Carpathians mainly comprise unmetamorphosed Palaeozoic and Mesozoic complexes that form a south-directed fold-and-thrust belt (Figs 18.19, 18.20 & 18.21). In terms of palaeogeography, the Pelso Megaunit is closely related to the southern Austro-Alpine or to the South Alpine facies realm (Transdanubian Range), or even to that of the Dinarides (Bükk Mountains).

In plate tectonic terms, the External Western Carpathians correspond to a Tertiary accretionary complex related to the southward subduction of the North Penninic (Magura), and possibly also the Moldavide, oceanic realms. The Pieniny Klippen Belt forms a narrow steep transpressional zone between the accretionary wedge and the Central Western Carpathians representing the backstop. It is also the boundary between the Penninic and Austro-Alpine-related units in the Western Carpathians. Accordingly, it is a fossil plate boundary (i.e. suture), although surface evidence for ophiolite complexes is lacking. The Central and Internal Western Carpathians represent the stacks of crustal-scale units within the late Mesozoic collisional pro- and retro-wedge, respectively, shaped by the latest Jurassic to mid-Cretaceous subduction–collision processes related to the closure of the Triassic–Jurassic Meliata Ocean. The ophiolite-bearing Meliata Suture is approximately located in a central position between the pro- and the retro-wedge.

### *Palaeogeographic subdivision*

From the point of view of Mesozoic–Cenozoic palaeogeography, two principal evolutionary periods can be distinguished in the Western Carpathians. During the first (Triassic–Jurassic), units of the External and Central Western Carpathians were located on the northern, rifted, passive European margin of the Meliata Ocean. In contrast, the Internal Western Carpathian elements formed the southern (in present co-ordinates) passive (Triassic) and later active (Jurassic) margin of the ocean, corresponding to the NE part of Apulia-Adria. The Central and Internal Western Carpathians were welded together by the time of the Jurassic–Cretaceous boundary following closure of the Meliata Ocean. From this time onward, the Western Carpathian Cretaceous palaeogeography generally corresponds to that of the Alps, including the following zones from north to south: European continental margin, largely overridden by the Western Carpathians units during the Tertiary; Moldavide basins (partly oceanic?) in the north to NE only, and widening eastwards; elongated continental fragment (the Silesian Ridge); the North Penninic–Magura Oceanic Basin; rifted continental fragment in a Middle Penninic position (the Oravic or Czorsztyn Ridge); the South Penninic–Vahic Oceanic Zone; and the broad and dissected Adria-related continental margin including the Slovako-Carpathian (Austro-Alpine) and Pelso (Upper Austro-Alpine–South Alpine–Dinaridic) units.

### *Outline of the palaeogeographic and palaeotectonic evolution*

The Alpine edifice of the Western Carpathians was constructed on the remnants of the Variscan orogenic belt of Central Europe. The Variscan Orogeny culminated with the late Early to early Late Carboniferous collision processes (Putiš 1992; Plašienka *et al.* 1997; see also Kroner *et al.* 2008; McCann *et al.* 2008*a*). Post-Variscan unconformable successions include marine Pennsylvanian and continental Permian clastic units. During the Permian, several narrow rift basins formed, which were filled with continental red-beds (Vozárová & Vozár 1988; Vozárová 1996). Some of these were accompanied by alkaline to calc-alkaline, acid to intermediate (locally also basic) volcanism (Dostal *et al.* 2003). Permian to Lower Triassic granitoid intrusions include late-orogenic S-type granites in the Gemeric Superunit and several small, post-orogenic A-type intrusions in the southern zones of the Western Carpathians (e.g. Broska & Uher 2001). In the Early Triassic, the region had become a peneplain and most of the Western Carpathians area was covered by mature continental to beach siliciclastics (Mišík & Jablonský 2000), subsequently overlain by lagoonal or sabkha deposits. However, the site of the later Meliata Rift was already marked by the presence of much thicker Scythian-age deposits (Hips 2001).

Gradual subsidence of the European shelf during the Anisian is reflected by the deposition of widespread carbonate ramp and platform deposits, with intervening narrow intrashelf basins (e.g. Michalík 1994*a*). Early Mesozoic Tethyan rifting culminated in the opening of the Meliata Ocean in the early Middle Triassic (as indicated by the Pelsonian breakup unconformity in adjacent areas; cf. Kozur 1991). The Meliata Ocean has been interpreted as a backarc basin related to the northward subduction of Palaeotethys (Stampfli *et al.* 1998). This ocean subsequently separated the stable European shelf in the north from the mobile Adria-related continental fragments in the SW (Figs 18.22 & 18.23). The broad northern shelf exhibited only restricted

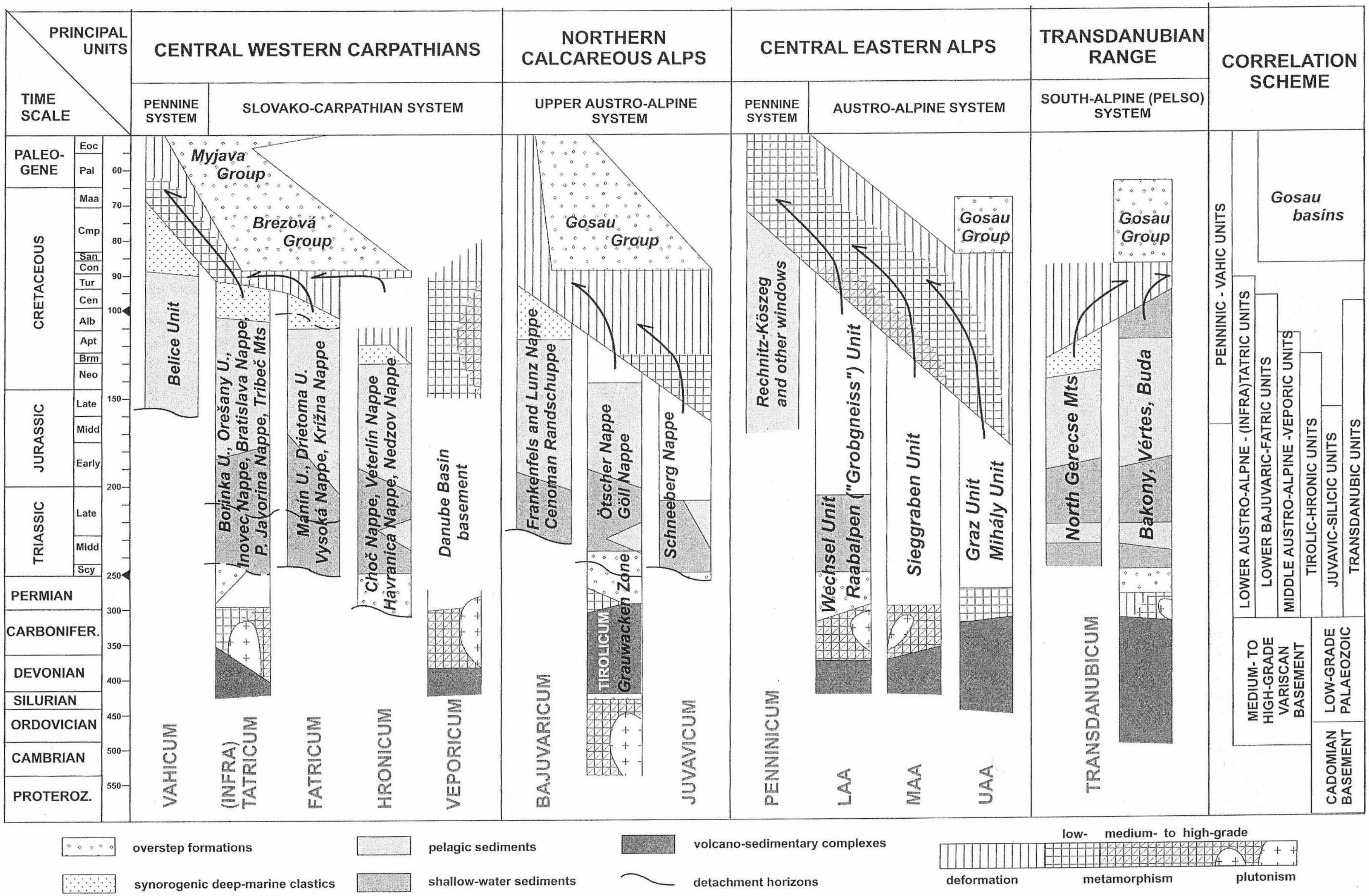



**Fig. 18.21.** Synoptic correlation table of pre-Tertiary units at the Alpine–Carpathian–Pannonian junction area (east Austria, SW Slovakia, NW Hungary), after Plašienka (1999*b*), modified. Nomenclature of Alpine units is mainly according to Tollmann (1977) and Wessely (1992).

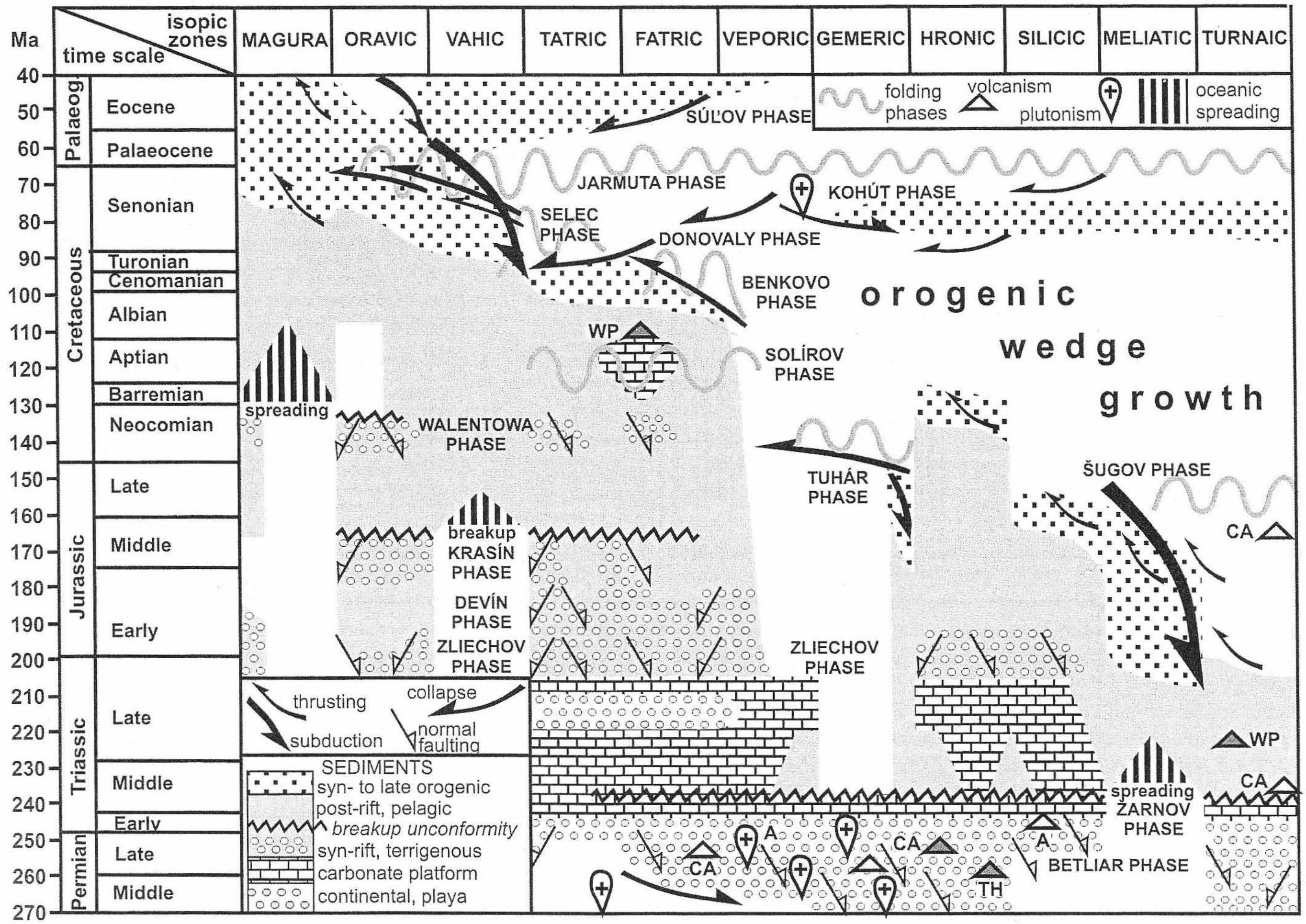



**Fig. 18.22.** Synoptic overview of the principal early Alpine orogenic processes and regional tectonic phases defined in the Central Western Carpathians and adjacent zones.

subsidence during the Triassic. Epicontinental successions similar to those of the German Basin were deposited, e.g. the Upper Triassic Carpathian Keuper Formation. Along the edge of the northern shelf, however, extensive, reef-cored Middle–Upper Triassic carbonate platforms developed. These reflect the thermal subsidence of the Meliata Ocean flanks. In contrast, the southern Meliata margin records Tethyan-type, mostly deep-water pelagic sedimentation, since deposition did not keep pace with subsidence (Haas *et al.* 1995; Csontos 2000). In the axial zones of the Meliata Ocean, sedimentation below the CCD predominated. Ladinian and Upper Triassic radiolarites today occur as blocks within the Jurassic olistostromes and mélanges, together with dismembered ophiolite fragments (Mock *et al.* 1998).

From the Triassic–Jurassic boundary onwards, important palaeogeographic changes occurred. The northern shelf experienced several strong rifting phases that led to the disintegration of the Triassic carbonate platform and ultimately resulted in the opening of new, Penninic-related oceanic basins (Plašienka 1995*a*, 2003*a*). The Slovako-Carpathian domain was separated from the North European Platform by the, probably Bathonian, breakup of the South Penninic–Vahic ocean, and was dissected into several longitudinal, wide, subsiding basins, floored by extended lithosphere and filled with deep-water, predominantly calcareous pelagic sediments (Figs 18.22 & 18.23b). These Jurassic–Lower Cretaceous basins were separated by narrow, occasionally emergent ridges with shallower and/or condensed sedimentation (e.g. Wieczorek 2001).

The southward subduction of the Meliata Ocean probably commenced during the Early Jurassic and its closure occurred in the latest Jurassic (Kozur 1991; Kovács 1992; Haas *et al.* 1995; Csontos 1999; Kovács *et al.* 2000). Subsequent shortening migrated from the collision zone to both sides (Figs 18.22 & 18.23a,b). During the Early Cretaceous, the Internal Western Carpathian retro-wedge (Pelso Megaunit) was formed by generally southward thrusting and obduction of a small Upper Jurassic backarc oceanic basin, as indicated by the presence of the Szarvaskő ophiolites of the Bükk Mountains (Csontos 1999, 2000) and by backthrusting and formation of a retro-arc, ophiolite-detritus-bearing deep-marine clastic basin in the Transdanubian Range (Császár & Árgyelán 1994). Subsequently, the Pelso units were unconformably overlain by the Senonian (Gosau Group) and by Eocene to Lower Miocene epicontinental deposits of the North Hungarian (Buda) Basin.

The Central Western Carpathian pro-wedge grew significantly during the late Early and early Late Cretaceous. Considerable shortening of the Central Western Carpathian crust, attenuated by previous Jurassic rifting, took place and the principal thick- and thin-skinned nappe sheets were thrust northward (e.g. Plašienka *et al.* 1997). Mid-Cretaceous deep-marine clastic basins developed in front of the advancing basement thrust wedges. The

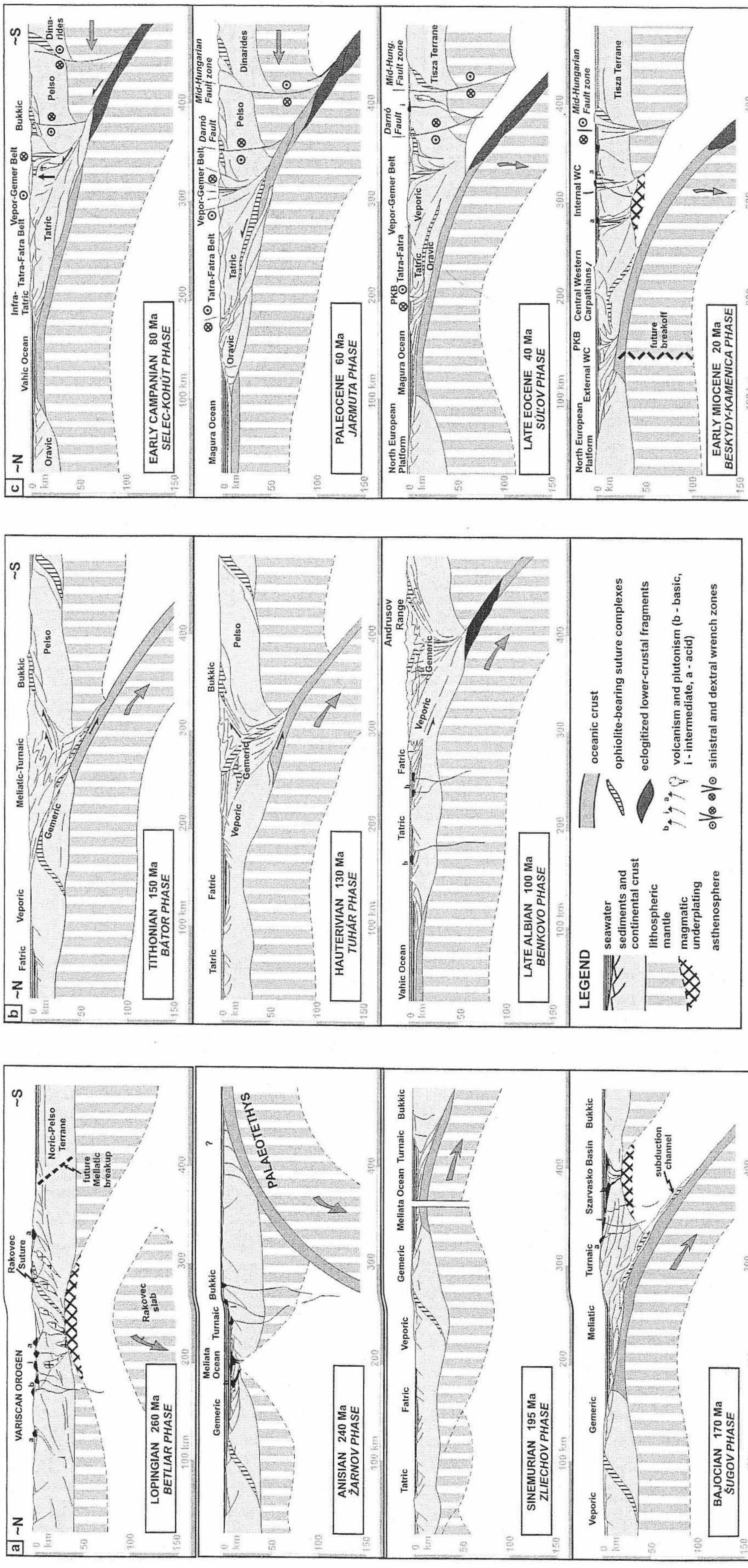



**Fig. 18.23.** Tentative palaeogeodynamic sections of the Western Carpathian orogen during **(a)** Late Permian to Middle Jurassic; **(b)** Late Jurassic to mid-Cretaceous; **(c)** Late Cretaceous to Early Miocene.

Cretaceous nappe stack in the Central Western Carpathians was completed by the Turonian, and is unconformably overlain by the Senonian deposits, although these are only locally preserved (Wagreich & Marschalko 1995). The thickened crust collapsed in the southern zones of the Central Western Carpathians in the Late Cretaceous, when the Veporic metamorphic core complex was exhumed by an orogen-parallel, extensional unroofing (Janák *et al.* 2001).

Deep-water clastic deposits, several thousand metres thick, of the Central Carpathian Palaeogene Basin form the main seal in the Central Western Carpathians. These indicate the rapid collapse of the northern part of the Central Western Carpathian following the eastward-migrating Eocene transgression (Wagreich 1995; Soták *et al.* 2001; Kázmér *et al.* 2003). The Central Western Carpathian Palaeogene deposits are generally undeformed and the Central–Internal Western Carpathian orogenic wedge functioned as a relatively rigid buttress for the developing accretionary wedge of the External Western Carpathians during the late Palaeogene and Miocene (Figs 18.22 & 18.23c).

North of the Slovako-Carpathian domain, Penninic rifting occurred in three main phases. The first Early Jurassic phase affected broad areas, but did not yet lead to continental breakup. The Middle Jurassic localized asymmetric rifting led to opening of the South Penninic–Vahic ocean basin, while the onset of oceanic crust production in the North Penninic–Magura Ocean probably began as late as the earliest Cretaceous (Plašienka 2003*a*). There is, however, no direct evidence for the oceanic nature of these basins, since no ophiolites are found. Southward-directed subduction of the Vahic oceanic crust commenced in Senonian times (Plašienka 1995*a*, *b*). The Oravic continental fragment began to collide with the Slovako-Carpathian orogenic wedge at the Cretaceous–Palaeogene boundary (Figs 18.22 & 18.23c). Only detached Vahic, Oravic and Magura sediments were frontally accreted to the orogenic wedge, while the entire Penninic lithosphere, either oceanic or continental, was subducted. Subduction of the Magura Ocean probably began during the Eocene and lasted until the earliest Miocene. The eastward-migrating subduction of the basement of the Moldavian units, followed by underthrusting of the European margin below the External Western Carpathian accretionary wedge, commenced in the Late Oligocene and ended by the Early Miocene in the west and by the Middle Miocene further to the east (Nemčok *et al.* 1998, 2000).

The Middle to Late Miocene retreat (roll-back) of the subduction zone in the External Western Carpathians had crucial consequences for the evolution of the Central and Internal Western Carpathians in the upper plate position (Figs 18.22 & 18.23c). From the Middle Miocene onwards, the Pannonian Basin System formed in a backarc position (e.g. Royden *et al.* 1982; Csontos *et al.* 1992; Horváth 1993; Csontos 1995; Kováč *et al.* 1998; Bada & Horváth 2001). Initially, small pull-apart basins opened, followed by widespread rifting and thermal subsidence that unified numerous small depocentres into the extensive Pannonian Basin. In the Western Carpathians, this backarc rifting was also associated with the uplift of small horsts, forming mountains from the Late Miocene up until the present (Kováč *et al.* 1994). The peaks of the High Tatra Mountains are the highest in the entire Carpathians, though the highest point (Gerlachovský štít) is only 2654 m above sea level. Marine conditions in the Pannonian Basin System were replaced by brackish and then by freshwater conditions during the Late Miocene to Pliocene, as the connections with the Mediterranean and Black seas were closed (Kováč *et al.* 1993; Kováč 2000). Lithospheric stretching and asthenospheric upwelling were associated with voluminous calc-alkaline andesitic volcanism, which culminated during the Middle and Late Miocene and was terminated by minor basanitic extrusions during the Pliocene and Quaternary (e.g. Konečný *et al.* 2002).

### *Carpathian foredeep*

The Carpathian foredeep is the eastward lateral prolongation of the East Alpine Molasse Basin. It is the outermost element of the Carpathian Orogen which developed at the orogenic front of the Carpathian arc from the area north of Vienna to the Iron Gate of the Danube in Romania i.e. near Orşova (Fig. 18.18). It is *c.* 1500 km long with a curvature of nearly 270°. In the Western Carpathian sector, the foredeep is filled with marine to brackish sediments and with some continental sediments of Lower to Middle Miocene age. It represents the peripheral flexural foreland basin formed in response to the loading of the lower, subducted plate by the advancing nappes of the accretionary wedge (Carpathian Flysch Belt). The main phases of the foredeep subsidence and depocentre migration were closely related to the thrusting phases in the Carpathian Flysch Belt (Meulenkamp *et al.* 1996; Oszczypko 1998). Due to the oblique, eastward-migrating soft collision of the frontal part of the West Carpathian orogenic system with the North European Platform, the foredeep exhibits a conspicuous eastward younging. Forming during the terminal stages of the orogeny, the foredeep reflects the gradual frontal and lateral cessation of convergence between the stable platform foreland and the mobile Alpine-Carpathian Orogen.

The sedimentary infill of the Carpathian foredeep forms a characteristic clastic wedge thinning towards the foreland, and is subdivided into: (1) the lower and outer autochthonous part directly overlying various pre-Tertiary basement and cover rock complexes of the Bohemian Massif and Malopolska Block of the North European Platform; and (2) the parautochthonous inner and upper part, which is partly incorporated into the Outer Carpathian accretionary wedge (Carpathian Flysch Belt). Three lateral sectors can be identified in the Western Carpathian foredeep, differing both in terms of the character and the age of the sedimentary filling.

The westernmost, **Moravian sector,** extending from NE Austria to south-central Moravia (Czech Republic) is <20 km wide and comprises sediments predominantly of Lower Miocene age (e.g. Brzobohatý & Cicha 1993). However, two deep, buried, NW–SE orientated canyons (Nesvačilka, Vranovice), filled with Palaeocene–Eocene clastic sediments, have been encountered by drilling below the foredeep deposits and the nappe front of the Carpathian Flysch Belt in southern Moravia (Pícha 1979). These presumably represent the channels that fed the Outer Carpathian deep-marine clastic basins.

The oldest (Egerian–Eggenburgian, i.e. Aquitanian to Early Burdigalian) deposits consist of reworked material from the weathered crystalline rocks of the underlying Bohemian Massif. The subsequent Eggenburgian marine transgression resulted in the deposition of thick clastic units. During the Ottnangian (Middle Burdigalian), older sediments were partly eroded and brackish, lagoonal and freshwater sediments developed locally. A new, Carpathian (Late Burdigalian) sedimentary cycle began with the deposition of basal clastics, followed by up to 1200 m of deep-marine calcareous clays, turbiditic siltstones, and sandstones. During the Carpathian, the foredeep width was laterally reduced due to the advance of the Flysch Belt nappes. Some of the older sediments were buried or detached to create transitional parautochthonous elements at the boundary between the foredeep and the accretionary wedge (the Pouzdřany Unit in southern Moravia). The final marine transgression, which extended far into the foreland (Bohemian Massif) along tectonically predisposed zones, took place in the Early Badenian (Langhian). Hundreds of metres of shallowing-upward, terrigenous clastics and calcareous clays were deposited in the axial zones of the foredeep. From the Middle Badenian (Serravalian) onwards, the foredeep depocentres migrated to the NE. The youngest, uppermost Miocene and Pliocene continental clastic sediments are mainly found in wide grabens orientated perpendicular to the foredeep axis, thus reflecting their post-convergence formation.

The middle, **Silesian sector** of the foredeep in northern Moravia and southern Poland (west of Kraków) is *c.* 30–40 km wide and is filled with deposits of Badenian to Sarmatian age. Lower Miocene sediments are known only from boreholes that extend through the Carpathian Flysch Belt nappes. The Middle to Late Badenian (Serravalian) foredeep is filled with characteristic evaporitic deposits (e.g. famous salt mines at Wieliczka near Kraków and sulphur pits at Tarnobrzeg). The decline in subsidence (or even uplift) was associated with a phase of decreased Carpathian convergence (Oszczypko 1998). The latest Badenian and Sarmatian are represented by pelitic and psammitic sediments reflecting decreasing salinity.

The eastern, **Galician sector** of the Western Carpathian foredeep between the Kraków and Przemyśl at the Polish–Ukrainian border is younger still, and filled with thick (up to 3500 m) units of marine sediments of Late Badenian to Sarmatian (Serravalian to Early Tortonian) age. The foredeep reaches its greatest width in this area (>80 km at the surface). Its inner part, composed of Lower to Middle Miocene units, mainly terrestrial sediments >1.0 km thick, is overridden by the Carpathian Flysch Belt nappes. Folded Badenian- and Sarmatian-age sediments are also partially incorporated into the outermost parts of the Carpathian Flysch Belt (narrow Zgłobice Unit in front of the Silesian and Skole nappes).

## *Carpathian Flysch Belt*

The Carpathian Flysch Belt extends from the area north of Vienna up to the Vrancea region at the boundary between the Eastern and Southern Carpathians (Fig. 18.18). It represents the frontal Tertiary accretionary wedge of the Carpathian Orogen and forms a crescent-shaped arc around the various pre-Tertiary units, blocks or terranes in the inner part of the orogen, which have complex, and sometimes ambiguous, mutual relationships (e.g. Csontos *et al.* 1992; Kováč *et al.* 1994; Csontos & Vörös 2004). Its inner arc is *c.* 1000 km and the outer arc >1200 km long. The width of the belt varies between 60 and 120 km. It is composed exclusively of Jurassic to Miocene sediments that were scraped off the subducted, presumably at least partly oceanic, basement of the Carpathian embayment. However, no ophiolite remnants are found at the surface in the Carpathian Flysch Belt.

### *Silesian-Krosno units (Moldavides)*

The Carpathian Flysch Belt consists of numerous elongate tectonic units though none of these is continuous all along the belt. Two first-order, large-scale tectonic systems can be discerned. The outer one, the Moldavian system (Silesian-Krosno units), comprises most of the Eastern Carpathian Flysch Belt, but narrows westward and is progressively replaced by the more internal Magura Superunit of the (North) Penninic tectonic system (Fig. 18.19). Hence, in the External Western Carpathians both Moldavian and Penninic units are present, the former first appearing at the front of the Carpathian Orogen in the hinge area between the Eastern Alps and Western Carpathians, the latter gradually wedging out in the area of the Western–Eastern Carpathian boundary.

The Silesian-Krosno (or Krosno-menilite) units form a thin-skinned fold-and-thrust belt that overlies the foredeep sediments and autochthonous cover and basement of the North European Platform (Tomek 1993; Tomek & Hall 1993; see Fig. 18.20). The outermost **Pouzdřany Unit** appears as a narrow belt of deformed Oligocene to Lower Miocene clays, marls and silts at the boundary between the foredeep and the Carpathian Flysch Belt proper in southern Moravia. The **Ždánice Unit** in south-central Moravia overrides the Pouzdřany Unit and includes sediments of a much wider stratigraphic range. It extends to the SW into Austria where it is known as the **Waschberg Zone**. Upper Jurassic platform limestones form isolated tectonic slices at the nappe front. The Senonian- to Eocene-age succession is only a few hundred metres thick and contains mainly hemipelagic claystones with occasional bodies of sandstones and conglomerates. The Oligocene Menilite Formation, which is typical of all of the Silesian-Krosno units, is composed of deep-water, anoxic bituminous shales with characteristic fish remains and black cherts ('menilites'), overlain by >1200 m of Egerian (Chattian–Aquitanian) synorogenic, turbiditic sediments. The Eggenburgian and Carpathian (Burdigalian) sediments were deposited in localized piggyback basins. The Ždánice Unit extends northeastwards into the **Sub-Silesian Unit** of Silesia which comprises a Senonian to Oligocene, hemipelagic and partly turbiditic succession.

The **Skole Unit** in SE Poland is detached along the mid-Cretaceous black shale horizon and ranges stratigraphically up to the Lower Miocene (Oszczypko 2004). It contains Upper Cretaceous siliciclastic and calcareous turbidites, Palaeocene–Eocene shales and distal turbidites, Lower Oligocene 'Globigerina' marls and black shales and the Egerian Krosno Flysch Formation. The Skole Unit is the areally extensive and internally tightly imbricated thrust sheet that extends from the NE part of the Western Carpathians for more than 700 km up to the Eastern/Southern Carpathian hinge. Its prolongation is known as the Skiba Unit in the Ukraine and the Tarcău Unit in Romania.

The **Silesian Unit** is one of the largest in the Carpathian Flysch Belt. It extends from Silesia to the Ukrainian Eastern Carpathians north of the Marmarosh Massif (here termed the Chornohora Unit) and comprises a continuous stratigraphic succession up to 6000 m thick from the Late Jurassic through to the Egerian. Maximum thicknesses are typical in the dominant basinal ***Godula succession*** while other successions, representing slope and ridge environments, are much thinner and occur only in Silesia. Pre-Tertiary beds crop out mainly in the west in Silesia, while Palaeogene deposits are widespread further to the east. The oldest sediments are Oxfordian to Berriasian hemipelagic marlstones and allodapic detrital limestones, overlain by Valanginian–Aptian, proximal to distal, calcareous deep-marine clastics with interdigitated submarine lava flows and dykes (known as 'teshenites') which are rift-related alkaline basalts. Mid-Cretaceous sediments are partly anoxic hemipelagic shales, passing upwards into thick Cenomanian–Senonian deep-marine clastics. The Palaeocene and Eocene are dominated by deep-marine shales and distal turbidites. The Lower Oligocene Menilite Formation is overlain by the turbiditic Krosno Formation of Egerian age which attains a thickness of >1000 m in SE Poland. The latter was deposited immediately prior to the onset of thrusting and folding. The ridge facies (***Baška succession***) is represented mainly by blocks and slices of Tithonian platform limestones (Stramberg Formation) which are found in the frontal part of the Silesian nappe.

The **Dukla Unit** first appears on the surface between the Silesian and Magura units near the Polish–Slovakian border and then widens to the SE from NE Slovakia to the Ukrainian Eastern Carpathians. The Dukla Unit is composed of mid-Cretaceous shales, Senonian to Eocene turbidites and the Lower Oligocene Menilite Formation which passes upwards into distal turbidites (e.g. Oszczypko 2004). The Dukla and related units (e.g. Grybów, Obidowa-Slopnice) are extensive in the subsurface below the frontal Magura nappes in Poland and NE Slovakia, as recorded by deep wells and tectonic windows (e.g. Mszana Dolna, Smilno). Balanced cross-sections reveal that the Dukla Unit and its equivalents (sometimes referred to as the Fore-Magura group of nappes; Oszczypko 2004) are volumetrically the largest unit of the entire External Western Carpathians (Roca *et al.* 1995; Nemčok *et al.* 2000). The **Fore-Magura Unit** *sensu stricto*. in Silesia is the western counterpart of the Dukla Unit. It is restricted to narrow tectonic slices in front of the Magura Superunit.

*Magura Superunit*

The Magura Superunit, which overthrusts the Silesian-Krosno units, emerges from the pre-Neogene units of the Vienna Basin in southern Moravia and western Slovakia and continues in a belt up to 50 km wide as far as the Western–Eastern Carpathian boundary in SW Ukraine, where it wedges out eastwards (Fig. 18.19). Its southern boundary follows the northward-convex arcuate Pieniny Klippen Belt for c. 400 km. Four principal units are generally recognized within the Magura Superunit (Figs 18.19 & 18.20): from north to south the Rača, Bystrica and Krynica subunits can be distinguished. These are broadly related, whereas the innermost and westernmost Biele (Bílé) Karpaty Unit at the Slovakian–Moravian border has some special features and is sometimes considered to be an independent unit. The various Magura units comprise detached sediments of Late Cretaceous to Early Oligocene age dominated by deep-marine clastic lithologies (e.g. Oszczypko 1992, 2004).

The outer **Rača Subunit** (including the Siary Subunit recognised in Poland; Oszczypko 2004) is the areally most extensive of the Magura subunits. Its sole thrust is moderately south-dipping and overrides the Silesian and Fore-Magura units in the west and the Dukla and related units in the north and east. Fragments of Lower Cretaceous basinal sediments occur sporadically in front of the Magura Superunit. The Rača Subunit is composed of mid-Cretaceous non-calcareous shales, Senonian thick-bedded turbiditic sandstones, Lower Eocene deep-water shales (Beloveža Formation), and Middle–Upper Eocene and locally Oligocene synorogenic turbidites (Magura and Malcov formations). Rare Lower Miocene sediments also occur in remnants of small piggyback basins.

The Rača Subunit is deformed into numerous imbricates and tight north-vergent asymmetric macrofolds formed by fault propagation. Fold axes are subhorizontal and parallel, hence the regional strike directions are quite uniform. Based on data from tectonic windows of the Fore-Magura units, borehole data and section balancing, it is clear that the Magura basal overthrust is folded due to the duplexing of underlying units.

The **Bystrica Subunit** comprises a narrow (5–10 km), but long and continuous band of Senonian- to Oligocene-age sediments similar to the Rača Subunit. It represents the axial zone of the Magura Basin with Eocene deep-marine clastics attaining a thickness of *c.* 1000 m.

The inner **Krynica Subunit** (also called the Oravská Magura Unit in NW Slovakia) is restricted to the central segment of the Western Carpathian Flysch Belt. The Eocene thick-bedded sandstones (Magura Formation) of this subunit form the morphologically most prominent part of the Magura Superunit. The Lower Oligocene Menilite Formation is overlain by the turbiditic Malcov Formation deposited in a small ponded basin. Lower Miocene deposits occur in restricted piggyback basins. In some places, mainly in Poland, narrow slices of basinal Jurassic and Cretaceous sediments known as the **Grajcarek Unit** occur along the boundary between the Krynica Subunit and the Pieniny Klippen Belt. These are considered to represent the innermost part of the Magura Basin incorporated into the Pieniny Klippen Belt structure (Birkenmajer 1986; Oszczypko 2004). Partial synonyms for the Grajcarek Unit include the Hulina Unit (Sikora 1974), and the Fodorka succession of western Slovakia.

Lithological complexes within the Bystrica and Krynica subunits are often steeply dipping (or even overturned) and tightly folded with steep axial planes, and affected by transpressional deformation. The southernmost elements are locally back-thrust onto the Pieniny Klippen Belt. These features suggest that the southernmost part of the accretionary wedge of the Carpathian Flysch Belt plunges to great depth at the contact of the colliding North European Platform and Carpathian Orogen (Fig. 18.20).

The **Biele Karpaty Unit** (Bílé Karpaty in the Czech literature) comprises sediments of Cretaceous to earliest Eocene age (e.g. Švábenická *et al.* 1997). Lower Cretaceous deep-water siliceous shales pass into distal and then, during the latest Cretaceous and Palaeocene, into proximal calcareous turbidites similar to the Jarmuta Formation of the Pieniny Klippen Belt. This succession is dominated by carbonate terrigenous material derived from the areas of the Pieniny Klippen Belt and the Central Western Carpathians to the south, unlike other units of the Carpathian Flysch Belt which contain predominantly siliciclastic material derived from more distant sources. In contrast to the underlying and more external, imbricated Magura thrust stack, the Biele Karpaty Unit forms a flat-lying and comparatively less deformed, thin thrust sheet that was most probably emplaced out of sequence.

The structural styles of the Outer Carpathian nappes include piggyback thrusting, the formation of imbricated duplexes, fault-bend and fault-propagation folding, ramp and flat geometry in frontal parts, and oblique out-of-sequence thrusting of the Magura Superunit (Nemčok *et al.* 2000). Internal shortening of the Magura Superunit was estimated to be 50%, while that of the Silesian-Krosno units was 60%. The overall shortening attains at least *c.* 160–180 km in the Polish part (Nemčok *et al.* 2000; Roca *et al.* 1995), resulting in a shortening rate of 1.1–1.4 cm/a during the Oligocene–Sarmatian. According to Behrmann *et al.* (2000), the cumulative shortening of the Carpathian Flysch Belt in the eastern West Carpathians was 260 km, resulting in a convergence rate of 2.2 cm/a. The amount of shortening in the Carpathian Flysch Belt closely matches estimates of the total extension in the Pannonian Basin.

## *Pieniny Klippen Belt*

The Pieniny Klippen Belt is probably the most famous, and doubtlessly also the most complicated and enigmatic, of the Western Carpathian units. It typically follows the boundary between the external and central Carpathian zones for *c.* 500 km (from the eastern margin of the Vienna Basin to Novoselica in SW Ukraine), or for nearly 700 km if its lateral counterparts, with similar positions and structures, are taken into account (i.e.

the St. Veit Klippenzone in the Vienna Forest and the Poiana Botizei Zone in northern Romania). In contrast, the width of the belt is just several kilometres, and it may be as narrow as several tens of metres in some areas (Fig. 18.19).

Several deep boreholes and seismic lines indicate a nearly vertical subsurface dip for the Pieniny Klippen Belt down to depths of at least 5 km (e.g. Birkenmajer 1986; Tomek 1993; Vozár *et al.* 1998; Bielik *et al.* 2004; Hrušecký *et al.* 2006). The surface trend of the Pieniny Klippen Belt largely coincides with the course of the 'Peri-Pieniny Lineament' at greater depths (Máška & Zoubek in Buday *et al.* 1960); this is a deep crustal-scale fault associated with several geophysical anomalies (gravity minimum, change in polarity of the Wiese vectors, Moho steps). This fault has been assumed to coincide with the boundary between the subducted margin of the North European Platform and the Carpathian orogenic stack (Fig. 18.20). However, it is rather improbable that the units of the Pieniny Klippen Belt extend along this fault into the lower crust. It is more likely that they become inclined to the south at a depth of 5 to 10 km and extend southwards below the outer part of the Central Western Carpathians (Bielik *et al.* 2004). In the middle to lower crust, the Peri-Pieniny Lineament may represent the contact between the North European Platform and the original basement of most of the Pieniny Klippen Belt units (the underthrust Oravic continental fragment).

The present surface structure of the Pieniny Klippen Belt can be best explained as resulting from extensive and complex transpressional to transtensional movements (e.g. Ratschbacher *et al.* 1993; Nemčok & Nemčok 1994; Kováč & Hók 1996) that affected an originally shallow fold-and-thrust belt. In most areas, the Pieniny Klippen Belt forms the axial part of a broad flower structure which also includes the inner units of the Carpathian Flysch Belt and the frontal units of the Central Western Carpathians (Fig. 18.24). In NW Slovakia (Kysuce and Orava regions) the axis of the flower strucure is located north of the Pieniny Klippen Belt and, therefore, its structure is here dominated by oblique backthrusts. Additionally, the Zázrivá-Párnica sigmoid, the only important transversal structure in the Pieniny Klippen Belt, is located in this area. This is an abrupt dextral offset of *c.* 5 km affecting the otherwise linear trend of the Pieniny Klippen Belt. Its formation was related to the Late Miocene activity of a north–south trending wrench fault zone which affected the entire Central Western Carpathians (Central Slovakian Fault System; Kováč & Hók 1993).

The rock complexes cropping out in the Pieniny Klippen Belt are exclusively non-metamorphic sediments of Jurassic (very rarely also Triassic) to Palaeogene age. These are usually subdivided into morphologically positive forms (called 'klippen') formed by comparatively hard rocks (mostly Jurassic to Lower Cretaceous limestones), surrounded by less competent Upper Creaceous to Palaeogene 'klippen mantle' (marlstones, deep-marine clastics). The typical klippen structure was compared to a megabreccia, a mélange, or a chaotic mixture of lens-shaped, or often nearly isometric, rigid inclusions, several metres to kilometres in size, floating within an incompetent, highly deformed matrix. The complex internal structure of the Pieniny Klippen Belt units developed in several stages, from shallow, downward-propagating (piggyback) thrusting through tight upright folding, out-of-sequence thrusting, and imbrication, to transpression associated with the pop-up of blocks (i.e. klippen) (Fig. 18.25). However, there are some parts of the Pieniny Klippen Belt where the structure is less disintegrated and instead is dominated by tight linear folds which are generally orientated parallel to the

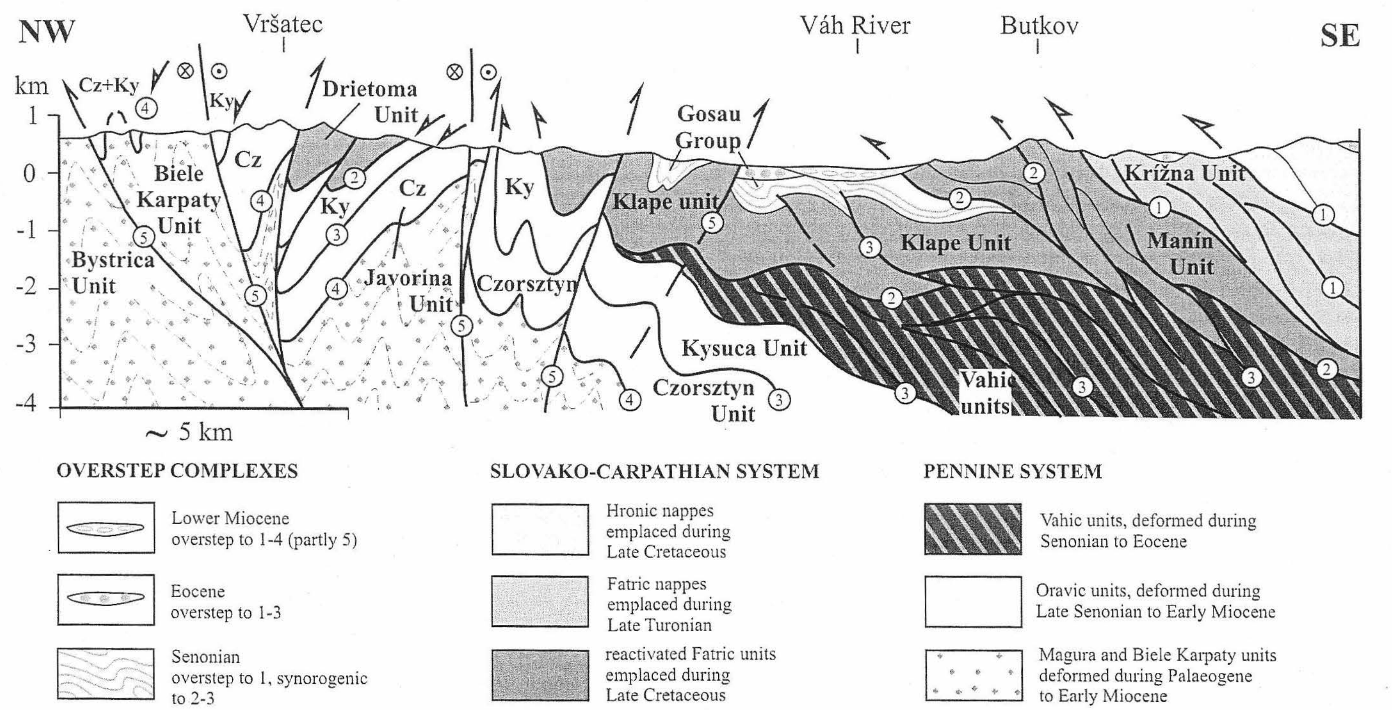



**Fig. 18.24.** Schematic cross-section through the Pieniny Klippen Belt and adjacent zones in western Slovakia (middle Váh Valley). Note that the Vršatec area represents the Pieniny Klippen Belt *sensu stricto* with tight imbricate flower structure, whereas the Váh Valley represents the Peri-Klippen Belt as a mixture of Penninic and Slovako-Carpathian units and several unconformably overlying formations. Abbreviations: Cz, Czorsztyn Unit; Ky, Kysuca-Pieniny Unit. Numbers in circles refer to the age of principal fault and fold structures: 1, late Turonian; 2, late Turonian, reactivated during Senonian; 3, Senonian to Palaeocene; 4, Palaeocene to Eocene; 5, Oligocene to Early Miocene.

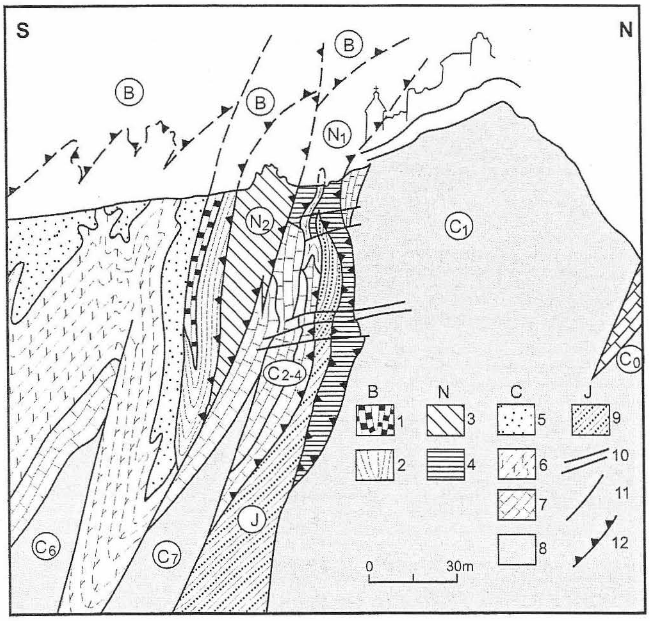



**Fig. 18.25.** A detailed profile of the Pieniny Klippen Belt illustrating its tight, imbricated small-scale structure. Niedzica castle hill, Polish Pieniny Mountains (after Birkenmajer (1999), slightly modified). Legend: B, Branisko Unit; $N_1$–$N_2$, scales of the Niedzica Unit; $C_1$–$C_6$, scales of the Czorsztyn Unit; J, Jarmuta Formation. 1, Oxfordian radiolarites; 2, Middle Jurassic shales; 3, Bajocian–Berriasian limestones; 4, Toarcian–Aalenian shales and turbidites; 5, Senonian deep-marine clastics; 6, Albian–Turonian marlstones; 7, Bajocian–Berriasian limestones; 8, Bajocian massive crinoidal limestones; 9, Maastrichtian–Palaeocene deep-marine clastics; 10, Upper Miocene transversal strike-slip faults; 11, Lower Miocene reverse faults; 12, Lower Palaeogene overthrusts.

boundaries of the Pieniny Klippen Belt. In these parts, the stratigraphic successions can also be precisely reconstructed and the relationships between the klippen and their mantle can be studied. Based on these sections, and thanks to the mainly excellent preservation of fossils in the Pieniny Klippen Belt deposits, numerous individual formations, successions or tectonic units have been defined over the last c. 150 years of intensive research (e.g. Stur 1860; Neumayr 1871; Uhlig 1903; Andrusov 1938, 1968; Scheibner in Buday *et al.* 1967; Birkenmajer 1977, 1986; Marschalko 1986; Mišík 1994). The most characteristic of these, the Oravic units, occur (at least rudimentarily) in all parts of the Pieniny Klippen Belt.

The western Slovakian sector of the Pieniny Klippen Belt is the broadest and, at the same time, the most complex part. It can be subdivided parallel to strike into the 'Klippen Belt *sensu stricto*', which is a very narrow outer strip bounding the Magura units, and the 'Periklippen Zone', which is a much wider zone (up to 15 km) with less intricate structure. The former zone only includes units derived from an independent palaeogeographic region which is designated as the **Oravic** area (Pieninic or Pienidic in older literature) and is thought to represent the (Middle) Penninic tectonic element in the Western Carpathians. The latter zone, however, also incorporates cover units undoubtedly derived from the frontal parts of the Slovako-Carpathian–Austro-Alpine tectonic system (Fig. 18.24). The tectonic mixture of both types of units leads to many uncertainties and misunderstandings which still provoke lively discussion on the origin, evolution and tectonic significance of the Pieniny Klippen Belt. Although it has often been considered to represent a suture, it should be pointed out that no ophiolite or blueschist units have been encountered. On the other hand, the deep-marine clastic formations from the klippen mantle often include boulder beds and conglomerates which are designated as exotic, since the provenance of many rock types occurring as pebbles (including ophiolites and blueschists) is not known.

*Oravic Superunit*

The Oravic Superunit is composed of two palaeogeographically contrasting units – the Czorsztyn and the Pieniny – which are interconnected by several 'transitional' successions. Both units comprise sedimentary successions whose deposition commenced in the Early Jurassic. Triassic dolomites occur only in the Maríková Klippe in western Slovakia. However, frequent dolomite clasts are present in Jurassic synrift sediments, indicating that a carbonate platform formed the original Triassic substratum of the Oravic units.

The **Czorsztyn Unit** (or Subpieniny in older literature; e.g. Uhlig 1903) represents a former ridge or high environment and was characterized by prevailing shallow-water facies during the Jurassic and Early Cretaceous (Mišík 1979*a*, 1994). It is interpreted as having been derived from a subducted continental ribbon in a Middle Penninic position (Fig. 18.26). Due to the marked dissection of the ridge as a result of several rifting events, the sedimentary formations show considerable lateral and vertical variations. Several sedimentary successions can be distinguished. The Czorsztyn succession is generally composed of (Birkenmajer 1977): (1) Middle Liassic to Aalenian deep-water, partly anoxic bioturbated marlstones and black shales ('Fleckenmergel', Allgäu Group); (2) Bajocian–Bathonian very shallow-water, sandy-crinoidal limestones (Smolegowa and Krupianka formations), or in places scarp breccias (Krasín Breccia; Aubrecht & Szulc 2006); (3) following the Upper Bathonian breakup unconformity, a thick complex of condensed 'Ammonitico rosso' nodular limestones (Czorsztyn Formation); (4) Tithonian–Berriasian coquinas, breccias, Maiolica-type Calpionella limestones and Neocomian crinoidal limestones, only locally preserved. After a hiatus indicated by numerous Neptunian dykes and partial erosion down to the Dogger deposits (Barremian–Aptian gap), the locally karstified surface was covered by an Albian hardground followed by; (5) deepening Upper Cretaceous 'couches rouges', Globotruncana-bearing marlstones (Púchov Formation); and (6) Maastrichtian–Eocene deep-marine clastics (Jarmuta and Proč formations).

Other successions related to the Czorsztyn Unit represent dissected slope environments. The ***Pruské succession*** (analogous to the ***Niedzica succession*** of the Polish sector of the Pieniny Klippen Belt, Fig. 18.27) contains mostly redeposited (allodapic/turbiditic) Dogger crinoidal limestones, intercalations of Oxfordian radiolarites within the Czorsztyn Formation, and a poorly developed Lower Cretaceous hiatus. In the ***Czertezik succession*** of the Polish Pieniny Klippen Belt, the Oxfordian radiolarites directly overlie the crinoidal limestones.

Since the Czorsztyn Unit is mainly composed of massive or thickly bedded competent limestones, it forms block- or plate-shaped klippen of various sizes (metre- to kilometre-scale) which are often in a subvertical or overturned position. These mostly occur along the outer limit of the Pieniny Klippen Belt, but in places also in a more internal position, in tectonic windows surrounded by other units of the Pieniny Klippen Belt (Fig. 18.24). Therefore, the Czorsztyn Unit is regarded as the lowermost tectonic unit, and is in a subautochthonous position with respect to all of the other Pieniny Klippen Belt units (Birkenmajer 1986). Locally it clearly overrides the innermost Magura

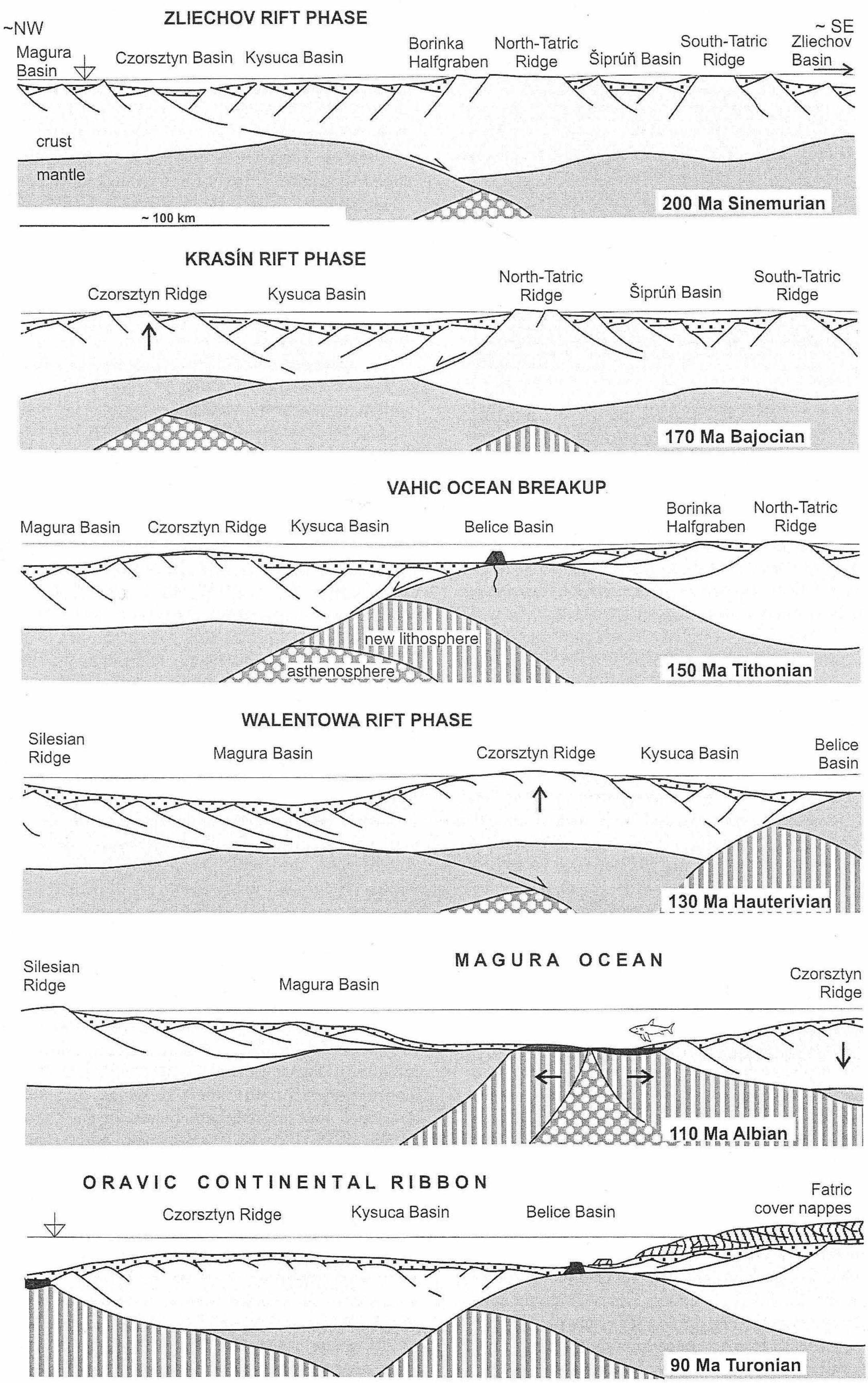



**Fig. 18.26.** Tentative reconstruction of Jurassic–Early Cretaceous rifting events in the Pennine–northern Tatric realm. Modified after Plašienka (2003*a*).

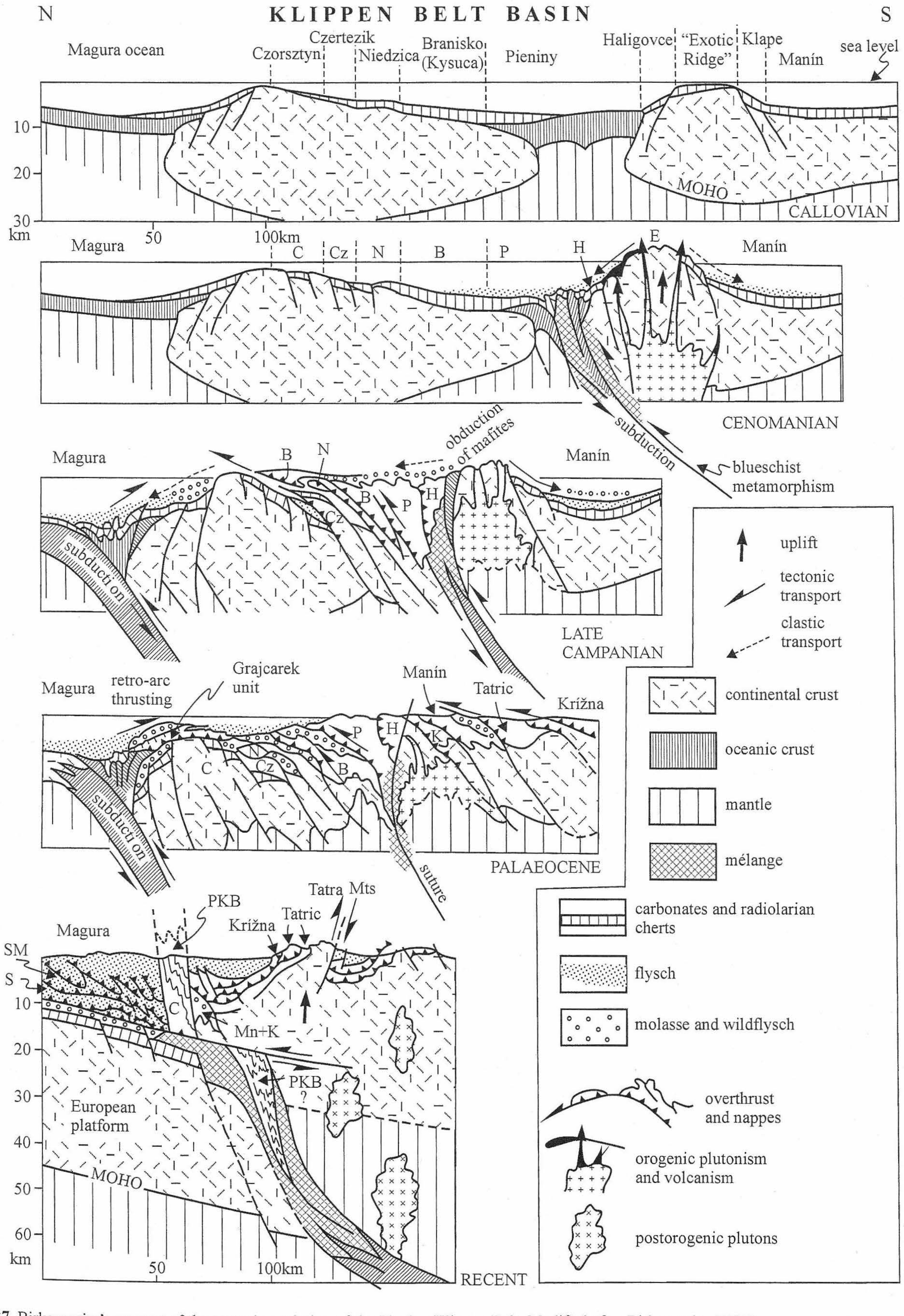



**Fig. 18.27.** Birkenmajer's concept of the tectonic evolution of the Pieniny Klippen Belt. Modified after Birkenmajer (1986).

elements (Jurewicz 1997). In places the slope successions were thrust over the ridge-derived Czorsztyn succession (Niedzica Nappe; Jurewicz 1994), indicating that the Pieniny Klippen Belt represents a former fold-thrust belt.

The **Pieniny Unit** includes basinal successions with continuous stratigraphic successions ranging from the Lias to the Late Cretaceous. The most widespread of these, the ***Kysuca succession*** (analogous to the ***Branisko succession*** in Poland), consists of: (1) lowermost Jurassic synrift siliciclastics (Gresten Formation); (2) Liassic to Aalenian hemipelagic marlstones (Allgäu Formation); (3) Middle Jurassic 'Posidonia' (*Bositra buchi*) marlstones, in places with synrift siliciclastic turbidites ('Aalenian flysch', Szlachtowa Formation), passing gradually into; (4) Callovian–Oxfordian radiolarites (Sokolica and Czajakowa formations); (5) Kimmeridgian red nodular limestones; (6) Tithonian–Neocomian Biancone- and Maiolica-type cherty limestones (Pieniny Formation); (7) various mid-Cretaceous hemipelagic marlstones; (8) Turonian–Santonian upward-coarsening, deep-marine clastics with conglomerates containing exotic pebbles (Snežnica and Sromowce formations); and (9) Campanian red pelagic 'Globotruncana' marls (Púchov Formation).

The ***Pieniny succession*** is nearly identical to the Kysuca-Branisko succession, but was deposited in even deeper waters, with radiolarites passing into the Kimmeridgian and with a very thick complex of Tithonian–Neocomian pelagic cherty limestones (Pieniny Formation). The ***Nižná succession,*** recognized from the Orava region in NW Slovakia, is marked by the presence of Barremian–Aptian platform, Urgonian-type limestones within an otherwise deep-water succession. The ***Orava (Podbiel) succession,*** of problematic affiliation, also includes Toarcian red nodular limestones (Adnet Formation).

In contrast to the Czorsztyn Unit, the klippen of the Pieniny Unit are composed of well-bedded strata that are generally folded rather than faulted. Consequently, there are long sections in this unit with continuous stratigraphic successions and quite simple macrofold structures (e.g. in the Kysuce region of NW Slovakia). Some smaller klippen of the Pieniny Unit represent disrupted cores of major folds.

### *Peri-Klippen Zone*

The Peri-Klippen Zone is typically well-developed in western Slovakia and includes several units of problematic affiliation. In the area of the westernmost Pieniny Klippen Belt it comprises the Drietoma Unit, with a succession that closely resembles the Upper Triassic–mid-Cretaceous formations of the deep-water Zliechov succession of the Krížna Nappe System (Fatricum) of the Central Western Carpathians. This unit is, therefore, considered to represent a frontal Fatric element incorporated into the Pieniny Klippen Belt following its emplacement as a nappe during the Turonian. The Manín Unit, cropping out in the inner part of the Pieniny Klippen Belt in the middle Váh valley, may also be derived from the Fatric Nappe System and may represent its shallow-water Jurassic–Lower Cretaceous Vysoká succession (the second alternative is that it corresponds to a frontal Tatric element; e.g. Rakús & Hók 2005; Birkenmajer 1986; Fig. 18.27). Generally, the 'klippen style' is poorly developed in this area and the Manín klippen are in fact brachyantiforms with more or less continuous stratigraphic successions. The most characteristic members of the Manín succession are the Barremian–Aptian (Urgonian) platform limestones and Albian pelagic marls (Butkov Formation). Along the internal boundary of the Manín Unit, the Kostolec Unit, of uncertain position, is found as a narrow strip. Its klippen are now interpreted as olistoliths of shallow-water Jurassic limestones derived from the frontal parts of the Central Western Carpathian cover nappes (Hronicum?).

In addition to all of these complications, the western part of the Pieniny Klippen Belt contains another large and most puzzling element, the **Klape Unit**. This unit crops out mainly in the middle Váh valley in western Slovakia between the Manín Unit and the Oravic units of the 'Klippen Belt *sensu stricto*' (Fig. 18.24). The Klape Unit is composed of prisms several thousand metres thick of mostly proximal deep-marine clastic to mélange ('wildflysch') complexes, which are Albian–Senonian in age. The Klape Klippe, formed by shallow-water Jurassic limestones, has recently been interpreted as a large olistolith embedded in Cretaceous deep-marine clastics (Marschalko 1986). In addition, the deep-marine clastics of the Klape Unit also contain olistoliths of Triassic carbonates. However, the most conspicuous feature of the Klape Unit is the presence of 'exotic' conglomerates and chaotic boulder beds.

The exotic, so-called Upohlav conglomerates occur in two stratigraphic levels: Albian–Lower Cenomanian and Coniacian–Santonian, and are separated by the Upper Cenomanian–Turonian shallowing-upward sequence of massive sandstones with tempestites and oyster banks (Orlové Formation). The conglomerates show a wide range of composition with numerous rock types, which have been thoroughly studied and described (e.g. Mišík *et al.* 1977, 1981, 1991; Mišík & Sýkora 1981; Marschalko 1986; Mišík & Marschalko 1988; Birkenmajer *et al.* 1990; Faryad & Schreyer 1996). In addition to many common rock types, the most noticeable exotic material includes Triassic basinal limestones, Upper Jurassic platform limestones, Urgonian limestones with serpentinite clasts, Permian A-type granites with Lower Cretaceous fission-track cooling ages (Uher & Pushkarev 1994; Kissová *et al.* 2005), large amounts of calc-alkaline volcanics of uncertain age (Permian, Upper Jurassic?), Upper Jurassic glaucophanites (e.g. Dal Piaz *et al.* 1995), and heavy-mineral spectra rich in Cr-spinel (Mišík *et al.* 1980). To explain the source of these exotic clasts, which cannot have been derived from successions in the Pieniny Klippen Belt and neighbouring zones, the concept of a short-lived Cretaceous 'exotic ridge' (Fig. 18.27), the 'Klape ridge' or 'Pieniny (ultra-Pieninic) cordillera', has been developed over many decades. Birkenmajer (1988) renamed this structure as the **Andrusov Ridge** in honour of Dimitrij Andrusov, the outstanding twentieth-century Carpathian geologist and leading expert on the geology of the Pieniny Klippen Belt.

Following the advent of plate tectonics, the exotic ridge has been interpreted as a compressional tectonic structure in an active margin setting: a complex of imbricated, obducted oceanic material or a subduction mélange temporarily cropping out on the outer structural high of an accretionary prism (Mišík 1979*b*), a subduction complex exhumed in the rear part of the South Penninic–Vahic accretionary wedge (Klape Unit; Mahel' 1989) or a magmatic island arc (Birkenmajer 1986, 1988; Fig. 18.27). The exotic pebble material would suggest that the corresponding ocean basin opened during the Triassic and was closed during the Late Jurassic–Early Cretaceous (e.g. Birkenmajer 1988; Dal Piaz *et al.* 1995). However, this is in marked contradiction to the geological record of all of the other Pieniny Klippen Belt and neighbouring units, where no such events can be documented. On the other hand, these events are documented from the southern zones of the Western Carpathians, where they were associated with the opening and closing of the Meliata Ocean. The problem remains, however, as to how this material could have been transported across the Central Western Carpathian zones (with their rugged morphology) and be deposited in the neighbourhood of the Pieniny Klippen Belt zones, where an extensional tectonic regime was active during the entire Jurassic–Early Cretaceous period (cf. Plašienka 1995*a*, 2003*a*). Therefore, Plašienka (1995*b*) proposed a hypothetical solution which considers the Klape Unit to have been derived from the Fatric Zliechov Basin, which was adjacent to the Meliatic collisional stack in mid-Cretaceous times and received exotic material from it. During the Turonian, the Klape Unit (as a part of the Fatric Krížna Nappe System) moved far to the north to occupy a position in the neighbourhood of the subsequently-formed Pieniny Klippen Belt, and was then incorporated into its structure. The Lower Senonian conglomerates with a similar composition to the Albian exotic deep-marine clastics would contain, at least partly, recycled material. If this is correct, the Klape Unit and its exotic conglomerates record important tectonic events in completely different tectonic zones in the Carpathians, which have nothing in common with events occurring in the broad area of the present-day Pieniny Klippen Belt.

Moving further east, the Pieniny Klippen Belt has a constant, *c.* 5 km width and comprises just the Oravic units, with the exception of the **Haligovce Unit** which is found in one large klippe in the Pieniny Mountains at the Polish–Slovakian border NE of the High Tatra Mountains. The presence of Urgonian limestones and some additional features make the Haligovce Unit comparable with the Manín Unit of the western sector of the Pieniny Klippen Belt. It also includes Triassic dolomites.

The structure of the Pieniny Klippen Belt in its key area in the Polish–Slovakian Pieniny Mountains was influenced by pronounced dextral transpression (Ratschbacher *et al.* 1993). Transpression also affected the adjacent Krynica Unit to the north, as well as the narrow peri-Klippen Šambron-Kamenica Zone of the Central Carpathian Palaeogene Basin (Podhale Basin in the Polish literature) to the south. This zone was formed by several en-echelon, east–west trending, slightly asymmetric, south-vergent brachyanticlines (Plašienka *et al.* 1998). Further to the east, close to the Slovakian–Ukrainian border, the backthrusts in the southern limb of the flower structure centred on the Pieniny Klippen Belt create a local antiformal thrust stack in the Humenné Mountains comprising Mesozoic (Krížna-Fatric units from the outermost part of the Central Western Carpathians; Gosau deposits) as well as Palaeogene deposits (Soták *et al.* 1997).

In many areas, especially in western Slovakia, the southern boundary of the Pieniny Klippen Belt against the Central Western Carpathians cannot be precisely defined. It often coincides with deformed Palaeocene–Lower Eocene sediments referred to as the 'Peri-Klippen Palaeogene' (currently designated as the Myjava-Hričov Group). In the westernmost parts of the Pieniny Klippen Belt and the Central Western Carpathians (Malé Karpaty Mountains), these form the upper part of the Gosau Supergroup in a position similar to that of the Gosau sediments in the Northern Calcareous Alps (e.g. Wagreich & Marschalko 1995). In sections close to the Pieniny Klippen Belt, the transition from the Senonian to the Palaeocene does not contain a stratigraphic gap, and Palaeogene deposits seem to form a continuation of the 'klippen mantle'. To the south, however, the Cretaceous–Tertiary boundary is marked by an unconformity (Salaj & Began 1983). There, the Senonian Brezová Group rests transgressively on Triassic carbonates of the Hronic Nappe System (analogous to the Upper Bajuvaric and Lower Tyrolic nappes of the Northern Calcareous Alps). Gosau sediments in the Northern Calcareous Alps and Malé Karpaty Mountains are connected by the prolongation of the Giesshübl Syncline which was drilled in the substratum of the Neogene Vienna Basin (e.g. Wessely 1992). Further to the NE in the central Váh Valley, Palaeocene to Lower Eocene sediments of the Myjava-Hričov Group are closely related to those of the Central Carpathian Palaeogene Basin which extends far to the south, overlying the cover nappes of the Central Western Carpathians. In the areas close to the Pieniny Klippen Belt, the transgressive base of the Central Carpathian Palaeogene Basin is formed by exceptionally thick Eocene dolomite breccias (Súl'ov Conglomerates; Marschalko & Samuel 1993). In general, the Gosau-type Brezová and Myjava-Hričov groups are dominated by pelagic marls and calcareous deep-marine clastic formations with shallow-water biogenic detritus and reef-derived olistoliths. They were presumably deposited in narrow compressional or transpressional basins within the rear part of the developing Vahic-Oravic accretionary wedge along the northern edge of the Central Western Carpathian units.

In inner zones of the Central Western Carpathians, south of the Pieniny Klippen Belt, only rare occurrences of the Senonian-age Gosau sediments are known (Šumiac village and Dobšiná Ice Cave in central Slovakia). These represent remnants of an originally extensive marine channel which crossed the Central Western Carpathian area from northern Hungary. This interpretation is based on drilling of the pre-Tertiary basement in southern Slovakia (Mišík 1978). Other occurrences of Senonian sediments are known from the Slovak Karst area (Gombasek, Miglinc) and from the Uppony Mountains in NE Hungary.

In the Polish sector of the Pieniny Klippen Belt, carbonate conglomerates of the Jarmuta Formation transgressively overlie various klippen successions with an angular discordance, thus suggesting that there was an important phase of pre-Palaeogene ('Laramian') deformation within the Pieniny Klippen Belt (Birkenmajer 1970, 1986). This concept has been questioned, however, by some workers who emphasized that there was a record of continuous sedimentation through the Senonian and across the Cretaceous–Palaeogene boundary (e.g. Sikora 1974). However, this continuous record is probably applicable only to some 'marginal' Pieniny Klippen Belt successions (Hulina and Zlatna according to Sikora 1974). In eastern Slovakia, there are several localities where carbonate conglomerates containing olistoliths of klippen successions (Gregorianka Breccia; Nemčok *et al.* 1989) occur as olistostromes within the Palaeocene–Eocene deep-marine clastic sequence (Proč Formation). Indeed, Nemčok (1980) regards the entire Pieniny Klippen Belt as a gigantic olistostrome body, although this opinion is not accepted by any other workers in the area.

### *Vahic Superunit*

The origin and evolution of the Pieniny Klippen Belt are closely related to the long-term problem of the continuation of the South Penninic (Piemont-Liguria) Ocean into the ancient Western Carpathian realm. Since there are no surface exposures of ophiolites or Bündnerschiefer complexes in the Western Carpathians, such a continuation has been considered speculative. Some authors have assumed that the deep-water Fatric (Krížna-Zliechov) Trough represented this continuation (Tollmann 1978; Kozur & Mock 1996), but this is obviously a Slovako-Carpathian–Austro-Alpine element that was underlain by thinned continental crust and its tectonic history is very different from that of the Penninic zones. Since the Slovako-Carpathian units of the Central Western Carpathians clearly form a lateral continuation of the Austro-Alpine system, and the Oravic units have many common features with the Middle Penninic zones (Briançonnais), the majority of authors consider that this hypothetical ocean was located between the present Pieniny Klippen Belt and the Central Western Carpathians, where its suture is obliterated by superimposed nappe units and unconformably overlying sedimentary complexes. This ocean has been named 'Ocean X' (Birkenmajer 1986, 1988) or the Vahic Ocean, and its vestiges were referred to as the Vahicum (Mahel' 1981). However, there has never been any general agreement as to which particular unit represents relicts of the Vahic Ocean (for Mahel' (1981) it was the Klape Unit).

Tomek (1993) interpreted a series of seismic reflections on the Carpathian deep seismic profile 2T, which underlie the Tatric basement sheet in the northern parts of the Central Western Carpathians, to represent the shallow-dipping Piemont-Ligurian suture. Plašienka (1995*a*, *b*) assigned the **Belice Unit** to the Vahicum and assumed that this unit is the only Vahic element cropping out at the surface. It is exposed in small tectonic windows in the Považský Inovec Mountains (some 10 km south of the Pieniny Klippen Belt). The Belice Unit is composed of a thin (only several tens of metres) but strongly imbricated pelagic succession of Upper Jurassic red radiolarites and Lower Cretaceous siliceous slates (closely resembling the 'Palombini Shales' of the Ligurian Alps) with thin intercalations of Calpionella-bearing limestones in the lower part. This is overlain by a thicker, coarsening-upward Senonian deep-marine clastic succession which is topped by chaotic breccias composed of material from the Tatric basement thrust sheet that overrides the Belice Unit. Thus, the Belice Unit is positioned below the Tatric thrust sheet and its Infra-Tatric elements, representing the most external Slovako-Carpathian units. Volcanic material possibly derived from oceanic crust of the Vahic Ocean is present only as small fragments and olistoliths in the Senonian deep-marine clastics. The Belice Unit crops out in the northern and southernmost parts of the Považský Inovec Moutains. The outcrops in the north formed as a result of antiformal thrust stacking, and those in the south resulted from extensional exhumation during Miocene formation of the Danube Basin (Fig. 18.28).

The **Iňačovce-Krichevo Unit** is another element possibly representing the Carpathian analogue of the (South?) Penninic Ocean (Fig. 18.20A). This is known solely from deep boreholes that extended into the pre-Neogene substratum of the East Slovakian Basin south of the Pieniny Klippen Belt. The unit comprises Triassic carbonates and slates (similar to the Quartenschiefer in the Helvetic units), thick low-grade metasediments strongly resembling the Bündnerschiefer of the Tauern Window, Eocene deep-marine clastics, and serpentinite bodies (Soták *et al.* 1993, 1994, 2000).

## *Tectonic evolution of the Pieniny Klippen Belt*

Based on the existing data, mostly from western Slovakia, a series of evolutionary stages can be reconstructed for the Magura, Oravic, Vahic and other units involved in the Pieniny Klippen Belt and adjacent zones (Plašienka, 1995*a*, 2003*a*; Figs 18.26 & 18.27).

1. Triassic: establishment of a carbonate platform in the Oravic domain.
2. Liassic to Aalenian: broad symmetric rifting and related tectonic subsidence, resulting in the formation of a series of half-grabens with sedimentation under mainly anoxic conditions.
3. Bajocian: strongly asymmetric rifting phase with development of the Czorsztyn Ridge due to thermal uplift above a lithospheric-scale, north-dipping detachment fault.
4. Bathonian: continental breakup on the internal side of the Czorsztyn Ridge, resulting in the opening of the South Penninic-Vahic Ocean with the Kysuca-Pieniny Basin being located on its northern (in present-day coordinates) flank.
5. Callovian to Tithonian: thermal subsidence of the entire Oravic domain.
6. Early Neocomian: renewed asymmetric rifting and thermal uplift of the Czorsztyn Ridge; development of a south-dipping detachment fault on the northern side of the Czorsztyn Ridge.
7. Late Neocomian: breakup of the North Penninic-Magura Ocean to the north of the Czorsztyn Ridge.
8. Mid-Cretaceous to Senonian: thermal subsidence, resulting in the transformation of the Czorsztyn Ridge into a pelagic high.
9. Turonian: nappe emplacement of frontal elements of the Krížna (Fatric) Nappe System of the Central Western Carpathians (Drietoma, Klape, Manín units) onto the southern parts of the Vahic Ocean.
10. Early Senonian: onset of subduction of the Vahic Ocean lithosphere below the Central Western Carpathians; deformation of the Krížna elements in the position of a 'false' accretionary complex, erosion and resedimentation of Albian conglomerate material, including exotic pebbles, from the Klape Unit into deep-marine clastic deposits of the Pieniny Unit.
11. Late Senonian: gradual closure of the Vahic Ocean; partial inversion of the Magura Ocean.
12. Maastrichtian to Palaeocene: final closure (still ongoing) of the Vahic Ocean with the formation of numerous narrow remnant and piggyback deep-marine clastic basins; collision of the accretionary complex with the Czorsztyn Ridge; detachment and thrusting of the internal Oravic units along Liassic black shale horizons and formation of a foreland fold-and-thrust belt.
13. Eocene: onset of subduction of the Magura Ocean; detachment of the Czorsztyn Unit from its basement which was underthrust beneath the Central Western Carpathians; duplexing, recumbent folding, and thrusting of the Czorsztyn Unit over the southern Magura elements.
14. Oligocene to earliest Miocene: closure of the Magura Ocean; dextral transpression and oroclinal bending in the Pieniny Klippen Belt due to counterclockwise rotation of the Central Western Carpathian block; development of a positive flower structure usually centred by a narrow, generally vertical zone within the Pieniny Klippen Belt *sensu stricto*, in which strike-slip movement predominated, leading to the formation of the typical 'klippen' tectonic style as a result of pervasive brittle faulting.
15. Middle to Late Miocene: sinistral transtension along the western, SW–NE trending sector of the Pieniny Klippen Belt; the NW–SE trending eastern Slovakian sector underwent only dextral wrenching: initial transpression during the Early–Middle Miocene, followed by transtension and later extension during the Late Miocene and Pliocene (leading to the opening of the Transcarpathian Basin); volcanic activity in the so-called Pieniny andesite line along the eastern Pieniny Klippen Belt, which extends from the Polish Pieniny Mountains (Mount Wżar) to the eastern Slovakian–Ukrainian chain of Sarmatian–Pannonian subduction related stratovolcanoes.

## *Central Western Carpathians*

### *Regional subdivision*

The Central Western Carpathian region forms the core of the Carpathian Orogen which underwent a complex evolution and

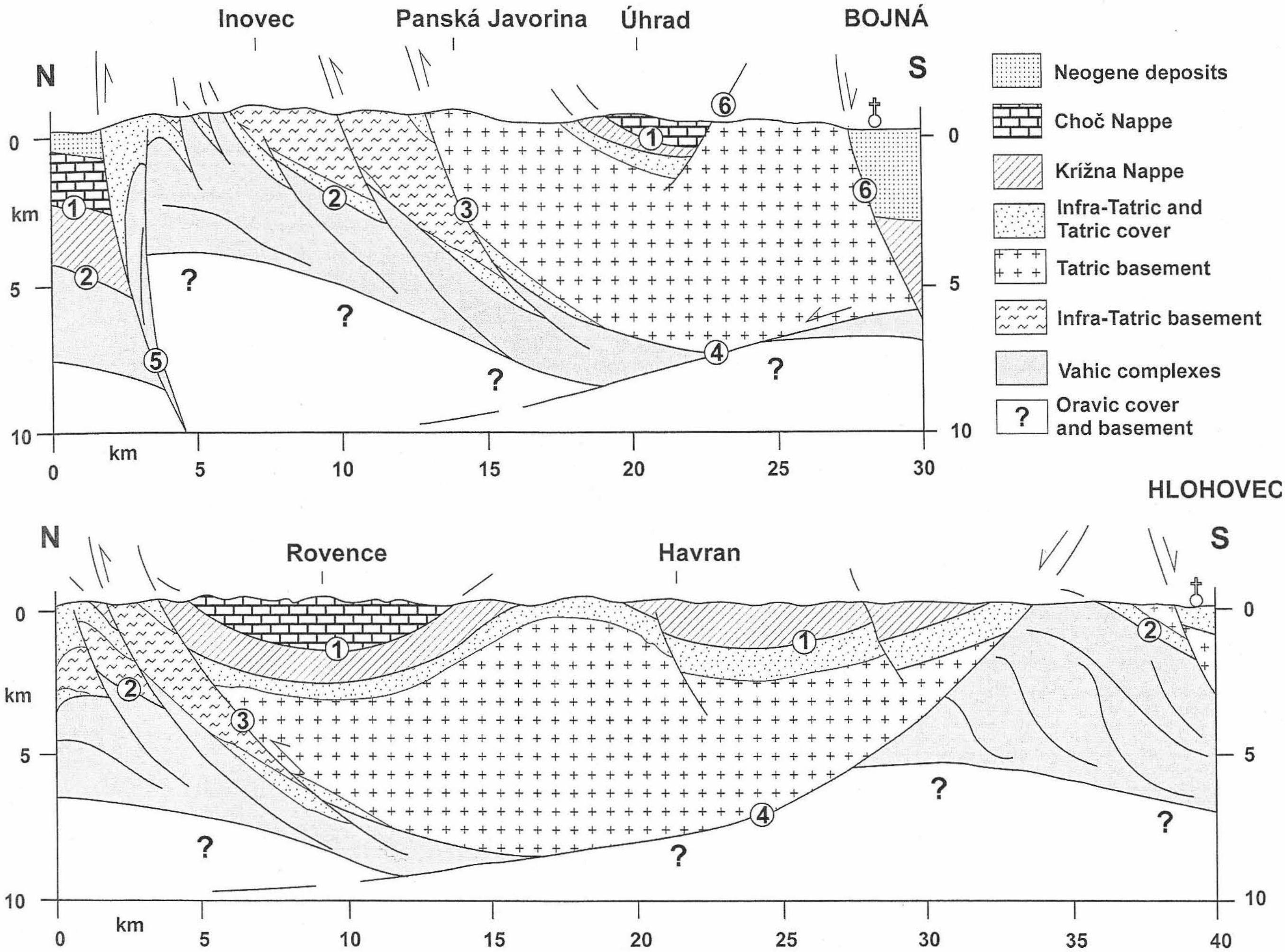



Fig. 18.28. Deep profiles of the Považský Inovec Mountains horst (after Plašienka 1999*b*). Numbered fault structures: 1, upper Turonian overthrust planes of the Fatric (Krížna) and Hronic (Choč) cover nappe systems; 2, uppermost Cretaceous overthrust of the Slovako-Carpathian (Infratatric–Tatric and overlying) units onto the Vahic Belice Unit; 3, early Palaeogene out-of-sequence reverse faults; 4, Middle Miocene low-angle normal faults; 5, Upper Miocene rotational oblique reverse faults; 6, Upper Miocene normal faults.

contains a number of nappe units of various orders and ages. Two main types of tectonic units are present: thick-skinned thrust sheets which comprise the pre-Alpine crystalline basement along with its Late Palaeozoic–Mesozoic sedimentary cover (the Tatric, Veporic and Gemeric superunits), and detached cover nappe systems containing Late Palaeozoic to Mesozoic sedimentary rocks with rare volcanics (the Fatric, Hronic and Silicic superunits) (e.g. Andrusov 1968, 1975; Andrusov *et al.* 1973; Mahel' 1986; Biely 1989; Plašienka *et al.* 1997).

The Southernmost Central Western Carpathian zones were overridden by nappe units belonging to the Internal Western Carpathians (Meliatic and partly Turnaic Superunits). The main phase of crustal shortening and nappe formation took place during the early Late Cretaceous in the Central Western Carpathians. Deformation processes reflect an outward (northward) vergence of the orogen. Regionally, the Central Western Carpathians are divided into two principal morphostructural zones: the Tatra-Fatra Belt and the Vepor-Gemer Belt.

The **Tatra-Fatra Belt** includes so-called core mountains in western and northern Slovakia with a Tatric-type (see below) crystalline basement and the classic 'Sub-Tatra' nappes (in the sense of Uhlig 1907; Andrusov 1936). These mountains represent mainly asymmetric, northward-tilted horst structures (*c.* 40–50 km long and 15–20 km wide on average) resulting from Late Tertiary uplift, surrounded by cover nappe units and Tertiary sedimentary basins. There are nine principal core mountains, namely, from west to east: the Malé Karpaty (Lesser Carpathians), Považský Inovec, Tribeč, Strážovské vrchy, Žiar, Malá Fatra, Vel'ká Fatra, Nízke Tatry (Low Tatra), and Tatry (High Tatra) mountains (Figs 18.19 & 18.29).

The core mountains comprise the Tatric Superunit including Variscan high-grade crystalline basement and its Late Palaeozoic and Mesozoic sedimentary cover, overridden by the superficial nappe systems: the Fatric Superunit (Krížna nappe *sensu lato*) and the Hronic Superunit (Choč nappe *sensu lato*) (Fig. 18.30). The post-nappe cover is formed by Palaeogene (Central Carpathian Palaeogene Basin), Neogene and Quaternary rocks. The northern boundary of the Tatra-Fatra Belt is formed by the above-noted, locally ill-defined contact with the Pieniny Klippen Belt. The southern boundary, against the Vepor-Gemer Belt, is the contact between the Tatric sheet and the overriding Veporic crystalline basement, which is a crustal-scale thrust fault (the

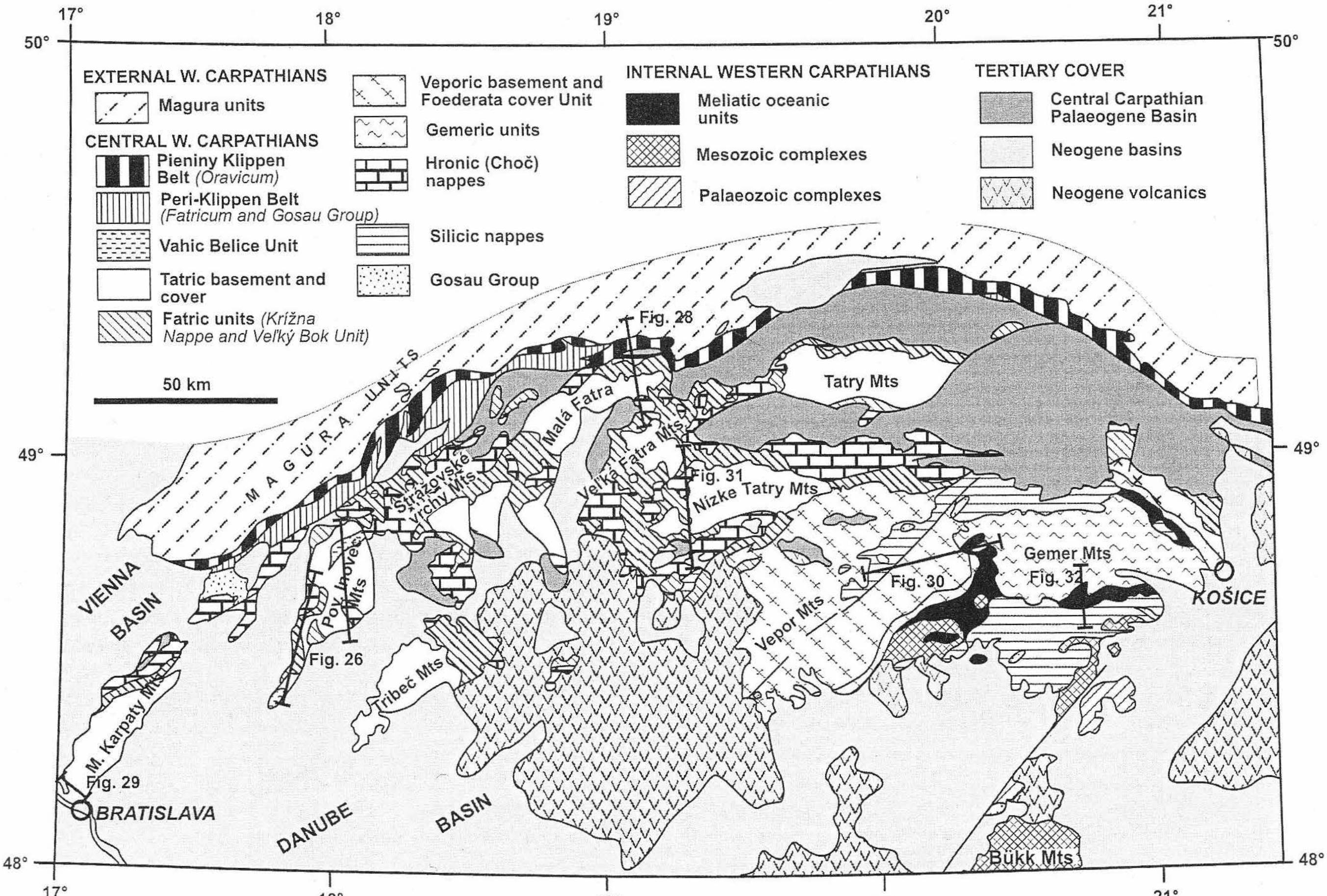



**Fig. 18.29.** Simplified geological map of the Central and Internal Western Carpathians. All boundaries of pre-Tertiary units are tectonic in origin.

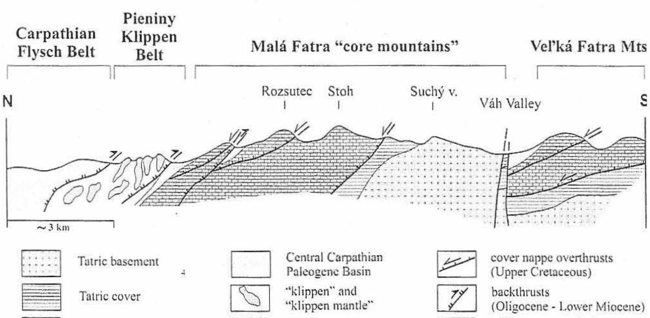



**Fig. 18.30.** Cross-section of typical 'core mountains' – the Malá Fatra Mountains of NW Slovakia (modified after Polák 1979). Note backthrusting that modified the nappe structures in the proximity of the Pieniny Klippen Belt.

Čertovica Fault). However, this fault is precisely defined only in the Nízke Tatry Mountains, where it is exposed over a short distance.

The westward continuation of the Čertovica Fault is beneath the Central Slovakian neovolcanic area and to the south of the Tribeč Mountains, where it was reactivated as a Neogene extensional normal fault system (Mojmírovce faults; Hrušecký *et al.* 1996; Hók *et al.* 1999). A further possible SW prolongation of this fault below the Neogene fill of the Danube Basin has not been verified (see Fig. 18.19). In the subsurface of the western part of the Danube Basin, only the basement units in the vicinity of the Malé Karpaty Mountains and the Hainburg Hills (and possibly the Leitha Mountains in the easternmost part of the Alps) can be considered to be Tatric. No typical Tatric elements occur in the Eastern Alps; here they are replaced by the Lower Austro-Alpine units. These are equivalent to the Infra-Tatric units in the Western Carpathians, a frontal, imbricated part of the Tatric Superunit. East of the Nízke Tatry Mountains, the Čertovica Fault is covered by the Hronic Nappes and Palaeogene sediments of the Central Carpathian Palaeogene Basin, and somewhere to the north of the Branisko Mountains it probably joins the Pieniny Klippen Belt.

The **Vepor-Gemer Belt** is the metamorphic core of the Western Carpathian Orogen and is predominantly built up of basement complexes, which were exhumed from deeper crustal levels early in the Alpine Orogeny. The Vepor-Gemer Belt includes the Kráľovohoľské Nízke Tatry Mountains, the Slovak Ore Mountains (Veporské, Stolické and Volovské vrchy mountains), and the Branisko and Čierna Hora mountains to the NW of Košice. These areas are constituted either by Veporic or Gemeric basement and cover complexes. In addition to the dominant underlying Veporic Superunit, the northern Vepor Sub-belt also includes remnants of the overriding Gemeric Superunit and large nappe outliers of the Silicic Superunit (Drienok, Muráň, Vernár and Stratená nappes). Unconformable cover complexes include Senonian Gosau sediments, Palaeogene sedimentary, and Neogene sedimentary and volcanic complexes.

The contact of the Vepor Sub-belt with the Gemer Sub-belt to the south is represented by the Lubeník-Margecany Fault. This is a complex, arcuate fault zone that underwent a complex kinematic evolution. Its precursor was a thrust plane along which the Gemeric Superunit was thrust over the Veporic Superunit, probably during the Early Cretaceous. During the Late Cretaceous, the SW–NE striking Lubeník sector of the Lubeník-Margecany Fault acted as a sinistral transpressional zone. Simultaneously, its presently NNW–SSE striking part was reactivated as a low-angle normal fault which accommodated top-to-the-east unroofing of a metamorphic core complex constituted by the Veporic Superunit (Plašienka *et al.* 1999).

The Vepor Sub-belt plunges to the SW beneath the Central Slovakian volcanic complexes and then continues in the subsurface of the southeastern part of the Neogene Danube Basin between the Mojmírovce faults in the north (reactivated Čertovica Fault) and the Rába-Hurbanovo Fault Zone (Figs 18.19 & 18.20). According to the borehole data, the Vepor Sub-belt in this area is dominantly composed of basement complexes. It continues westward into the Central Eastern Alps.

The Gemer Sub-belt includes the eastern part of the Slovak Ore Mountains and comprises mainly rocks belonging to the Gemeric Superunit: Palaeozoic, mainly low-grade metamorphic volcanosedimentary complexes with possible Mesozoic-age cover. It is overlain by remnants of the Jaklovce and Bôrka nappes belonging to the Meliatic Superunit and by cover nappes belonging to the Silicic Superunit (Stratená and Galmus nappes along the northern margin, Radzim and Opátka outliers in the central part, and the northern margin of the Silica Nappe *sensu stricto* at the boundary with the Slovak Karst Mountains). Tertiary cover rocks are present only at the peripheries of the exposed part of the sub-belt. The southern boundary of the sub-belt is poorly defined but it is herein considered to coincide with the course of the Meliatic suture separating the Central Western Carpathians and the Internal Western Carpathians. This suture was probably reactivated by fault zones such as the Rožňava Fault Zone, continuing toward the west and SW into the Plešivec, Diósjenő, Hurbanovo and Rába fault zones (Plašienka 1999*a*).

## *Slovako-Carpathian tectonic system*

### *Tatric Superunit*

The Tatric Superunit (or Tatricum) represents the most external and lowermost component of the Slovako-Carpathian tectonic system. Its internal structure is dominated by pre-Alpine crystalline basement complexes, though Mesozoic cover rocks are common at the surface. As revealed by deep seismic transects, the Tatric Superunit forms a tabular, slightly convex, >10 km thick upper-crustal body, rooted in the lower crust below the wedge-shaped Veporic Superunit (Tomek 1993; Bielik *et al.* 2004; see Fig. 18.20). At depth, the Tatric Superunit presumably overrides the Vahic Superunit. It is overlain by thin-skinned nappes of the Fatric and Hronic superunits, as well as by unconformably overlying Tertiary sediments and volcanics.

Compared to the overlying Veporic Superunit, the Tatric Superunit is internally less differentiated. Late Tertiary extensional faults (mainly the bounding faults of the horsts forming the core mountain), and minor north-vergent overthrusts along

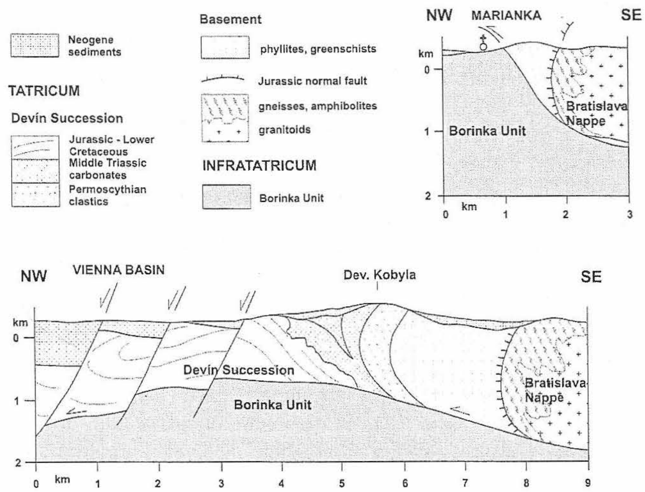



**Fig. 18.31.** Large-scale recumbent fold at the tip of the Tatric Bratislava nappe in the southern part of the Malé Karpaty Mountains (after Plašienka 1999*b*).

the northern edges of some of the core mountains (Tribeč, Ďumbierske Nízke Tatry, High Tatra) are also documented. The latter formed as a result of Cretaceous inversion of Jurassic extensional faults (Dumont *et al.* 1996; Plašienka 2003*b*). Major overthrusts are present along the northern edge of the Tatric Superunit, where Tatric units were thrust over various complexes of the so-called Infra-Tatricum (e.g. the Tatric Bratislava Nappe overriding the Infra-Tatric Borinka Unit in the Malé Karpaty Mountains; Plašienka 1990; Plašienka *et al.* 1991; Fig. 18.31).

The mid-crustal suture between the rigid Tatric basement sheet and the underlying Vahic complexes represents a crustal inhomogeneity coinciding with the ductile–brittle transition. It was, therefore, reactivated during Palaeogene and Miocene extension. Listric extensional faults flattening into the ductile middle crust were imaged by seismic profiling particularly in the Danube Basin area (Tomek *et al.* 1987; Tari *et al.* 1992; Tari 1996; Horváth 1993; Hrušecký 1999). The weak mid-crustal layer probably also controlled the dissection of the Tatric sheet in Tertiary times into a system of uplifted horsts (forming the core mountains) and subsiding grabens in the areas of Neogene extension (mainly in western Slovakia) and the rotation of these blocks (Kováč *et al.* 1994, 1997; Hrouda *et al.* 2002*a*).

The Tatric Superunit consists of pre-Alpine crystalline basement complexes and their Upper Palaeozoic–Mesozoic sedimentary cover. Upper Palaeozoic rocks are mainly present in the Infra-Tatric units. The Tatric crystalline basement was only slightly affected by Alpine deformation and metamorphism; it commonly retains Variscan structures and isotopic ages. The Tatric sedimentary cover is autochthonous to parautochthonous with respect to the basement.

The pre-Alpine, generally Variscan crystalline basement includes medium- to high-grade, rarely low-grade metamorphic volcanosedimentary complexes which were intruded by numerous differentiated Variscan granitoid plutons (Petrík *et al.* 1994; Petrík 2000; Broska & Uher 2001). In the western High Tatra Mountains, a Variscan nappe structure was documented (Kahan 1969; Janák 1994). It has been suggested that this nappe structure controls the distribution of metamorphic complexes within the Tatric and also the Veporic basement (Putiš 1992; Bezák *et al.* 1997).

The Tatric sedimentary cover is largely of Mesozoic age (Lower Triassic–Lower Turonian), although immature terrestrial clastics of probable Late Permian age are found locally (Vozárová & Vozár 1988). Triassic sediments are comparable to the record in the German Basin. Scythian strata are continental siliciclastics. Lower Scythian quartz sandstones (Lúžna Formation) usually directly overlie the pre-Alpine basement. The Upper Scythian Werfen Formation is composed of sandstones, siltstones, shales and, in the upper part, evaporites (dolomite, gypsum, cargneule). The Middle Triassic Gutenstein Formation represents a shallow carbonate ramp and includes evidence of hypersaline environments and submarine slumping. It is overlain by a partly hypersaline dolomitic complex (Ramsau Formation) with shale intercalations in its upper part. The maximum thickness of the Middle Triassic carbonate complex is *c.* 500 m. The Upper Carnian–Norian Carpathian Keuper Formation includes shales, quartz sandstones, rare conglomerates, and minor evaporites. Due to erosion associated with Lower Jurassic rifting, Rhaetian-age shallow-marine sediments are rarely preserved. Rhaetian limnic sediments with dinosaur footprints (Tomanová Formation) occur in the High Tatra Mountains.

The Jurassic sediments of the Tatric area were deposited on two bounding structural highs (south and north Tatric ridges) and in the intervening Šiprúň Basin (Fig. 18.26). The synrift Lower Jurassic sediments are represented by biodetritic and sandy limestones and sandstones and, in the Šiprúň Basin, also by the hemipelagic Allgäu Formation. While the ridges maintained their elevated position up into the Late Jurassic, the Middle and Upper Jurassic strata in the Šiprúň Basin are chiefly represented by a deep-water pelagic facies (cherty and siliceous limestones to radiolarites). The Upper Jurassic–Lower Cretaceous succession comprises limestones with chert nodules. Bioclastic limestones were widespread during the Barremian. Terrigenous allodapic sandy limestones of that age surround the north Tatric ridge (Jablonský *et al.* 1993). In the area of the south Tatric ridge (High Tatra Mountains), Urgonian platform limestones and related slope deposits occur. Small bodies of basanitic lavas locally occur within the Barremian and Aptian sediments (Spišiak & Hovorka 1997). Synorogenic turbidite sediments dominated by terrigenous siliciclastic material, together with bodies of exotic conglomerates (Poruba Formation), were deposited during the Late Albian and Cenomanian. Hemipelagic sedimentation ceased as late as the Early Turonian.

The **Infra-Tatric units** comprise the lower and frontal, imbricated parts of the Tatric sheet and crop out in the western part of the Tatra-Fatra Belt. The Borinka Unit of the Malé Karpaty Mountains (Fig. 18.31), the Inovec Nappe of the northern part of Považský Inovec Mountains, and the Kozol Unit in the Malá Fatra Mountains can be regarded as Infra-Tatric units (Plašienka *et al.* 1997). Units underlying the Infra-Tatric units crop out only in the Považský Inovec Mountains (Belice Unit of supposed Vahic provenance; Plašienka 1995*b*). Substantial portions of the Lower Austro-Alpine complexes occurring in the easternmost part of the Eastern Alps (Wechsel, Grobgneis and Semmering nappes) can be considered to form a westward prolongation of the Infra-Tatric units (Pahr 1991; Häusler *et al.* 1993). The Infra-Tatric units represent a system of recumbent fold nappes. They are exposed in uplifted areas forming tectonic windows and half-windows. The thrust planes of the nappes are accompanied by ductile/brittle shear zones formed under low-grade metamorphic conditions (*c.* 250–300°C; Putiš 1991; Plašienka *et al.* 1993; Korikovsky *et al.* 1997).

The Infra-Tatric basement is exposed only within the Inovec Nappe, and is composed mainly of mica schists and rarer amphibolites and metagranotoids. It is covered by the thick Upper Palaeozoic Kálnica Group, consisting principally of Permian continental clastics; similar sediments occur in the Kozol Unit of the Malá Fatra Mountains. The Permian sediments are overlain by Scythian-age clastics and by a Middle Triassic-age carbonate complex. Upper Triassic-age formations are often missing due to erosion associated with Early Jurassic rifting. Jurassic synrift sediments consist of prisms up to 1000 m thick of marine terrigenous clastics (Borinka Unit of the Malé Karpaty Mountains, interpreted as the infill of an extensional half-graben; Plašienka 1987; Plašienka *et al.* 1991; Fig. 18.26).

*Veporic Superunit*

The Veporic Superunit (or Veporicum) is the central of the three crustal-scale thick-skinned thrust sheets of the Central Western Carpathians. Based on the 2T and G1 deep seismic profiles (Tomek 1993; Vozár *et al.* 1998; Bielik *et al.* 2004), it can be characterized as a wedge-shaped, partly imbricated crustal slice. It is *c.* 15–20 km thick in its central part and up to 30 km thick in the rear part, where it occupies almost the entire crustal profile (profile 2T). Towards the east (profile G1), the Veporic Superunit is only *c.* 5 km thick in the northern part and 8–10 km thick in the southern part (see Fig. 18.20). In seismic profiles, the inner structure of the Veporic Superunit is complex, with numerous reflective zones, roughly paralleling its boundaries. These reflectors are interpreted as mylonitic zones created during the palaeo-Alpine (Cretaceous) collision and subsequent post-collisional extension (e.g. Hrouda *et al.* 2002*b*). Some of the reflections (i.e. the subhorizontal or slightly northward-dipping ones) can be considered as pre-Alpine, related to Variscan thrusting and collisional processes (Bezák *et al.* 1997; Bielik *et al.* 2004).

At the surface, the Veporic Superunit mainly comprises pre-Alpine crystalline basement, with a similar composition and pre-Alpine history to that of the Tatric Superunit. Its Upper Palaeozoic–Mesozoic cover is only locally preserved. The basement is, for the most part, composed of probable Lower Palaeozoic volcanosedimentary complexes with a low- to medium-grade Variscan metamorphic overprint. High-grade migmatitic and amphibolite complexes occur in the northern and central parts of the Veporic Superunit, where they are intruded by the Vepor Pluton and smaller granitoid massifs (Hrončok, Rimavica and Cretaceous Rochovce granitoid massifs; Putiš *et al.* 2000; Poller *et al.* 2001). In contrast to the Tatric Superunit, the basement of the southern Veporicum was considerably influenced by palaeo-Alpine tectonometamorphic processes, especially by low- to medium-grade metamorphism and the development of a penetrative subhorizontal mylonitic structure that superimposed the Variscan metamorphic features and tectonic relationships of the pre-Alpine complexes (Hók *et al.* 1993; Madarás *et al.* 1996; Plašienka *et al.* 1999; Janák *et al.* 2001).

The sedimentary cover of the northern part of the Veporic Superunit, the **Vel'ký Bok Unit**, represents the southern slope of the Fatric Basin. The Vel'ký Bok Unit crops out on the northern slopes of the Král'ovohol'ské Nízke Tatry Mountains, in the NW part of the Slovenské rudohorie Mountains, and in the Branisko and Čierna Hora Mountains. It is composed of a Permian–Lower Cretaceous sedimentary succession similar to that of the Zliechov succession of the Fatric Superunit (see below), and is subdivided into several partial units, commonly forming large recumbent folds (Plašienka 1995*c*, 2003*b*).

Thick (>1000 m in places) Permian clastics, including a volcanic horizon (L'ubietová Group), directly overlie the northern Veporic crystalline basement and grade up into Scythian quartzites and shales. The Middle Triassic carbonate complex comprises limestones and dolomites. The Upper Triassic is represented by the Carpathian Keuper Formation. The Jurassic commences with synrift sandy limestones, with local gaps in sedimentation and occasional erosion of the underlying Triassic. The post-rift units are mainly Middle Jurassic to Lower Cretaceous pelagic sediments (i.e. marly, cherty and siliceous limestones, locally also radiolarites). The Neocomian–Lower Albian pelagic marlstones attain a thickness of >1000 m. Younger sediments are not known from the Vel'ký Bok Unit.

The **Foederata Unit** represents the cover of the southern and central zones of the Veporicum. However, it is poorly preserved. In the southernmost part of the Veporic Superunit, along the Lubeník Fault, only Upper Palaeozoic and Scythian-age units are preserved (Revúca Group; Vozárová & Vozár 1988). The Foederata Unit in the central Veporicum comprises Permian–Triassic metamorphosed sedimentary complexes made up of Permian–Scythian-age clastics and a Middle to Upper Triassic carbonate succession. All of the rocks of the Foederata Unit are metamorphosed to greenschist facies and underwent ductile deformation (Plašienka 1993; Lupták *et al.* 2000). However, they remained in a parautochthonous position with respect to their basement.

The structural association of the central and southern parts of the Veporic Superunit is dominated by a flat, or moderately NE-dipping, metamorphic/mylonitic $S_1$ foliation that is penetrative in both the topmost parts of the basement granitoids and the cover (Foederata Unit). The foliation planes exhibit a distinct stretching lineation $L_1$ plunging generally to the east. The association of this first Alpine deformation stage $D_1$ is completed by moderately east- to NE-dipping shear bands which are mesoscopically penetrative in the vicinity of the NW–SE trending sector of the Lubeník line (C′-type shear bands). Shear bands and other shear-sense criteria indicate top-to-the-east kinematics (Fig. 18.32) (i.e. orogen-parallel extension). The growth of new metamorphic minerals (micas, chloritoid, kyanite) within the cover rocks is generally syn- to early post-kinematic with respect to the $S_1$ foliation, and pre- to synkinematic with respect to the C' planes (Lupták *et al.* 2000).

The $D_1$ structural association represents a large-scale subhorizontal ductile shear zone. This shear zone is interpreted as a low-angle detachment fault that parallels the basement–cover interface and lithological boundaries within the Foederata Unit, as well as the original Veporicum/Gemericum overthrust contact (Plašienka 1993; Hók *et al.* 1993; Madarás *et al.* 1996). Basement complexes in the footwall of the detachment exhibit abruptly downward-decreasing strain and increasing metamorphic grade due to vertical thinning and the telescoping of isograds within the shear zone. The maximum P-T conditions reach the lower amphibolite facies along a prograde loop, up to 620°C at about 1 GPa (Janák *et al.* 2001) in the deepest exposed unit (garnetiferous mica schists). The Foederata Unit indicates metamorphic conditions of 530–560°C at 600–800 MPa for Permian metasediments (chloritoid-kyanite schists; Lupták *et al.* 2000) and only 330–380°C at *c.* 400 MPa for Mesozoic metasediments (Lupták *et al.* 2003). Rocks in the hanging wall of the detachment fault, i.e. extensional allochthons of the Gemeric and Silicic superunits, are only anchizonal to unmetamorphosed. The cataclastic shear zone along their base was formed under considerable fluid overpressure (Milovský *et al.* 2003).

The detachment fault at the top of the Veporic Superunit was active during the Late Cretaceous unroofing of the Veporic metamorphic core complex as suggested by $^{40}Ar/^{39}Ar$ data (Maluski *et al.* 1993; Dallmeyer *et al.* 1996; Kováčik *et al.* 1997; Janák *et al.* 2001), which record cooling ages of various mineral phases between 110 and 75 Ma (Fig. 18.32). The extension and exhumation of the southern part of the Veporicum culminated in the intrusion of the Rochovce Granite at 81 to 76 Ma (Hraško *et al.* 1999; Poller *et al.* 2001). Apatite fission-track ages from Variscan granitoids suggest that cooling took place from the Late Cretaceous to the Palaeogene (Král' 1977).

Orogen-parallel extension was accompanied, and followed, by orogen-normal contraction resulting in superimposed deformation stages (Plašienka 1993; Plašienka *et al.* 1999). The $D_2$ structural association is dominated by steep cleavage and related tight upright folds that are associated with several SW–NE to

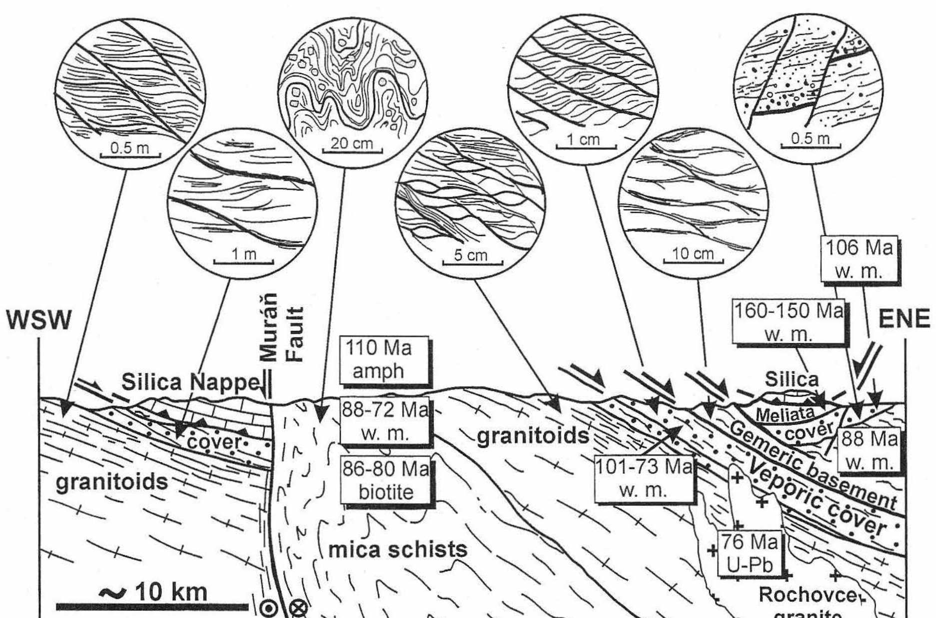



**Fig. 18.32.** Geological cross-section across the eastern margin of the Veporicum parallel to the regional trend of the stretching lineation (modified after Plašienka *et al.* 1999*b*). Typical structures related to east-vergent unroofing and important radiometric (mostly $^{40}Ar/^{39}Ar$) data are shown.

WSW–ENE trending high-strain zones. These are interpreted as deep-seated sinistral transpressional belts (e.g. Lexa *et al.* 2003). They dissect the exposed part of the Veporicum in central Slovakia into individual zones. The Osrblie Fault System separates the Ľubietová Zone from the Kraklová Zone, while the Pohorelá Fault divides the Kraklová and the Král'ova Hol'a zones (Hók & Hraško 1990; Putiš 1991; Madarás *et al.* 1994), and the Muráň Fault (Marko 1993) separates the Král'ova Hol'a Zone from the Kohút Zone. All of these fault systems, together with the Čertovica Fault (e.g. Siegl 1978) which separates the Veporic and Tatric superunits, and the Lubeník Fault at the boundary between the Veporic and Gemeric superunits, are latest Cretaceous–earliest Palaeogene in age. Outcrop data suggest that there was no reactivation of these faults in Neogene times. The same is true for conjugate transversal fault systems with a NW–SE direction (e.g. the Mýto-Tisovec Fault; Marko & Vojtko 2006).

*Gemeric Superunit*

The Gemeric Superunit (or Gemericum) is the uppermost basement-cover superunit of the Central Western Carpathians. It is volumetrically smaller than the Tatric and the Veporic superunits and, moreover, it wedges out laterally. Partial analogues of the Gemeric Superunit can be found in the Upper Austro-Alpine units of the Eastern Alps which include Palaeozoic formations, such as the Greywacke Zone (Noric nappe), the Graz Palaeozoic, the Gurktal nappes, and some units in the basements of the Styrian and Danube basins (e.g. Mihályi Ridge; Neubauer & Vozárová 1990; Balla 1994; Ebner 1992; Tari 1995).

During the Variscan cycle, the Gemeric Superunit represented the southern, external parts of the orogen, whereas during the Alpine cycle it formed the innermost zone of the Central Western Carpathians. Palaeogeographically, it represents the southern marginal zone of the Slovako-Carpathian system, abutting the Meliata Ocean which was located to the south. Following closure of the ocean, the Gemericum was thrust northward over the Veporic Superunit. According to data from the G1 deep-seismic profile (Vozár *et al.* 1998), the Gemeric sheet is a wedge-shaped, southward-thickening upper crustal body in its central part. Its lower limiting plane, dipping at *c.* 20°, can be traced downward along the Lubeník-Margecany Fault which merges with the steeply dipping Rožňava Fault Zone at at *c.* 15 km depth (Fig. 18.20).

The Gemeric Superunit comprises a series of northward-verging thrust imbricates and partial nappes, the number and terminology of which varies from author to author. Most frequently, the North Gemeric (Klátov, Rakovec, Črmel' and Ochtiná units) and South Gemeric (Volovec and Štós units) units are distinguished.

The North Gemeric basement is represented by the Rakovec and Klátov units with an oceanic affinity. The **Rakovec Unit** (phyllite–diabase complex) comprises low-grade metasediments and basic volcanics, of probable Devonian age. Together with the higher-metamorphosed **Klátov Unit** (gneiss–amphibolite complex) it probably represents a Variscan oceanic suture.

The North Gemeric Mississippian cover is represented by the **Ochtiná and Črmel' units** (the former to the west, the latter to the east; both are *c.* 1000 m thick). They comprise low-grade metamorphosed marine clastics and metasomatized carbonates (magnesites), together with basic volcanics and, in places, ultrabasics. Their age is Late Visean–Serpukhovian (Vozárová 1996). The Pennsylvanian Dobšiná Group lies transgressively on the Lower Palaeozoic substratum in the northern part of the Gemeric area. Basal clastics are overlain by shallow-marine fossiliferous carbonates, basic volcanics and a regressive paralic succession. The Permian Krompachy Group rests transgressively on various North Gemeric complexes. Its basal part consists of unsorted coarse clastics deposited in a terrestrial environment. These pass up into Upper Permian and Lower Triassic lagoonal-sabkha sediments, accompanied by subalkaline rhyolitic volcanism. These sediments are tectonically overlain by carbonate complexes considered to be part of the Silicic Superunit (Stratená Nappe; Mello 2000). Previously, the carbonates were assumed to be part of the cover of the Gemericum ('North-Gemeric Mesozoic'). Indeed, the tectonic contact cannot be convincingly documented in many places.

The largest unit of the Gemericum is the **Volovec Unit**, comprising low-grade metamorphosed Lower Palaeozoic volcanosedimentary complexes (Gelnica Group, several thousand metres thick), intruded by the Late Variscan (Permian) Spiš-Gemer granitoids (Finger & Broska 1999; Poller *et al.* 2002). The sedimentary units consist of upward-thickening deep-marine clastic megacycles containing abundant terrigenous and volcanogenic material. The bases of the megacycles comprise pelagic chert (lydites), anoxic shales, and locally carbonates (e.g. Grecula 1973, 1982; Ivanička *et al.* 1989). The age of the Gelnica Group ranges from Cambrian to Early Devonian. The volcanics (mostly volcaniclastics) are acidic to intermediate, and rarely basic. The **Štós Unit** is the uppermost imbricate of the Gemeric Superunit, composed of monotonous phyllites of uncertain, but possibly Mississippian age. The sedimentary cover of the Volovec and Štós units comprises the Permian–Scythian clastic Gočaltovo Group, including the continental, Verrucano-type Lower Permian Rožňava Formation and the Upper Permian–Lower Triassic, partly shallow-marine Štítnik Formation (Reichwalder 1973; Vozárová & Vozár 1988; Vozárová 1996). Some erosional remnants of Middle Triassic carbonates locally occur below the Meliatic Bôrka Nappe and are also probably part of the Gemeric sedimentary cover (Mello 1997).

The internal structure of the Gemeric Superunit formed as a result of the interaction of Variscan and Alpine deformation. Distinguishing between these two deformation events is a matter of much controversy (e.g. Reichwalder 1973; Ivanička *et al.* 1989; Grecula 1982; Jacko *et al.* 1996; Németh *et al.* 1997; Lexa *et al.* 2003). The most conspicuous feature of the Gemeric structure is the presence of a tight, isoclinally folded and imbricated fabric associated with a large-scale cleavage fan transversally cut by transpressional shear zones (Grecula *et al.* 1990) which were possibly caused by the indentation of a still-unrecognized basement block from the south (Lexa *et al.* 2003).

*Fatric Superunit*

The term Fatric Superunit, or Fatricum, is used to refer to units composed mainly of detached sedimentary cover, which were derived from zones between the present-day Tatric Superunit in the north and the Veporic Superunit in the south (Fig. 18.33) (i.e. this term is used both in a tectonic and a palaeogeographic sense). The term was introduced by Andrusov *et al.* (1973) for the 'Lower Sub-Tatra Nappe' (i.e. the Krížna and analogous units). However, the Fatricum includes not only the detached complexes of the Krížna nappe proper, but also their original (mostly underthrust) basement, although this basement is only rarely exposed at the surface.

Three types of tectonic units occur in the Fatric Superunit: (1) units with basement and cover; (2) cover nappes (Krížna Unit); and (3) detached units in the Peri-Klippen Zone. The first type is represented by units comprising pre-Alpine basement and its Permian–Scythian sedimentary cover, which form duplexes in

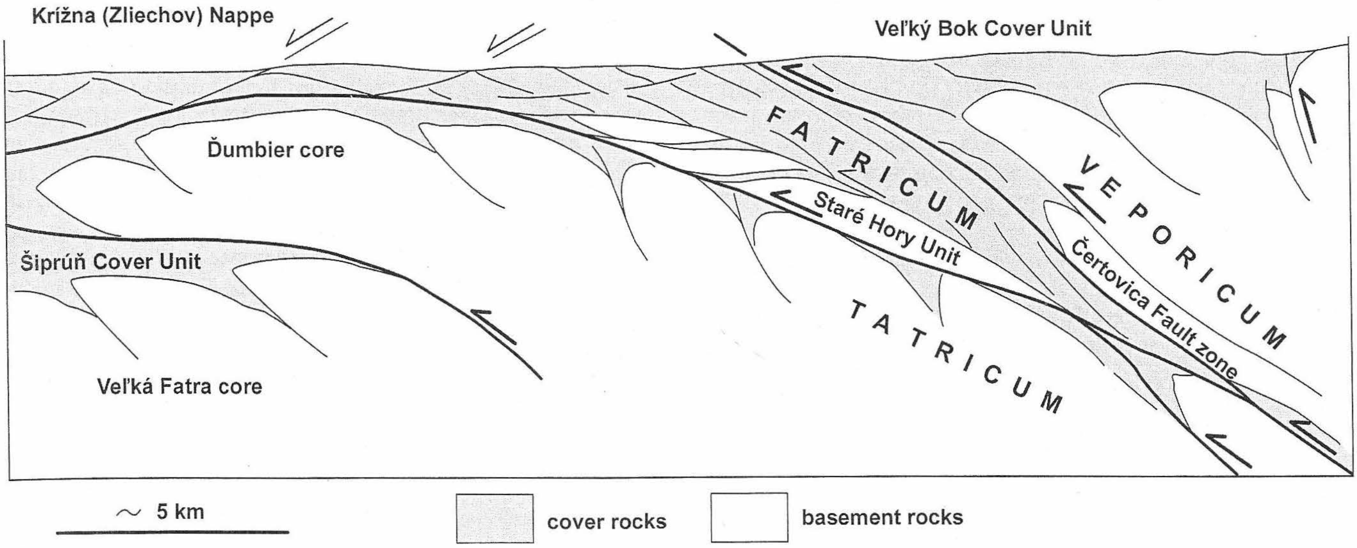



**Fig. 18.33.** Schematic section through the rear Fatric elements showing their relationships to the underlying Tatric and overlying Veporic units (after Plašienka 1999*b*).

the internal part of the Fatric nappe system (Jaroš 1971; Hók *et al.* 1994; Plašienka 2003*b*). The **Rázdiel Unit** in the Tribeč Mountains includes pre-Alpine basement composed of phyllites, mica schists, amphibolites, and mylonitized granitoids. The cover consists of Late Permian coarse-clastic and shaly sediments. An analogous, imbricated basement duplex (**Staré Hory Unit**, Fig. 18.33) is found in the Starohorské vrchy Mountains (western part of the Nízke Tatry Mountains) and contains an orthogneiss basement overlain by Permian and Triassic sediments. In the **Smrekovica Unit** of the Branisko Mountains, the basement comprises a pre-Alpine, high-grade metamorphosed crystalline complex, covered by Permian and Scythian clastics.

The second type of tectonic units is typified by the **Krížna Unit**, which is one of the fundamental units of the Tatra-Fatra Belt. This unit is composed of several detached partial nappes containing mainly Middle Triassic–mid-Cretaceous sedimentary successions of several lithotectonic types (e.g. Andrusov 1968; Mahel' 1983, 1986). These are in an allochthonous position above the Tatric cover and are tectonically overlain by the Hronic nappes (Fig. 18.33). The nappe sole is formed by incompetent Upper Scythian shales and evaporites and by carbonatic tectonic breccias (cargneule) (Jaroszewski 1982; Plašienka & Soták 1996). The overlying Middle Triassic carbonates form the rigid, mechanically dominating element of the nappe. As in the Tatric Superunit, the lower Middle Triassic strata are represented by the Gutenstein Formation, overlain by sabkha-type Ramsau dolomites. However, unlike the Tatric Superunit, the upper part of the Ladinian–Carnian dolomite complex includes a thin package of quartz sandstones and shales (Lunz Formation). The Norian-age Carpathian Keuper Formation (up to 300 m) contains fewer clastics than the stratigraphic equivalent in the Tatric Superunit and is composed of claystones, siltstones, rare sandstones, and in places gypsum, with the upper part dominated by evaporitic dolomites deposited in supratidal lagoons (Al-Jaboury & Ďurovič 1996). Secondary detachment zones were generated along this incompetent horizon. The Kössen Formation comprises shallow-marine fossiliferous limestones and reflects the Rhaetian marine transgression as well as the onset of rifting of the European shelf (Michalík 1977, 1978; Tomašových 2000).

The Jurassic–Lower Cretaceous deposits are found in two distinct lithotectonic units showing differing facies development. Substantial parts of the Krížna Unit (Krížna Nappe *sensu stricto* = Zliechov partial unit) comprise the deep-water ***Zliechov succession***, up to 2000 m thick, which begins with the Hettangian–Sinemurian, terrigenous, littoral to neritic Kopienec Formation. This passes up into the synrift, hemipelagic Allgäu Formation represented by bioturbated marlstones and limestones ('Fleckenmergel', up to 500 m thick). In marginal areas to the south (***Il'anovo succession***), originally close to the North Veporic Vel'ký Bok Unit, the Kopienec and Allgäu formations are replaced by swell facies of oolitic, crinoidal and nodular limestones (Hierlatz and Adnet formations). Nodular and marly limestones represent the entire Middle and Upper Jurassic. In the Zliechov succession, the Middle to Late Jurassic sediments reflect post-rift thermal subsidence. They are pelagic cherty and siliceous limestones, overlain by Oxfordian-age radiolarites (Ždiar Formation; Polák *et al.* 1998). During the Middle and Late Malm, siliceous, cherty and marly limestones were deposited in the axial zones of the Zliechov Basin (Jasenina and Osnica formations). The proportion of terrigenous clayey material increased during the Early Cretaceous (Mráznica and Párnica formations; e.g. Borza *et al.* 1980; Michalík *et al.* 1993*a*; Reháková 2000). Allodapic calciturbidite and slump breccia intercalations occur in several horizons (Michalík & Reháková 1995; Michalík *et al.* 1996). The Párnica Formation also includes numerous, though small bodies of submarine basanitic lavas (Hovorka & Spišiak 1988, 1993; Spišiak & Hovorka 1997) and olistostromes (Jablonský & Marschalko 1992). Sedimentation in all of the Krížna units terminated with the deposition of the mid-Cretaceous (Albian–Cenomanian), synorogenic, turbiditic Poruba Formation (Jablonský 1978). Following deposition of the Poruba Formation, in the Late Turonian, the main phase of nappe thrusting took place.

While the Zliechov succession builds up a substantial part of the Krížna Unit (the Krížna nappe *sensu stricto*), the ***Vysoká succession*** forms several smaller, partially independent subunits situated at the front and sole of the Krížna nappe *sensu stricto* (i.e. Vysoká, Beckov, Belá, Ďurčiná and Havran partial nappes). The Vysoká succession is characterized by the dominance of Jurassic sediments deposited in comparatively shallow water. Liassic strata are represented by bioclastic (crinoidal) and sandy cherty limestones, while the Dogger and Malm comprise massive crinoidal limestones (Vils Formation), as well as cherty, siliceous and nodular limestones (Koša 1998). The Lower Cretaceous deposits are similar to those of the Zliechov succession but allodapic limestones are more frequent, especially in the Barremian and Aptian (allodapic bioclastic limestones derived from adjacent Urgonian carbonate platforms; Michalík & Soták 1990; Michalík 1994*b*).

The third type of Fatric unit includes detached units containing mostly Lower Jurassic–Middle Cretaceous sedimentary successions, located in the Peri-Klippen Zone of the Pieniny Klippen Belt. These also include Upper Cretaceous, post-overthrust sedimentary sequences. The Drietoma, Manín, Klape and Haligovce units and, perhaps, some other 'non-Oravic' units of the Pieniny Klippen Belt belong to this subdivision.

The evolutionary tectonic model of the Krížna Unit (Plašienka & Prokešová 1996; Plašienka 1995*c*, 1999*b*, 2003*b*) assumes that the Zliechov partial unit (Krížna Nappe *sensu stricto*) was formed by underthrusting of an extensive basin floored by continental crust which had been strongly stretched and thinned during Early Jurassic rifting. The lithological variability of the cover in the Krížna Unit resulted in a mechanical stratification of the nappes. They contain three important décollement horizons: Upper Scythian, Keuper, and the base of the Poruba Formation. The first décollement horizon is the most important, since almost all of the nappes of the Krížna Unit were detached along it. The importance of the other two increases towards the frontal parts of the Krížna Unit. Cover units below the Upper Scythian décollement remained mostly attached to the underlying basement (Jaroš 1971). Massive Triassic carbonates form the relatively rigid basal part, and the well-bedded Jurassic to Lower Cretaceous sequences the incompetent middle part of the nappes. The uppermost Poruba Formation was the most incompetent; it is mostly absent in the rear and accumulated in the frontal parts of the nappes.

The structural evolution of the Fatric Superunit and the closely related, southerly adjacent Vel'ký Bok Unit (Veporic Superunit) is subdivided into several deformational stages, the first two of which occurred during the generation and emplacement of the Fatric Superunit. The Vel'ký Bok Unit (Fig. 18.33) displays complex structures formed under low- to very-low-grade metamorphic conditions (Plašienka 1995*c*, 2003*b*). The first stage involved the formation of bedding-parallel foliation, stretching lineation and flow folds in calcite-rich rocks. Macrostructures of this stage include large-scale recumbent to northward-plunging folds, all with top-to-the-north to NW kinematics. These structures originated at the northern tip of the Veporic Superunit overriding the underthrust Fatric basement. The second deformation stage involved the development of crenulation cleavage and upright folds during the collision of the Veporic Superunit with the northern, Tatric margin of the Zliechov Basin. In the Krížna Unit, these two stages are represented by compressional structures, such as intrastratal detachment zones, mesofolds, and imbricated duplexes formed during initial décollement and shortening of the Zliechov Basin fill. These contractional structures were passively transported during the gravitationally driven final emplacement of the Krížna Unit. Structures related to this gliding event are extensional and presumably formed synchronously with the second (shortening) deformation stage in the Vel'ký Bok Unit.

The inversion of the Zliechov Basin and the generation of the Krížna Unit occurred in mid-Cretaceous times. The Zliechov Basin was progressively shortened through underthrusting of its basement complexes below the Veporic thrust wedge (Plašienka 2003*b*). The sedimentary fill was detached along the lower décollement horizon and formed an initial fold-and-thrust stack prograding outwards. Simultaneously, deep-marine clastic prisms (Poruba Formation, Klape Unit) fed by rising hinterland units were deposited in piggyback basins as well as in the foreland (Tatric Superunit). Following the complete closure of the Zliechov Basin by southward underthrusting of its basement, the Tatric and Veporic basin margins collided and the detached Krížna Nappe *sensu stricto* was pushed over the frontal South Tatric ramp, from which the frontal Fatric elements (Vysoká- and Manín-type), with slope- and ridge-related sedimentary successions, were detached. In the late Turonian, the nappes of the Fatric Superunit moved northwards under the force of gravity from an uplifted area in the southern part of the Tatricum and over the basinal areas in the northern Tatricum. The Klape Unit, composed of a mid-Cretaceous deep-marine clastic complex, was the lowermost and most frontal sheet. Some frontal, mostly Manín- and Vysoká-type units, were emplaced above it as the second sheet. The Klape and Manín units glided furthest, and were emplaced on the (later-subducted) oceanic crust of the Vahic domain north of the Tatric Superunit. In the overlying main sheet, which is the Krížna nappe *sensu stricto* with its dominant basinal Zliechov succession, the emplacement event is recorded by extensional structures superimposed on older, compressional structures (Prokešová 1994). Finally, the frontal elements of the Veporic basement wedge overrode the southern Tatric basement and cover (the former northern margin of the Zliechov Basin), along with local imbricated basement duplexes detached from the original substratum of the Zliechov Basin (Plašienka 2003*b*).

*Hronic Superunit*

The Hronic Superunit (or Hronicum) represents the structurally highest nappe system in the Tatra-Fatra Belt where it forms numerous nappe outliers resting on the Fatric units. Hronic units also occur in the northern part of the Malé Karpaty Mountains (Veterlín, Havranica, Jablonica and Nedzov nappes). Upper Austro-Alpine nappes in the Northern Calcareous Alps (Upper Bajuvaric and Tirolic units: Lunz, Göll and Ötscher nappes), and beneath the fill of the Vienna Basin, are analogous to the Hronicum. In the Northern Calcareous Alps, however, the typical Hronic member, the Upper Palaeozoic Ipoltica Group, is absent.

The Hronic Superunit is a large system of unmetamorphosed sedimentary cover nappes which originated in the southern zones of the Slovako-Carpathian system, i.e. on the northern passive margin of the Meliata Ocean. They were detached at the base of the Upper Palaeozoic detrital succession, but in the northern parts of the Central Western Carpathians the Triassic carbonate complex is detached along the horizon of Upper Scythian shales and marlstones. The nappes underwent exclusively brittle deformation and represent thin-skinned, décollement cover nappes. Originally, during the Cretaceous, they formed thin tabular bodies subhorizontally covering (probably) the entire Tatra-Fatra Belt. Internally, they are not very deformed, but the portions containing rigid reef bodies usually form independent partial nappes (e.g. the Strážov nappe; Kováč & Havrila 1998). Along the northern margin of the Central Western Carpathians, in the vicinity of the Pieniny Klippen Belt, the Hronic nappes were largely affected by superimposed transpressional and transtensional deformation together with their syn- to post-tectonic sedimentary cover (Gosau Supergroup).

The basal part of the Hronic nappe system in the southern zones of the Tatra-Fatra Belt is represented by the thick (several thousand metres) Upper Palaeozoic–Scythian Ipoltica Group.

The Upper Pennsylvanian Nižná Boca Formation consists of a regressive lacustrine-deltaic succession including sandstones, conglomerates, sandy shales, and dacitic volcaniclastics (50–500 m thick). The overlying, synrift, fluvial-lacustrine and alluvial Malužiná Formation includes extensive basaltic volcanism and probably represents the entire Permian. This unit is up to 2000 m thick. Synsedimentary basaltic and andesitic volcanics, forming extensive lava flows, were generated during two major eruption phases (Vozár 1997; Dostal *et al.* 2003). Chemically, the basalts represent continental tholeiites suggesting a continental rift environment. The Lower Triassic strata comprise Lower Scythian quartz sandstones and Upper Scythian shales and sandstones, alternating with marlstones and sandy limestones in the upper part.

The Middle Triassic succession is comparatively thick and includes a wide range of sedimentary facies, representing various parts of the shelf environment, from tidal flats and reef platforms up to pelagic intrashelf basins. There were several subsiding and elevated zones within the Hronic sedimentary area during the Middle Triassic (e.g. Dobrá Voda Basin, Nedzov-Strážov Platform, Biely Váh Basin, Čierny Váh Platform; Kováč & Havrila 1998), which were later inverted to create numerous partial units within the Hronic Superunit. The Anisian-age carbonate ramp (Gutenstein and Annaberg formations) with restricted platforms (Steinalm Formation) was partly destroyed during the Pelsonian rifting event. This event ultimately led to the formation of a series of subsiding pelagic basins (Reifling Formation; e.g. Masaryk *et al.* 1993) rimmed by carbonate clastic aprons where the sediments were derived from adjoining, reef-cored prograding platforms (e.g. Wetterstein Formation, up to 1000 m thick; Michalík *et al.* 1993*b*; Polák *et al.* 1996). During the middle Carnian, the intrashelf basins were completely filled with up to 600 m of shales and siliciclastic turbidites (Lunz Formation). In the Norian and Rhaetian, carbonate platform conditions were re-established (Hauptdolomit and Dachstein formations). Locally, Rhaetian fossiliferous limestones infilled channels in tidal flats.

Two successions have been distinguished within the Middle and Upper Triassic part of the Hronic Superunit: the Čierny Váh and the Biely Váh successions. The ***Čierny Váh succession*** is characterized by a predominance of ramp and platform carbonates, now mainly dolomites, while the ***Biely Váh succession*** is more diverse, with predominantly basinal facies from late Anisian to early Carnian and shallow-water facies during the late Carnian to Rhaetian. This latter succession represents the fill of intrashelf depressions which originated due to late Anisian rifting (Michalík 1994*a*), probably related to the opening of the Meliata Ocean. The Middle to Upper Triassic carbonate platform complexes attain a thickness of 2000–3000 m.

In Early Liassic times, a hiatus has been noted in the Hronic Superunit and this was followed by the deposition of the Upper Lias and Dogger successions which are partly condensed (e.g. Hierlatz, Adnet, Klaus and Vils formations). The Upper Jurassic stage is represented by pelagic basinal limestones (Oberalm Formation) with intercalations of allodapic calciturbidites (Barmstein Formation; Mišík & Sýkora 1982). During the Early Cretaceous, a pelagic marly and cherty limestone succession was deposited, and this was overlain by the Hauterivian siliciclastic turbiditic Schrambach and Rossfeld formations (Michalík *et al.* 1996). These are the youngest sediments in the Hronic Superunit.

The Hronic nappes are typical 'rootless' nappes with no connection to their original basement. Circumstantial evidence suggests that this basement corresponded to the Veporic or an analogous palaeogeographic zone. Nevertheless, structural studies suggest that the Hronic nappes were not derived from the presently exposed part of the Veporic basement nor from the Lubeník-Margecany Fault (Plašienka & Soták 2001).

*Silicic Superunit*

The Silicic Superunit (Silicicum) includes the structurally highest, unmetamorphosed nappes of the region. They are restricted to the Vepor-Gemer Belt of the Central Western Carpathians and the Slovak-Aggtelek Karst area at the boundary between the Central Western Carpathians and the Internal Western Carpathians (i.e. Drienok, Muráň, Vernár, Stratená and Silica-Aggtelek nappes, Szőlősardó and Bódva nappes in the Aggtelek Karst and Rudabánya Mountains of northern Hungary). Based on their lithostratigraphy, the Silicic nappes can be correlated with the Upper Tirolic and/or Juvavic nappes of the Northern Calcareous Alps. The palinspastic position of the Silicic Superunit is uncertain. The facies relationships support a proposed original position along the northern passive margin of the Meliata Ocean. However, the structural position of the Silicic Superunit at the top of the nappe stack would suggest an origin from the southern margin of the Meliata Ocean (see discussion in Frisch & Gawlick 2003; Mello 1997).

The Silicic units form internally little-deformed thrust nappes, detached at the basis of a thick Triassic carbonate platform complex, usually located along the Upper Permian–Lower Scythian evaporitic horizon. In places, however, slivers of Meliatic oceanic rocks were found at the base of the Silicic nappes, incorporated into evaporite mélanges (Réti 1985; Havrila & Ožvoldová 1996; Horváth 2000; Vojtko 2000). In the Slovak Karst and Slovenský raj (Stratenská hornatina, Galmus) areas, the Silicic nappes were dissected by later transpressional movements along deep-seated wrench faults (e.g. the Rožňava and Muráň faults; Mello 1997, 2000).

The Silicic units contain sedimentary complexes of Late Permian to Late Jurassic age and dominated by extensive Middle to Upper Triassic carbonate platforms. The soles of the Silicic nappes are formed by the Upper Permian, evaporite-bearing Perkupa Formation, overlain by Lower Triassic sandstones and shales (Bódvaszilas Formation) and marlstones to limestones (Szin Formation), with a total thickness of 400–800 m. Rhyolite volcanics occur in the latter formation in the northern parts of the Silicic nappe system. The Anisian carbonate ramp (Gutenstein Formation) and platform (Steinalm Formation) are *c.* 500–600 m thick and were partly destroyed by a marked Pelsonian rifting event, which led to the formation of intrashelf basins filled with hemipelagic nodular limestones and resedimented carbonates (Schreyeralm, Reifling, Nádaska, Raming formations). Carnian-age siliciclastic turbidites also occur locally (Lunz and Reingraben formations). Thin intercalations of altered tuffites are known from the Ladinian succession. However, thick Middle–Upper Triassic carbonate platform complexes with extensive, prograding, 1000–2000 m thick reef bodies surrounded by peri-reef bioclastic aprons (Steinalm, Wetterstein, Tisovec-Waxeneck, Furmanec formations) predominate in the Silicic units. Barrier reefs grade into backreef lagoonal flats (Dachstein, Hauptdolomit, Bleskový Prameň formations; e.g. Mello 1997, 2000). Toward the south (Slovak-Aggtelek Karst), Middle Triassic platform carbonates are overlain by Upper Triassic basinal and slope pelagic limestones and marlstones (Aflenz, Pötschen, Hallstatt, Zlambach formations). The southernmost units, presumably belonging to the Silicic system (Szőlősardó and Bódva nappes in northern Hungary), are already marked by the prevalence of pelagic facies already in the Middle Triassic (subsequent to the Pelsonian rift event), such as the Ladinian Hallstatt-type Bódvalenke limestone and cherts deposited near the CCD (Szárhegy Formation). This facies indicates an area of transition to the Meliata Ocean (e.g. Kovács 1992).

Following the earliest Jurassic hiatus, limestones of the Hierlatz and Adnet formations were deposited on basin highs on top of the Upper Triassic platform sediments. In basinal areas, hemipelagic sediments of the Allgäu Formation were deposited (Rakús 1996; Rakús & Sýkora 2001; Mello 1997). Middle to Late Jurassic subsidence led to the deposition of deep-water shales and radiolarites with some terrigenous input. In the Slovak Karst area, bodies of chaotic breccias (olistostromes) are also found (Sýkora & Ožvoldová 1996). The youngest dated sediments of the Silicic Superunit are Oxfordian radiolarites. The Jurassic members are, however, poorly preserved; their overall thickness is only several tens of metres. Shallow-water Upper Jurassic limestones are found as clasts in Senonian and Tertiary conglomerates.

Relicts of Senonian-age Gosau Group sediments, related to the Silicic nappes, occur in several locations (e.g. Poniky, Šumiac, Dobšiná Ice Cave, Gombasek, Miglinc), and are preserved along younger fault structures. They comprise freshwater limestones, conglomerates, rudist limestones, pelagic marlstones (Campanian) and the clayey infill of palaeokarst cavities (Mello 1997, 2000).

*Meliatic Superunit*

Tectonic units belonging to the Meliatic Superunit (Meliaticum) represent the structurally deepest elements of the Slovak Karst Mountains. Parts of the Meliaticum which were thrust northward over the Gemeric basement form the Jaklovce Unit (unmetamorphosed ophiolite mélange; Mock *et al.* 1998), and the low-temperature, high-pressure (12 kbar) metamorphosed Bôrka Unit. The latter is a transitional Gemeric–Meliatic element (Faryad 1995*a*, *b*, 1997, 1999; Mello *et al.* 1998). The Meliata Unit *sensu stricto*. (Jurassic deep-water clastic complexes with radiolarites, olistostromes, mélanges and ophiolitic bodies; Kozur & Mock 1973, 1995, 1997; Kozur 1991; Mello 1997; Mock *et al.* 1998) crops out in tectonic windows from below the Tornaic and Silicic nappes in the western parts of the Slovak Karst area (Figs 18.19, 18.29 & 18.34). The most distinctive feature of the Meliata Superunit is the presence of slivers and olistoliths of ophiolites, blueschists and Triassic sediments. These latter include Lower Anisian platform carbonates showing evidence of a Pelsonian unconformity, and Upper Anisian and younger pelagic deposits (Upper Anisian red cherty limestones, Ladinian and Upper Triassic radiolarites, Norian nodular limestones etc.).

The original assumption of a Triassic Meliata Ocean was based on these pelagic Triassic strata which were thought to represent a continuous stratigraphic succession. Although the olistolith character of the Triassic sediments was later recognized, they still provide evidence, along with other features, for the existence of a Triassic oceanic realm that persisted up to Late Jurassic times. Hence, the Meliata Ocean in the Western Carpathian realm is assumed to have existed from the Middle Triassic to the early Late Jurassic.

Based on sparse biostratigraphic data, the deformation of the Meliatic units occurred after the closure of the Meliatic Ocean. This occurred before the Kimmeridgian (Kozur 1991; Rakús 1996). Blueschist-facies metamorphic basalts from the base of the Meliatic accretionary complex (Bôrka Nappe), overridding the southern, Gemeric, margin of the Slovako-Carpathian system, yielded Late Jurassic ages for the high-pressure/low-temperature metamorphism (150–160 Ma, $^{40}Ar/^{39}Ar$ on phengites; Maluski *et al.* 1993; Dallmeyer *et al.* 1996; Faryad & Henjes-Kunst 1997).

## Central Carpathian Palaeogene Basin

The Central Carpathian Palaeogene Basin (= Podhale Basin in Poland) most likely originally covered the entire Tatra-Fatra Belt as well as much of the Vepor Sub-belt in the Eocene and Oligocene. However, in western Slovakia only erosional remnants of its sediments are preserved in small basins between the 'core mountains'. In northern and northeastern Slovakia, the basin sediments (Podtatra Group) are more widespread, 3000–5000 m thick and subdivided into four formations (e.g. Soták *et al.* 2001; Fig. 18.29). The basal, transgressive Borové Formation consists of carbonate conglomerates and nummulitic limestones, as well as extensive prisms of the Súl'ov dolomitic breccia in areas close to the Pieniny Klippen Belt. The age of the Borové Formation decreases from the north (Pieniny Klippen Belt) towards the south and from the west (Early to Middle Eocene in the Malé Karpaty Mountains) to the east (Late Eocene to Early Oligocene in the Levoča Mountains). This reflects the direction of the Eocene transgression. The overlying Huty Formation is composed of shales and distal turbidites, locally with conglomeratic slump bodies. The siliciclastic turbiditic Zuberec Formation forms much of the basin fill. Thick bodies of massive, amalgamated sandstones (Biely Potok Formation) form the highest member (Egerian).

The Senonian and Palaeogene basins (Gosau Supergroup basins, Central Carpathian Palaeogene Basin) in the northern part of the Central Western Carpathians originated in a forearc position at the outer edge of the Central Western Carpathians. Their subsidence may have been caused by tectonic erosion during subduction of the South Penninic (Vahic) Ocean and later of the North Penninic (Magura) Ocean (Wagreich 1995; Kázmér *et al.* 2003).

## Palaeotectonic evolution of the Central Western Carpathians

The Central Western Carpathians represent a complex tectonic system that evolved from Late Palaeozoic times onward. This evolution can be summarized as follows (Plašienka 1995*a*, 2003*a*; Fig. 18.22 & 18.23).

1. Late Carboniferous to Permian: collapse and erosion of the Variscan Orogen accompanied by the formation of several grabens filled with immature continental red-beds; alkaline to calc-alkaline rift-related volcanism and plutonism, persisting until the Early Triassic.
2. Triassic: period of gradual subsidence; marked Late Anisian rifting related to breakup of the Meliata Ocean.
3. Early to Middle Jurassic: rifting proceeding in three phases; destruction of the Triassic carbonate platform; subsequent thermal subsidence during Middle and Late Jurassic.
4. Late Jurassic: closure of the Meliata Ocean; formation of a collisional belt in the southern zones of the Central Western Carpathians.
5. Early Cretaceous: thrusting (Gemericum over Veporicum) in zones adjoining the Meliata suture; extension and subsidence in foreland zones.
6. Mid-Cretaceous: northward (in present-day co-ordinates) progradation of crustal shortening (Veporicum over Tatricum); detachment and emplacement of cover nappe systems (Fatricum and Hronicum).

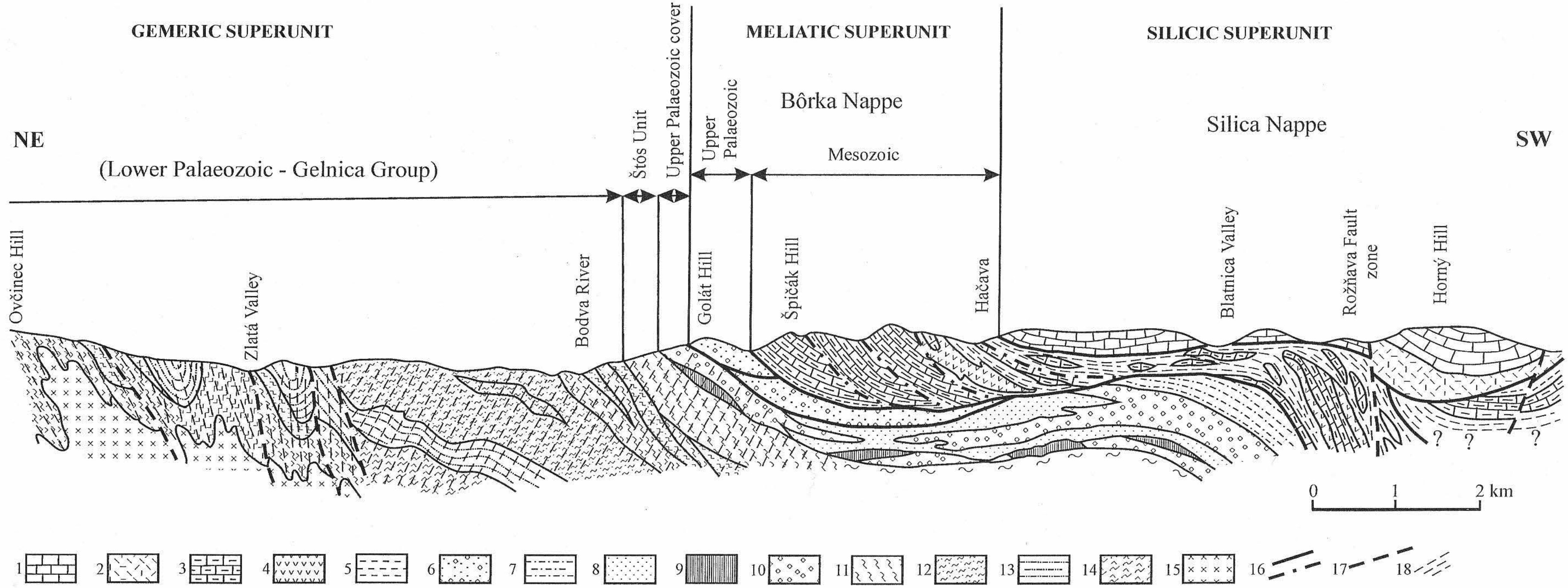



**Fig. 18.34.** Cross-section of the Slovak Karst Mountains and southern part of the Gemeric Superunit (modified after Reichwalder 1982). Note the cleavage fan in the Gemericum, imbrication in the Meliatic Bôrka Nappe and structural discordance at the base of the Silica Nappe above the steepened part of the Meliaticum within the Rožňava Fault Zone (suture?). Legend: Silica Nappe: 1, Triassic carbonates; 2, Scythian marls, shales and sandstones. Meliaticum: 3, crystalline limestone; 4, glaucophanized basalts; 5, phyllite, slates (Jurassic?); 6, metasandstones, metaconglomerate (Permian?). Gemericum, Permian cover: 7–10, conglomerates, sandstones, shales, rhyolites. Gemericum, Lower Palaeozoic metavolcanosedimentary complexes (Gelnica Group): 11–15, phyllites, metasandstones, porphyroids. Structures: 16, major overthrusts, 17, steep faults, 18, cleavage.

7. Senonian: extensional collapse of the southern zones of the Central Western Carpathians; exhumation of the Veporic metamorphic core complex; emplacement of the Silicic cover nappes.
8. Early Palaeogene: renewed shortening; sinistral transpression.
9. Eocene to Oligocene: extensional collapse of the northern zones of the Central Western Carpathians (Eocene) and subsidence of the Central Carpathian Palaeogene Basin.
10. Early Miocene: crustal shortening.
11. Middle Miocene: backarc extension leading to formation of the Pannonian Basin System and widespread calc-alkaline volcanism.
12. Late Miocene to Pliocene: gradual slowing of the rate of subsidence in the various basins of the region; uplift of 'core mountains'.

## *Internal Western Carpathians*

The term Inner (Mahel' 1986) or Internal Western Carpathians (Plašienka 1999*a*) is used for the units extending south of the inferred suture resulting from the closure of the Meliata Ocean. These units have many features in common with units of the Southern Alps and/or the Dinarides. The position of the suture of the Meliata Ocean plays a key role in the separation of the Internal Western Carpathians from the Central Western Carpathians. Since part of the Meliatic Superunit was emplaced on the Gemericum as a north-vergent nappe (Bôrka Unit), the main suture should be placed immediately south of the Gemericum. This area is designated the 'Rožňava Suture' (Rožňava-Šugov according to Kozur & Mock 1997). An almost identical structure was termed by Grecula (1973) the 'Carpathian-Pannonian Suture'. This author assumed that the Gemericum was expelled from it during the Alpine (Cretaceous) Orogeny; the Rožňava Fault Zone represents its surficial expression. The Rožňava Suture is, however, largely covered by the Tornaic and Silicic nappes in the Slovak Karst Mountains (Fig. 18.34) and covered by unconformably overlying Tertiary complexes elsewhere. The Rožňava Suture can be tentatively extended towards the west along the geophysically defined, deep-seated Plešivec, Diósjenő, Hurbanovo and Rába faults (Plašienka *et al.* 1997).

The Rožňava Suture marks the boundary between the Central and Inner Western Carpathians. The rock complexes within the Rožňava Suture are mainly composed of sheared-off slivers of oceanic sediments and dismembered ophiolites occurring within an ancient accretionary complex which separates crustal blocks of varying development, composition and structure. However, this boundary between the Central and Internal Western Carpathians is a conceptual one and cannot be drawn precisely, due to the many uncertainties in the exact position and lateral prolongation of the Meliatic Superunit and related units.

Regionally, the Internal Western Carpathians include the Slovak Karst Mountains and the north Hungarian mountains (i.e. Aggtelek, Rudabánya, Szendrő, Bükk, Uppony, Darnó, and Transdanubian Range), possibly also the Zemplínske Vrchy Mountains in SE Slovakia. The southern limits of the Internal Western Carpathians coincide with the Mid-Hungarian Lineament which separates them from the microcontinent Tisia (or Tisza-Dacia; Csontos 1995).

### *Pelso Megaunit*

The definition of the Internal Western Carpathians given above broadly corresponds to the original definition of the Pelso Megaunit by Fülöp *et al.* (1987). Subsequently, the 'Gemer-Bükk Unit' was also assigned to the Pelso Megaunit (e.g. Kovács 1992), and the entire megaunit was considered as a Tertiary block which was expelled, or escaped, from the Eastern Alps during collision and was welded to the Western Carpathians as late as the Oligocene–Lower Miocene. However, this interpretation does not conform with the observed surface structures of the Central Western Carpathians where a pre-Tertiary contact between the Veporicum and the Gemericum can be documented. Moreover, the northern boundary of the Pelso Megaunit (the Rába–Hurbanovo–Diósjenő Lineament) cannot, based on structural and palaeomagnetic investigations, be considered as a large-scale Tertiary sinistral strike-slip fault as would be expected if the Pelso Megaunit represented an escaping block. The Miocene escape included not only the Pelso Megaunit but also the Central Western Carpathians, i.e. the entire 'North Pannonian Unit' of Csontos *et al.* (1992), subsequently renamed the 'Alcapa' microcontinent (Csontos 1995; Fodor *et al.* 1999). These considerations support the idea of a Cretaceous-age welding of the Central Western Carpathians with the Pelso Megaunit, and suggest that the Pelso Megaunit is part of the Western Carpathians, a concept favoured here, despite the fact that the Central Western Carpathian (Slovako-Carpathian) units and the Pelso Megaunit were derived from very different Mesozoic palaeogeographic zones. A more recent definition of the Pelso Megaunit or 'Pelsonia Composite Terrane' by Kovács *et al.* (2000) includes only units occurring in Hungary, which implies a return, to some extent, to the original definition.

### *Tornaic Superunit*

The Tornaic Superunit (Tornaicum in the Hungarian, Turnaicum in the Slovak literature, South Rudabányaicum according to Kozur & Mock 1997) is defined as a rootless nappe system consisting of several partial units comprising sediments of Lower Pennsylvanian to Upper Triassic (probably up to Jurassic) age, overlying the Meliaticum and underlying the Silicicum (Mello 1997). In the western part of the Slovak Karst Mountains, it comprises the Slovenská skala Unit. The Turňa (Torna) Unit crops out from below the Silicicum in the Turňa Valley. In Hungary, the Tornaicum occurs mainly in the northern part of the Rudabánya Mountains (Mártonyi Unit; Fodor & Koroknai 2000; Less 2000). It has been suggested that Tornaic elements are present in the Alps, together with the Meliatic Florianikogel Unit (Kozur & Mostler 1992).

The stratigraphic succession of the Tornaicum includes the Bashkirian Turiec Formation (Vozárová & Vozár 1992), which is related to the Szendrő Phyllite Formation and the Hochwipfel Flysch of the Carnic Alps (e.g. Ebner 1992; Ebner *et al.* 2006). Permian continental sediments and Scythian clastics are overlain by Middle to Upper Triassic units. These are crucial for the definition of the Tornaicum and include Lower Anisian platform carbonates (mainly Steinalm Formation) and exclusively pelagic limestones from the Late Anisian on (= Žarnov, Nádaska, Reifling and Pötschen limestones). Carnian sediments, including shales, marlstones, sandstones and, locally, volcanics are also typical of the Tornaic Superunit.

The Tornaic Superunit displays a complex fold-and-thrust structure, forming a Jurassic accretionary complex together with the underlying Meliatic Superunit. Consequently, the boundaries between the Meliaticum and Tornaicum are sometimes uncertain. In most places, the Tornaicum shows evidence of low-grade, but relatively high-pressure metamorphism (about 7 kbar; Árkai & Kovács 1986).

*Bükk Superunit*

The Bükk Superunit (Bükkicum and Bükkium in the Slovak and Hungarian literature, respectively) comprises the eastern part of the Pelso Megaunit. It crops out in the Bükk, Uppony and Szendrö mountains. Some units in the Rudabánya Mountains probably also belong to the Bükk Superunit. Typically, the Bükk Superunit includes Palaeozoic and Triassic–Jurassic sedimentary and volcanic complexes which are lithologically similar to Dinaridic units and which underwent a low-grade Alpine, Cretaceous-aged metamorphism.

The structure of the Bükk Mountains is controversial. Csontos (1999, 2000) proposed that the Bükk Superunit consists of two fundamental units, the Bükk 'Parautochthon' and the Mónosbél-Szarvaskő nappe system (Mónosbél Unit and Szarvaskő Unit). This is the currently accepted model (albeit with some modifications).

The **Bükk Parautochthon** (Fennsíkum in terminology of Kozur & Mock 1997) consists of a Palaeozoic–Jurassic passive continental margin succession. The complexes of the Uppony Mountains and the 'North-Bükk Anticlinorium' include Silurian shales, Devonian–Lower Carboniferous platform carbonates, Middle Carboniferous deep-marine clastics, Lower Permian continental sediments, and Upper Permian and Lower Triassic shallow-water carbonates. Younger rocks, including Middle Triassic platform carbonates and calc-alkaline volcanics, Carnian clastics (Raibl Beds), and Upper Triassic platform carbonates only occur in the Bükk Mountains. Jurassic deep-water shales and radiolarites crop out in several synclines in the southern part of the Bükk Mountains.

The **Mónosbél Unit** and the **Szarvaskő Unit** are nappes (partly corresponding to the Bátor Nappe of Kozur 1991) and contain only Jurassic rocks: dismembered ophiolitic complexes, deep-water shales, radiolarites, and olistostromes with blocks of Triassic oceanic rocks. It is presumed that these south-vergent nappes were derived from a Jurassic backarc basin which formed to the north of the Bükk Parautochthon due to southward subduction of the Meliata Ocean (Balla *et al.* 1980, 1983; Kozur 1991). The nappes were thrust over the Bükk Parautochthon most likely during the Late Jurassic/Early Cretaceous.

There are no Cretaceous sediments in the entire area, with the exception of Gosau-type Senonian-age conglomerates in the Uppony Mountains. According to radiometric dating, the low-grade Alpine metamorphism in the Bükk Superunit is of Early Cretaceous age (Árkai *et al.* 1995). This metamorphism accompanied penetrative ductile deformation with several tight fold and cleavage sets. This deformation resulted in the overall south-vergent imbricated structure of the Bükk Mountains (Csontos 1999, 2000).

*Transdanubian Superunit*

The Transdanubian Superunit (= Transdanubicum or Bakonyicum) is a large tectonic complex occupying the western part of the Pelso Megaunit, and cropping out in the Transdanubian Range (Balaton Hills, Bakony Forest, Gerecse Mountains, Vértes Mountains, Buda Hills and Csővár Hills east of the Danube). Below the Tertiary cover, this superunit extends as far as the Hurbanovo-Diósjenő Fault to the north, the Rába Fault to the NW, and the Balaton Fault to the south. The superunit comprises weakly metamorphosed Lower Palaeozoic and unmetamorphosed Upper Palaeozoic and Mesozoic complexes overlain by Tertiary rocks. The Transdanubian Superunit is commonly considered to be analogous to the Upper Austro-Alpine nappe units, with facies links also to the South-Alpine realm (e.g. Tari 1995).

Palaeozoic rocks crop out locally near Lake Balaton and also occur below the Danube Basin in the vicinity of the Rába Fault. They consist of low-grade metamorphosed slates and phyllites with intercalations of metavolcanics and limestones. The age of the succession is Ordovician to Mississippian, the age of the metamorphism Variscan. In the Velence Mountains, late-Variscan granitoids also occur, showing a subalkaline geochemical trend. This succession is overlain by Pennsylvanian conglomerates.

The Upper Permian to Neocomian sediments represent a passive continental margin succession (e.g. Haas & Budai 1995, 1999; Császár & Haas 1984; Trunkó 1996; Haas *et al.* 2001). The Upper Permian sediments consist of terrestrial clastics and lagoonal evaporites, with marine dolomites occurring in the east. A general transgression at the Permian–Triassic boundary led to the deposition of shallow-marine carbonates followed by platform carbonates in the Middle Triassic. Anisian rifting resulted in the differentiation of the platform into subsiding intrashelf basins with Ladinian thin-bedded, cherty, sometimes bituminous limestones with thin tuffaceous 'Pietra Verde' intercalations (Buchenstein Formation), and elevated highs characterized by platform dolomites. The thickness of the Lower–Middle Triassic succession varies between 2000 and 2400 m.

Significant palaeogeographic changes occurred during the Late Triassic. The influx of fine-grained terrigenous material resulted in the deposition of Carnian marls which occur mainly above the basinal limestones, and are up to 800 m thick. In the Upper Carnian and Norian, almost the entire Transdanubian Range was the site of shallow-water carbonate platform sedimentation. The most widespread formation is the 1000–1500 m thick Hauptdolomit Formation. This is overlain by the Upper Triassic (Norian–Rhaetian) Dachstein Limestone Formation. In the eastern Csővár–Buda area, platform carbonates are replaced by basinal cherty limestones and dolomites. The total thickness of the Upper Triassic strata is up to 2000–2300 m.

The Triassic–Jurassic boundary is not reflected by a facies change within the Dachstein Limestone Formation. In the Jurassic, the sedimentary environments were differentiated into highs with comparatively thin, discontinuous and condensed lithofacies, and deeper-water areas with thicker, continuous successions showing less condensation.

Throughout the Liassic, nodular, cherty limestones, with interbedded Hierlatz Limestone and 'Ammonitico Rosso'-type limestones and marls (Adnet Formation) are found. Pelagic carbonate sedimentation continued up to the Middle Jurassic when it was replaced by siliceous marls and bedded radiolaritic cherts. Calcareous sediments reoccur in the Upper Jurassic ('Ammonitico Rosso'-type limestone and white pelagic cherty limestone and marls). The thickness of the Jurassic strata ranges from a few tens of metres up to 400–420 m. In the Bakony Mountains, the deposition of cherty limestones and marls was continuous from the Jurassic into the Lower Cretaceous. In contrast, the Lower Cretaceous strata in the Gerecse Mountains (northern part of the Transdanubian Range) represent a siliciclastic deep-marine succession including marls, turbiditic sandstones and conglomerates with ophiolitic detritus (e.g. Császár & Árgyelán 1994). Following the first Mesozoic phase of compressional deformation during the Albian, a new sedimentary cycle began with the deposition of freshwater and brackish marlstones, locally with bauxite lenses. Reef limestones (Zirc Formation) deposited on basin highs were overlain by shallow-marine marls extending up to the Cenomanian.

The surface structure of the Transdanubian Superunit appears to be rather simple and is dominated by a large syncline with some reverse faults along its flanks (e.g. the south-vergent Litér Thrust), which formed prior to the deposition of the Upper

Cretaceous sedimentary cover (Balla & Dudko 1993) and was overprinted by Tertiary transversal faults. However, according to Horváth (1993) and Tari (1995), the Transdanubian Superunit consists of nappes. The original thrust planes were reactivated during Miocene extension as low-angle normal faults (e.g. the Rába Fault). In contrast, other authors have interpreted the Rába Fault as a deep-reaching, steep crustal boundary of strike-slip character (Balla 1994). The Transdanubian Superunit, though influenced by Palaeogene back-thrusts and Neogene transtensional tectonics, represented a comparatively rigid block within the Pannonian Basin system.

*Zemplín Unit*

The Zemplín Unit is represented by pre-Tertiary strata of the Zemplín Mountains (southern part of the Neogene East Slovakian Basin, Fig. 18.19). The precise position of this unit is controversial, with some authors suggesting that it belongs to the Veporic Superunit while others derive it from Tisia (Tizia). The Zemplin Unit consists of high-grade crystalline basement of probably Variscan age, overlain by thick post-Variscan complexes including Pennsylvanian coal-bearing strata, Permian to Scythian continental sediments, and minor Middle Triassic carbonates. Along its NE boundary, the Zemplin Unit has a subsurface contact either with the Ptrukša Zone (= possible continuation of the Fatric Superunit), or directly with the Iňačovce-Krichevo Unit (Vozár *et al.* 1998).

## Senonian to Lower Miocene complexes of the Internal Western Carpathians

Large areas covered by Senonian-age, Gosau-type sediments occur in the western part of the Transdanubian Range (Bakony Mountains). These represent a new sedimentary cycle which began following Albian deformation and the Late Cenomanian–Late Santonian hiatus. The Senonian succession is more than 1000 m thick and commences with bauxites and terrestrial clastics with fluvial, limnic and paralic coal measures. Overlying marls are found in the basin lows while rudist-bearing platform limestones developed on basin highs. In the latest Cretaceous, the platforms were buried by pelagic pelitic-carbonatic sediments.

The **North Hungarian–South Slovakian Palaeogene–Lower Miocene Basin (Buda Basin)** had an elongate, SW–NE orientated shape. Palaeogene deposition began in the southern Bakony area with continental bauxite deposits during the Palaeocene to Early Eocene. The bauxites were overlain by Upper Eocene neritic nummulitic limestones and marls. In northern parts of the basin, coeval coal measures were deposited. Upper Eocene andesites occur near Recsk.

In most parts of the Transdanubian Range, Eocene sedimentation was followed by an erosional event ('infra-Oligocene denudation'). In the western part of the area, eroded Eocene rocks are unconformably overlain by brackish and freshwater Upper Oligocene deposits. The Oligocene transgression reached this area in the Late Kiscellian. Initially, coal beds were deposited, and these were overlain by littoral sandstones and subsequently by deep-water clays. During the Egerian, deep-water deposition continued (marls, calcareous siltstones and sandstones up to 700 m thick). Late Eggerian uplift and regression are recorded by prograding delta fans. During the Eggenburgian (Early Burdigalian), a new sedimentary cycle began in a sub-basin which was located further to the east (Fil'akovo-Pétervásara Partial Basin). This was a short-lived sub-basin filled with shallow-marine sandstones that ended with terrestrial clastics, uplift and erosion during the Late Eggenburgian. This uplift was associated with the first widespread evidence of explosive rhyolitic volcanism in the Pannonian area which recorded the onset of backarc lithospheric extension, asthenospheric upwelling and crustal melting. During the Ottnangian the deposition of coal measures was widespread, while basinal marls were again deposited during the Karpatian (Novohrad-Nógrád Partial Basin). Sedimentation was terminated in the Early Badenian (Langhian) due to regional uplift and erosion of the underlying strata. This was followed by rifting of the Pannonian Basin System and widespread andesitic volcanism.

## Palaeotectonic evolution of the Internal Western Carpathians

As shown in Figures 18.22 and 18.23, the tectonic history of Internal Western Carpathians can be summarized as follows.

1. Early Palaeozoic to Mississippian: Internal Western Carpathians located in the external and foreland zones of the Variscan Orogen.
2. Pennsylvanian: marine transgression; Permian to Scythian: shallow sea.
3. Triassic: gradual subsidence; marked Upper Anisian (Pelsonian) rifting event related to the breakup of the Meliata Ocean.
4. Early to Middle Jurassic: rifting; destruction of the Triassic carbonate platform.
5. Upper Jurassic: closure of the Meliata Ocean; formation of a collisional belt.
6. Early Cretaceous: backthrusting in the Bükk Superunit; deposition of synorogenic deep-marine clastics with input from obducted ophiolites in the NW part of the Transdanubian Superunit, probably adjoining the suture of the Meliata Ocean.
7. Mid-Cretaceous: southward (in present-day co-ordinates) thrusting in the Transdanubian Superunit.
8. Senonian: extension and shallow-marine deposition (Gosau Supergroup).
9. Eocene to Early Miocene: formation and subsidence of the Buda Basin.
10. Middle to Late Miocene: backarc extension (Pannonian Basin System) and widespread calc-alkaline volcanism.

## Pannonian Basin System and Neogene volcanics

During Miocene times, there was thrusting in the External Western Carpathians, the Central and Internal Western Carpathians were largely influenced by lithospheric stretching, basin formation and volcanism (see reviews by Kováč *et al.* 1997, 1998; Kováč 2000). These processes resulted in the formation of the Pannonian Basin System, including, among others, the Danube Basin and the Transcarpathian Basin. The Vienna Basin is also part of the Pannonian Basin System but occupies a slightly different position, partly covering the External Western Carpathians.

Remnants of Lower Miocene (dominantly Eggenburgian) sediments are found in the northern parts of the present Vienna and Danube basins, where they were deposited in a transpressional, wrench-fault dominated setting. The lozenge-shaped **Vienna Basin** is a pull-apart basin formed by late Early Miocene (Karpatian) sinistral wrenching along the Mur-Mürz-Leitha-Láb-Dobrá Voda Fault System trending SW–NE. The basin subsided during the Middle–Upper Miocene and in Pliocene–Quaternary times.

Large basins within the Central and Internal Western Carpathians include the Danube Basin and the Transcarpathian Basin. The **Danube Basin** formed during the Middle–Late Miocene by NW–SE extension, accommodated by NE–SW to NNE–SSW striking listric normal faults and detachment faults. The detachment faults partly reactivated Cretaceous-age thrusts (e.g. Horváth 1993).

The **Transcarpathian Basin** to the east had a complicated history. Lower Miocene (Eggenburgian) basinal sediments found in the NW part of the basin were deposited in a remnant forearc compressional depression, while the Transcarpathian Basin itself was formed as a result of Karpatian–Early Badenian dextral transtension along NW–SE orientated oblique-slip faults. Transtension was followed by general extension during the Late Badenian to Sarmatian, accompanied by extensive volcanism. Compressional inversion characterizes the basin evolution during the Pliocene.

Several smaller intramontane Neogene basins are restricted to western and central Slovakia, located between the 'core mountains' of the Fatra-Tatra Belt. Their evolution was controlled by block rotation, tilting and the uplift of adjacent highs.

Tertiary volcanics related to subduction in the External Western Carpathians and to backarc extension connected with asthenospheric updoming, are mainly found in the Internal and Central Western Carpathians, and more rarely in the External Western Carpathians. These calc-alkaline and alkaline volcanics include basaltic, intermediate and acidic lavas, subvolcanic complexes, and pyroclastic rocks of Neogene to Pleistocene age. Based on their age, spatial distribution and geochemical/petrological character, four stages of volcanism can be distinguished (Konečný & Lexa 1984; Konečný *et al.* 2002; Lexa & Konečný 1998). These are:

(1) *Areally extensive dacitic to rhyolitic volcanism* (Eggenburgian–Early Badenian). This stage involved the widespread deposition of tuffs and ignimbrites and the formation of extrusive domes in the source areas. Thickness varies from tens to hundreds of metres, exceeding 1000 m in the central part of the Pannonian Basin System. The rocks were derived from crustal anatectic melts that originated due to overheating of the continental crust as a result of mantle updoming in an extensional regime.

(2) *Areally extensive andesitic volcanism* (Early Badenian–Early Pannonian). Products of this stage are most widespread. Several stratovolcanoes formed which exhibit alternating effusive and extrusive activity. Additional stratovolcanoes are buried below the younger sediments of the Danube Basin. This stage commenced during the Early Badenian in the west and NW part of the Pannonian Basin System. In the Central Slovakian volcanic area (Fig. 18.29) it continued intermittently through to the Early Pannonian. Pyroxene-amphibole andesites (sometimes with garnet) and their pyroclastic products are dominant. Beginning in the Sarmatian, acidic volcanic rocks of rhyolite to rhyodacite composition also occur. This calc-alkaline, mainly intermediate magmatism represents mantle-derived melts that were significantly contaminated by crustal material. Magmatic activity can be genetically related to asthenosphere upwelling and lithospheric stretching due to the subduction of the oceanic basement of the Magura Zone (External Western Carpathians) beneath the continental Tatric-Veporic plate margin.

(3) *Arc-type andesitic volcanism* (Late Badenian–Sarmatian). Rare products of this stage are found along the Pieniny Klippen Belt (Horné Sŕnie, Pieniny Andesite Line in Poland). In Moravia they are also found in the Carpathian Flysch Belt (Uherský Brod). More significant occurrences are in eastern Slovakia (Slanské vrchy Mountains, Vihorlat Mountains) where they form a chain of stratovolcanoes continuing into the Ukraine and Romania. Basaltic and pyroxene andesites (with trachybasalts) originated due to the subduction of the Silesian-Krosno-Moldavian oceanic crust of the External Western Carpathians.

(4) *Post-orogenic alkaline-basaltic to basanitic volcanism* (Pannonian–Pleistocene). This stage is represented by relatively small occurrences of basaltic volcanics and involved the formation of diatremes, maars, scoria cones and lava flows. This type of volcanic activity began in the west (SE part of Austria) during the Pannonian and extended through the Danube Basin and the Transdanubian Range as far as the South Slovakian–North Hungarian Basin. Lava flows and a scoria cone overlying the Hron river terrace near Nová Baňa appear to be the youngest products of this stage of volcanic activity (Šimon & Halouzka 1996). Their age was determined as *c.* 100 ka (Šimon & Maglay 2005). The volcanics are interpreted as the products of post-orogenic extension which was characterized by asthenospheric mantle updoming and the ascent of basaltic magmas through fault-weakened zones of the thinned crust.

## *Permian and Triassic tectonics of the Western Carpathians*

The Variscan Orogeny in Central Europe terminated in a phase of gravitational collapse associated with rifting and extensive magmatism (regional **Betliar Phase**, see Figs 18.22 & 18.23a). Thus, the Early Mesozoic rifting was preceded by significant extensional tectonic events during the Permian (Vozárová & Vozár 1988; see also McCann 2008*b*). A particularly thick, rift-related, continental succession (Ipoltica Group) forms the sole of the Hronic cover nappe system. In this succession, the Permian continental sediments are associated with voluminous calc-alkaline to continental tholeiitic, andesitic and basaltic volcanism, which erupted in two distinct phases during the Early and Late Permian (Vozárová & Vozár 1988; Vozár 1997; Dostal *et al.* 2003). The original palaeogeographic position of this mature rift sequence is not precisely known. Other deep, but comparatively narrow Permian rift basins filled with immature continental clastics originated in the North Tatric, Veporic and Gemeric zones. Their formation was usually accompanied by calc-alkaline and/or subalkaline volcanism and A-type (Uher & Broska 1996, Broska & Uher 2001) as well as specific S-type (Poller *et al.* 2002) granitic plutonism. This rift-related silicic magmatism probably extended up into the Middle Triassic (*c.* 240 Ma; Kotov *et al.* 1996; Putiš *et al.* 2000; Uher & Broska 2000). The site of the future Meliata oceanic rift was characterized by strong Scythian subsidence and shallow-marine terrigenous sedimentation accompanied by alkaline rhyolitic volcanism (Uher *et al.* 2002). These features indicate that the terminal Variscan events (i.e. orogenic collapse and lithospheric attenuation) could have been genetically related to the early Alpine rifting in the southern zones of the Central Western Carpathians, which ultimately led to the opening of the Meliata Ocean.

At *c.* 240 Ma (Upper Anisian) the breakup of the Meliata Ocean occurred (Kozur 1991). This ocean formed probably as a backarc basin related to the northward subduction of Palaeotethys under the Eurasian continent (e.g. Stampfli 1996). Middle Triassic rifting was accompanied by widespread calc-alkaline volcanism in the Internal Western Carpathian zones (Buchenstein Formation in the Transdanubian and Bükk superunits). In contrast, the axial zones of the Meliatic superunit were marked by backarc-basin- and mid-ocean-ridge-type basalts (Ivan 2002).

The oldest deep-sea sediments, radiolarites, are of Ladinian age (Mock *et al.* 1998).

The Upper Anisian (Pelsonian) rifting and initial opening of the Meliata Ocean is termed the **Žarnov Phase** (Figs 18.22 & 18.23a). In the areas adjacent to the Meliaticum, this event was recorded by a breakup unconformity between the Lower Anisian ramp and platform carbonates and Pelsonian pelagic limestones. The younger Triassic pelagic facies include deep-water, partially condensed nodular limestones and Ladinian–Norian radiolarites. In more northern Slovako-Carpathian superunits (Tatricum, Fatricum) the evidence of Anisian rifting is much weaker, although slumps, tempestites and tsunamites occur within the carbonate ramp or platform complexes (e.g. Michalík 1997).

## *Jurassic tectonics*

The regional geodynamic situation changed considerably by the earliest Jurassic. The broad northern shelf of the Meliata Ocean underwent widespread rifting and, contemporaneously, subduction of Meliata oceanic lithosphere commenced. Most probably these processes were related to a change in the large plate movement kinematics (i.e. the beginning of southeastward drift of Africa and Adria relative to Europe during the opening of the Central Atlantic). The Western Carpathian orogenic wedge began to form by accretion of material scraped off from the subducted Meliata lithosphere to the tip of the upper plate. Thus the Jurassic period represents a nascent stage of the Western Carpathian orogenic wedge growth. Based on the timing of marked changes in the bathymetric evolution and on the character and distribution of syn- and post-rift sedimentary sequences, four main Jurassic–Cretaceous rifting phases can be identified within the Western Carpathian area (Plašienka 2003*a*): two Early Jurassic rifting phases resulting from lithospheric stretching and leading to the fragmentation of the Triassic platform (Zliechov Phase and Devin Phase); one rifting phase that led to the breakup of the South Penninic-Vahic Ocean in the late Middle Jurassic (Krasín Phase); and one rifting phase resulting in the breakup of the North Penninic-Magura Ocean in the Early Cretaceous (Walentowa Phase).

The Hettangian–Sinemurian-age **Zliechov Phase** of the 'wide rift' type is well recorded especially in the Tatric, Fatric and Hronic domains (Figs 18.22 & 18.23a). It involved overall uniform stretching of the continental lithosphere and resulted in the formation of broad subsiding intracontinental basins (Zliechov, Šiprúň, Kysuca-Czorsztyn-Magura), which were separated by subaerial highs (i.e. South Tatric and North Tatric ridges). For *c.* 100 Ma following the Zliechov Phase, sediment-starved basins within the Tatricum (Šiprúň Basin) and the Fatricum (Zliechov Basin) were subjected to slow thermal subsidence with accompanying deep-water pelagic sedimentation.

The Toarcian-age **Devín Phase** was more localized in terms of its effects. The area of the North Tatric Ridge (partly analogous to the Lungau Swell of Tollmann 1977) underwent Lower Jurassic uplift and erosion of the pre-rift Triassic successions, so that Upper Liassic–Lower Dogger clastic limestones were deposited on deeply eroded Triassic strata or on pre-Alpine basement complexes (Michalík *et al.* 1993*c*; Plašienka 1999*a*). The Devín Phase was probably also the main stage of extensional block tilting as indicated by the increasing contrast between the hemipelagic, partly anoxic sedimentation in the half-grabens (Allgäu Formation) and the deposition of 'Ammonitico Rosso'-type limestones (Adnet Formation) in well-aerated environments on the elevated edges of the tilted fault blocks (Soták & Plašienka 1996; Wieczorek 2000, 2001).

The Bajocian-age **Krasín Phase** strongly affected areas to the north of the North Tatric Ridge. Simple shear extension along a lithospheric detachment fault dipping towards the NW resulted in additional stretching in the Kysuca Basin and the late Bajocian to early Bathonian breakup and opening of the Vahic (South Penninic) Ocean (Plašienka 2003*a*). The breakaway fault of the detachment was the bounding fault of the Infra-Tatric Borinka half-graben (Malé Karpaty Mountains) where exceptionally thick Middle Jurassic scarp breccias were deposited (Somár Formation; Plašienka 1987). These were derived from the North Tatric Ridge which at that time formed the lower plate. The Czorsztyn Ridge formed at this time (e.g. Aubrecht & Szulc 2006), probably as a result of thinning of the lithosphere under the distal upper-plate margin by the detachment fault. Coeval subsidence of the basinal areas (i.e. Magura, Vahic, Šiprúň, Zliechov) to abyssal depths below the CCD is indicated by widespread deposition of radiolarites from Callovian through to Kimmeridgian times.

Collision between the Slovako-Carpathian continental margin and the Pelso Megaunit, following the closure of the Meliata Ocean, commenced during the late Middle Jurassic. Subsequently, the Slovako-Carpathian margin was overridden by the Meliatic accretionary complex in the Late Jurassic (**Šugov Phase**). Subsequently, the Gemeric Superunit was stacked over the Veporic Superunit (**Tuhár Phase**). An extensional tectonic regime and low-energy pelagic sedimentation still prevailed in the foreland of the convergent system (Figs 18.22 & 18.23a, b).

## *Cretaceous tectonics*

The Berriasian–Hauterivian **Walentowa Phase** represents the migration of rifting into the foreland area and is interpreted as recording the breakup of the Magura Ocean. This breakup was preceded by asymmetric rifting. Unlike the Krasín Phase, the detachment fault in this case dipped to the SE (Plašienka 2003*a*). Lithospheric extension and mantle upwelling triggered the second uplift event of the upper plate (Czorsztyn Ridge), which was manifested through the shallowing of depositional environments and the deposition of synrift debris flows and carbonate scarp breccias (Walentowa Breccia; Birkenmajer 1977; Krobicki & Słomka 1999). This was followed by widespread uplift, karstification, erosion and non-deposition extending into Albian times (Aubrecht *et al.* 2006). The Tatricum and Fatricum (including the basin highs) were characterized by generally uniform pelagic sedimentation during the Neocomian, interrupted by occasional incursions of turbidites (Michalík *et al.* 1996).

Marked subsidence of the Magura Basin to abyssal depths following the Walentowa Phase and the breakup of the Magura Ocean is evidenced by the presence of hemipelagic and turbiditic sedimentation commonly below the CCD, beginning in the Barremian and continuing throughout the Cretaceous. The Walentowa Phase was also accompanied, and followed, by submarine extrusion of primitive, mantle-derived alkaline basalts. This magmatic activity commenced in the Berriasian and continued up into the Lower Albian. Volcanic products occur in the Fatric Superunit, Tatric Superunit and the Silesian-Krosno units (e.g. Spišiak & Hovorka 1997; Lucińska-Anczkiewicz *et al.* 2002; Dostál & Owen 1998).

The above-noted extensional tectonic phases were followed by the **Solírov Phase** (Barremian–Early Albian) which is marked by the growth of the Urgonian carbonate platforms on the former South-Tatric Ridge and by syntectonic sedimentation around the North Tatric Ridge (calciturbiditic Solírov Formation; Jablonský *et al.* 1993). Following the Solírov Phase, in the Middle Albian, synorogenic, coarsening-upward deep-marine clastic sedimentation commenced in the Fatric and Tatric domains (Poruba Formation). These mid-Cretaceous successions are interpreted as reflecting underthrusting of the Fatricum beneath the northern part of the Veporicum (**Benkovo Phase**; Figs 18.22 & 18.23b). This epoch ended with the emplacement of extensive Fatric and Hronic cover nappes on the Tatricum (**Donovaly Phase**).

Changes in both the kinematics and dynamics of tectonic activity in the western Tethyan realm during the Late Cretaceous were related to changing relative motions of large plates, particularly the onset of convergence between Africa-Adria and Europe (e.g. Dewey *et al.*, 1989; Savostin *et al.* 1986). The advance of the Adriatic indenter, represented by the Pelso Megaunit, led to shortening and modification of the central, thermally-weakened parts of the original collisional wedge and triggered transpressional exhumation of the Veporic metamorphic dome in the Senonian (**Kohút Phase**). In upper structural levels, this exhumation was achieved by orogen-parallel, extensional unroofing (Plašienka *et al.* 1999; Janák *et al.* 2001). At the same time, the compressional stresses were transmitted from the indenter towards the front of the rigid Tatricum where subduction of the Vahic Ocean lithosphere commenced (**Selec Phase**; Figs. 18.22 & 18.23c). Moreover, the contractional regime affected the entire Alpine-Carpathian foreland with far-field effects extending into the distant interior of the North European Platform, where it initiated transpressional inversion of basins (see Reicherter *et al.* 2008).

### *Tertiary tectonics*

Late Cretaceous lateral extension reduced the angle of taper of the Western Carpathian orogenic wedge. Consequently, the collision of the wedge tip with the Oravic continental fragment generated a strong contraction event within the wedge (Maastrichtian–Danian **Jarmuta Phase**; Figs 18.22 & 18.23c). Similarly, the gravitational collapse of the wedge during the Eocene (**Súl'ov Phase**, formation of the Central Carpathian Palaeogene Basin) was replaced by compression during the Late Oligocene–Early Miocene collision of the External Carpathian accretionary wedge with the North European Platform (**Beskydy Phase**). This compressional event migrated backward into the entire wedge (**Kamenica Phase**).

During the final Middle–Late Miocene tectonic phases, the roll-back rate of the subduction zone of oceanic domains of the External Western Carpathians (first Magura, then Silesian-Krosno and Moldavian) exceeded the rate of advance of the indenter. Consequently, the convergent system changed from advancing to retreating (Royden 1993). As a result, the Western Carpathian orogenic system began to be governed by a general extensional regime (**Alföld Phase**). The compressional regime persisted only in the most easterly areas. There, the wedge continued to grow by frontal accretion of sedimentary units scraped off the subducting Moldavian lithosphere and the passive margin of the North European Platform.

## *Present-day crustal structure and neotectonics*

The present knowledge about the deep structure of the Western Carpathians is based on an extensive database of geophysical measurements. The structure of the Western Carpathian lithosphere and crust has been investigated using a wide range of geophysical methods, including seismic refraction and reflection profiling, seismology, gravimetry, magnetotellurics, geothermics and magnetometrics. The section is largely based on the review papers of Bielik & Šefara (2002) and Bielik *et al.* (2004), which summarize the results of various geophysical studies.

The complicated structure of the Western Carpathians resulted from the interference of the pre-Alpine, early Alpine and, predominantly, the late Alpine deep-seated tectonic activity. The fundamental role of the youngest lithosphere-scale collisional processes associated with the formation of the backarc Pannonian Basin System has been emphasized by many authors (e.g. Lillie *et al.* 1994; Lenkey 1999; Kováč 2000). Geophysical studies in this region have clearly shown that tectonic units differing in thermotectonic age, lithologic/rheologic stratification and crustal thickness produce important variations in the structure and geodynamics of the lithosphere (Horváth 1993; Lillie *et al.* 1994; Šefara *et al.* 1996; Lenkey 1999; Bada 1999; Szafián 1999).

### *The Western Carpathian lithosphere*

Recent estimates of the lithosphere thickness in the Carpathians have been made by Zeyen *et al.* (2002) using integrated lithosphere modelling. They presented 2D numerical models based on a combined interpretation of heat flow, gravity data and absolute topographic elevation. Unlike older models, this integrated modelling shows important differences in lithosphere thickness along the strike of the Western Carpathian Orogen. In contrast to the western part of the orogen, the lithosphere increases in thickness to a maximum of 140–150 km beneath the central and eastern parts of the Western Carpathians. To explain this lithospheric root, Zeyen *et al.* (2002) suggested that during the relatively short time following collision and slab break-off, convergence continued and resulted in lithospheric thickening of *c.* 40 km over a period of 2–4 Ma assuming a convergence rate of 1–2 cm/a.

Based on the results of this integrated modelling, critical analysis of earlier models and new interpolation of older data, the approximation of the lithosphere thickness in the Carpathian region was modified. The estimated total error of lithosphere thickness is not greater than 20 km. The hinterland of the Western Carpathians in the backarc Pannonian Basin System is characterized by a thin lithosphere. Based on the seismic and magnetotelluric surveys, the lithosphere may be as thin as 40–60 km beneath some of the sub-basins of the Pannonian Basin System (Posgay *et al.* 1995; Ádám *et al.* 1996). It has been suggested that stretching of the plate and asthenospheric mantle updoming caused this extreme lithospheric thinning (e.g. Bielik 1988; Lillie *et al.* 1994; Kováč 2000; Konečný *et al.* 2002).

The surface heat flow density in the Pannonian Basin System is one of the highest terrestrial heat flows measured. Additionally, a very strong blanketing effect was determined (up to 32%) in the central part of the basin system (Lenkey 1999), which emphasizes its temperature abnormality. The high present-day heat flow in this region may be attributed to the phase of Early–Middle Miocene lithospheric extension (Royden *et al.* 1983*a*, *b*). Thus, the areas characterized by high heat flow are underlain by a thin lithosphere.

The Carpathian–Pannonian area represents one of the key areas for studying the influences of various mechanical parameters on lithospheric rheology since it is possible to simultaneously observe several thermotectonic regimes in a relatively small area. Results of pioneering studies dealing with the prediction of lithospheric rheological behaviour in the Western Carpathians and in the surrounding tectonic units have been presented by Lankreijer *et al.* (1999) and Bielik *et al.* (2000). These indicate a general decrease in the mechanical strength of the lithosphere beneath the Western Carpathians from north to south. The effective elastic thickness varies between 15 and 23 km. In comparison with the older European platform, the rheological strength of the lower crust is strongly reduced in the Western Carpathians. This is a consequence of the increased temperature of the crust. The lithosphere strength further decreases and eventually completely disappears in the Pannonian Basin System.

The Bohemian Massif represents a mechanically rigid block extending down to a depth of *c.* 60 km. The predicted effective elastic thickness values for this region are 20–40 km. Based on this result, Lankreijer *et al.* (1999) assumed that, as a consequence of the high strength, that the Bohemian Massif blocked the northward movement of the colliding Alpine region. This was, possibly, the main reason for the Miocene-age sinistral strike-slip movement in a NE–SW direction in the Eastern Alps and its continuation into the Western Carpathians (Mur-Mürz-Leitha-Láb-Dobrá Voda Fault Zone). The result of this process was the opening of the Vienna Basin at the East Alpine–Western Carpathian junction by a pull-apart mechanism during the Karpatian (17.5–16.4 Ma).

The Pannonian Basin System, including the Danube and East Slovakian Basins, is characterized by only one, relatively thin rigid layer located in the uppermost 10 km of the crust. Thus, there is no strength in the lower crust or lithosphere in this region. The extreme flexibility of the Pannonian lithosphere is a direct result of the high values of heat flow and the very shallow and warm asthenosphere. The predicted effective elastic thickness is only 0–10 km. The calculations of Lankreijer (1998) also indicate that the peripheral parts of the Pannonian Basin are, at present, more resistant to deformation than the central depressions, which suggests that the effective elastic thickness decreases from the margins to the central part of the Pannonian Basin System.

The rheological prediction that only a thin layer of upper crust carries most of the strength of the lithosphere in the Carpathian–Pannonian area is in good agreement with the earthquake hypocentres which occur only to a depth of 15–17 km (e.g. Kováč *et al.* 2002).

## *The Western Carpathian crust*

The crust of the Western Carpathians and neighbouring areas has a complicated structure and is composed of fragments formed during the Variscan, palaeo-Alpine and neo-Alpine orogenic events. The accretionary wedge of the External Western Carpathians originated in the Tertiary. The Central and Internal Western Carpathians are composed of tectonic units that originated during the palaeo-Alpine Orogeny in the Mesozoic. The main, crustal-scale Alpine tectonic units of the Central Western Carpathians (Tatric, Veporic and Gemeric superunits) consist of pre-Alpine crystalline basement and its Upper Palaeozoic–Mesozoic cover. The tectonic units of the Variscan Orogeny form the crystalline basement.

The crustal thickness varies considerably across the Western Carpathians. Initial data on the regional Moho depth were based on deep seismic surveys. However, these incorporate large uncertainties. A more accurate determination of the Moho depth in the Western Carpathians and Bohemian Massif was achieved by the common depth point method (e.g. Tomek *et al.* 1987, 1989; Tomek & Hall 1993).

Crustal thickness (Moho depth) clearly tends to increase from west to east along the Carpathian Orogen. The Western Carpathians are characterized by crustal thicknesses of *c.* 30–35 km, while in regions influenced by Tertiary extension such as the Pannonian Basin System the Moho rises up to *c.* 25 km depth.

The present-day internal crustal structure of the Western Carpathians is a complex combination of structures originated during the Variscan and Alpine orogenies. A dominant role has been played by the Late Alpine, Tertiary extensional tectonics. This was documented for the entire Central Western Carpathian region (Bezák *et al.* 1993; Plašienka *et al.* 1997; Kováč 2000). On the deep seismic transect 2T (Tomek *et al.* 1987; Tomek 1993; Vozár *et al.* 1998; Bielik *et al.* 2004), a northerly inclined package of reflections is observed in the upper crust in the northern part of the profile (Pieniny Klippen Belt and adjacent areas). This package is interpreted as resulting from backthrusts related to Tertiary collision. In the southern part of the profile the reflectors are inclined towards the south or SE and are interpreted as palaeo-Alpine faults (Cretaceous). In the southernmost part (Pannonian Basin System) reflectors appear to cross not only the lower crust, but also the Moho. In the continuation of one of the reflectors a zone of low resistivity in the upper mantle (down to a depth of about 50 km) was modelled based on magnetotelluric measurements. This was interpreted as partially molten asthenospheric material, the probable source for the youngest Plio-Quaternary volcanic rocks (alkaline basalts) found on the surface (Šefara *et al.* 1998).

The fragmentation of the crust is more pronounced in the western part of the Western Carpathians (3T profile; Tomek *et al.* 1987; Vozár *et al.* 1998). The subsurface of the Danube Basin shows seismic reflectors which are interpreted as a low-angle normal fault system unroofing the basin floor during the formation of the basin (Šefara *et al.* 1998). The final phase of sedimentation (starting from the lower Pannonian) is related to a period of thermal subsidence in this area (Royden *et al.* 1983*a*, *b*; Lankreijer 1998).

The crustal structure in the basement of the East Slovakian Basin is characterized by three main features (Kováč *et al.* 1995; Bielik 1998): (1) down-bending of the North European Platform margin in the collision area into a steep orientation; (2) flower structure developed by transpression along the Pieniny Klippen Belt; (3) tectonic unroofing under transtensional/extensional conditions in the basement of the East Slovakian Basin. Unroofing of the Penninic-related Iňačovce-Krichevo Unit (Soták *et al.* 1993) and associated crustal thinning and increase of heat flow were related to Middle Miocene extension.

## *Neotectonics*

Ongoing neotectonic processes in the Western Carpathians are characterized by generally constant stress field characteristics and corresponding tectonic regimes for the Pliocene to Holocene, i.e. for the last approximately 5.3 Ma (e.g. Hók *et al.* 2000). The main source of the contemporaneous stress field in the Carpatho-Pannonian area is the counterclockwise rotation of Adria microplate ('Adriatic push'). Secondary sources are compressions generated by collisions in the Vrancea region of the Eastern Carpathians and at the southern corner of the Bohemian Massif in the Western Carpathians (Bada 1999). Variations in the effective elastic thickness of the lithosphere and topography-related sources have only a local significance.

Data on the recent stress field in the Western Carpathians are comparatively scarce, and are derived from focal mechanisms, structural analyses and *in situ* measurements (Pospíšil *et al.* 1992; Jarosiński 1998; Bada 1999; Hók *et al.* 2000; Kováč *et al.* 2002). These data suggest ongoing, albeit very slow, convergence between the North European Platform and the Carpathian orogen with the maximum horizontal stress axis being perpendicular to the orogen front along the arc. On the other hand, the interior regions of the Pannonian Basin are undergoing general extension and subsidence.

Moving from north to south, several provinces with differing structural regimes can be distinguished (Hók *et al.* 2000; Kováč *et al.* 2002): (1) the External Western Carpathian area which is characterized by a general NW–SE to north–south compression generating thrust faulting and uplift; (2) the Pieniny Klippen Belt which exhibits ongoing sinistral strike-slip movements; (3) the Central Western Carpathian area which is dominated by NW–SE compression, although NE–SW extension also occurs.

Important north-trending wrench-fault zones separate regions with differing vertical movements. The Central Slovakian Fault System separates western Slovakia, dominated by block rotation, uplift of 'core mountains' and subsidence of intramontane basins, from central Slovakia showing general uplift. The Hornád Fault System delineates the western boundary of the subsiding area in eastern Slovakia. The transition zone between the mountainous Central Western Carpathians and the Pannonian Basin shows predominantly orogen-parallel extension and subsidence.

Recent seismic activity in the Western Carpathians is concentrated into several zones that are interpreted to be a result of the interference of the actual stress field and the pre-existing crustal discontinuities (Šefara *et al.* 1998; Kováč *et al.* 2002). The most notable is the Pieniny Klippen Belt, which represents the deep-seated contact of the North European Platform with the Central Western Carpathians. The Mur-Mürz-Leitha-Láb-Dobrá Voda Fault Zone, trending SW–NE along the southern margin of the Vienna Basin, is the still-active sinistral wrench zone that accommodated the eastward extrusion of the Alcapa plate out of the collision zone in the Eastern Alps during the Neogene. The extensionally reactivated Čertovica and Rába-Hurbanovo-Diósjenő low-angle fault zones generate occasional earthquakes. The expected maximum epicentral intensity in these zones has been estimated as $I_{max} = 7–9$ (Šefara *et al.* 1998).

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