17 Palaeogene and Neogene

MICHAEL W. RASSER, MATHIAS HARZHAUSER (co-ordinators), OLGA Y. ANISTRATENKO, VITALIY V. ANISTRATENKO, DAVIDE BASSI, MIRKO BELAK, JEAN-PIERRE BERGER, GIANLUCA BIANCHINI, SAFET ČIČIĆ, VLASTA ĆOSOVIĆ, NELA DOLÁKOVÁ, KATICA DROBNE, SORIN FILIPESCU, KARL GÜRS, ŠÁRKA HLADILOVÁ, HAZIM HRVATOVIĆ, BOGOMIR JELEN, JACEK ROBERT KASIŃSKI, MICHAL KOVÁČ, POLONA KRALJ, TIHOMIR MARJANAC, EMŐ MÁRTON, PAOLO MIETTO, ALAN MORO, ANDRÁS NAGYMAROSY, JAMES H. NEBELSICK, SLAVOMÍR NEHYBA, BOJAN OGORELEC, NESTOR OSZCZYPKO, DAVOR PAVELIĆ, RAJKO PAVLOVEC, JERNEJ PAVŠIČ, PAVLA PETROVÁ, MARCIN PIWOCKI, MARIJAN POLJAK, NEVIO PUGLIESE, REJHANA REDŽEPOVIĆ, HELENA RIFELJ, REINHARD ROETZEL, DRAGOMIR SKABERNE, L'UBOMÍR SLIVA, GERDA STANDKE, GIORGIO TUNIS, DIONÝZ VASS, MICHAEL WAGREICH & FRANK WESSELINGH

Over the last 65 Ma, our world assumed its modern shape. This timespan is divided into the Palaeogene Period, lasting from 65 to 23 Ma and the Neogene, which extends up to the present day (see Gradstein & Ogg (2004) and Gregory *et al.* (2005) for discussion about the Quaternary).

Throughout the Cenozoic Era, Africa was moving towards Eurasia in a northward direction and with a counterclockwise rotation. Numerous microplates in the Mediterranean area were compressed, gradually fusing, and Eurasia underwent a shift from a marine archipelago to continental environments, related to the rising Alpine mountain chains (Figs 17.1 & 17.2). Around the Eocene–Oligocene boundary, Africa's movement and subduction beneath the European plate led to the final disintegration of the ancient Tethys Ocean. The Indo-Pacific Ocean came into existence in the east while various relict marine basins remained in the west. In addition to the emerging early Mediterranean Sea, another relict of the closure of the Tethys was the vast Eurasian Paratethys Sea.

The Oligocene and Miocene deposits of Central Europe are largely related to the North Sea in the north, the Mediterranean Sea in the south and the intermediate Paratethys Sea and its late Miocene to Pliocene successor Lake Pannon. At its maximum extent, the Paratethys extended from the Rhône Basin in France towards Inner Asia. Subsequently, it was partitioned into a smaller western part consisting of the Western and the Central Paratethys and the larger Eastern Paratethys. The Western Paratethys comprises the Rhône Basin and the Alpine Foreland Basin of Switzerland, Bavaria and Austria. The Central Paratethys extends from the Vienna Basin in the west to the Carpathian Foreland in the east where it abuts the area of the Eastern Paratethys. Eurasian ecosystems and landscapes were impacted by a complex pattern of changing seaways and land bridges between the Paratethys, the North Sea and the Mediterranean as well as the western Indo-Pacific (e.g. Rögl 1998; Popov *et al.* 2004). This geodynamically controlled biogeographic differentiation necessitates the establishment of different chronostratigraphic/geochronologic scales.

The geodynamic changes in landscapes and environments were further amplified by drastic climate changes during the Cenozoic. The warm Cretaceous climate continued into the early Palaeogene with a distinct optimum near the Palaeocene–Eocene boundary (Palaeocene–Eocene Thermal Maximum) and the

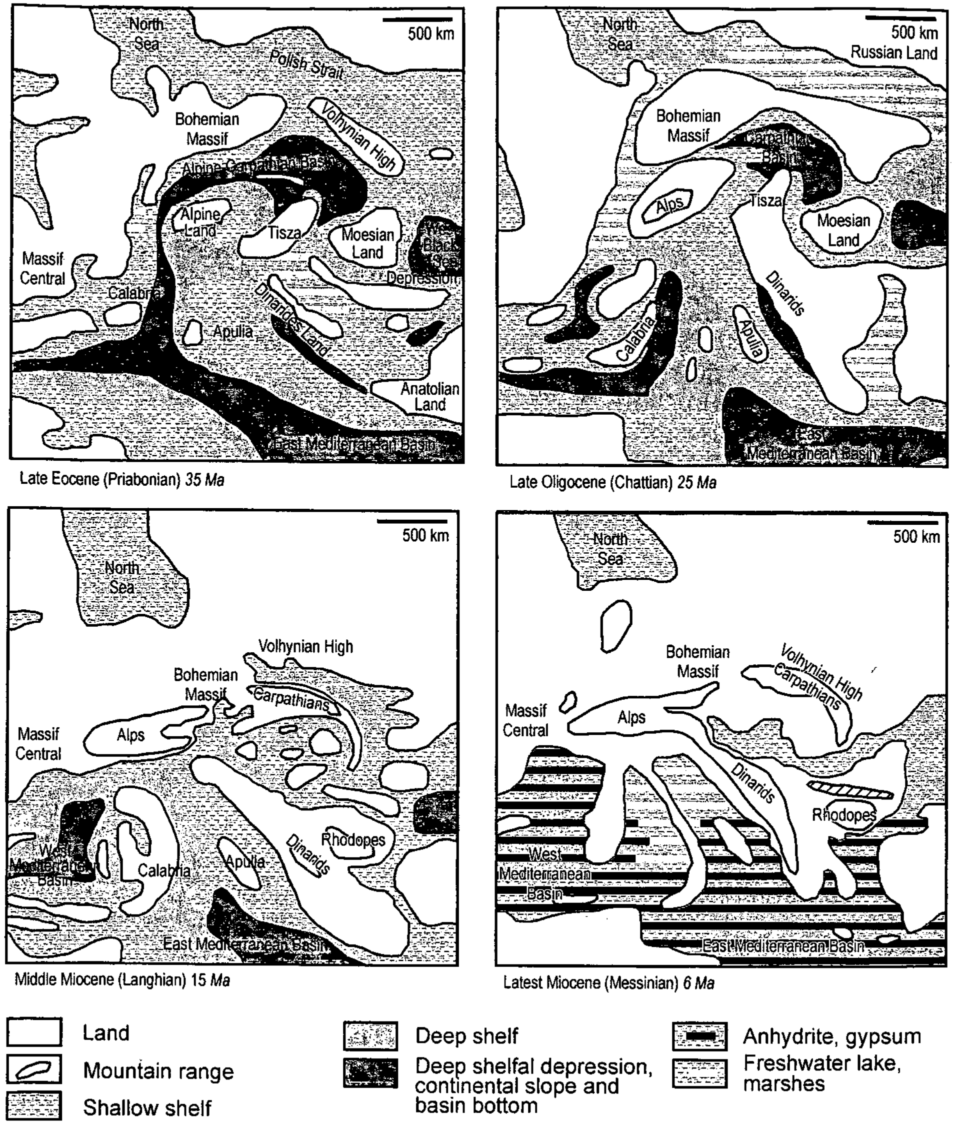


Fig. 17.1. Palaeogene–Neogene palaeogeography of Central Europe (after Popov *et al.* 2004).

Early Eocene (Early Eocene Climate Optimum). A gradual decrease in temperature during the later Eocene culminated in the formation of the first icesheets in Antarctica around the Eocene–Oligocene boundary (Zachos *et al.* 2001; Prothero *et al.* 2003). A renewed warming trend that began during the Late Oligocene continued into the Middle Miocene with a climax at the Mid-Miocene Climatic Optimum. The turning point at around 14.2 Ma led to the onset of the Middle Miocene Climate

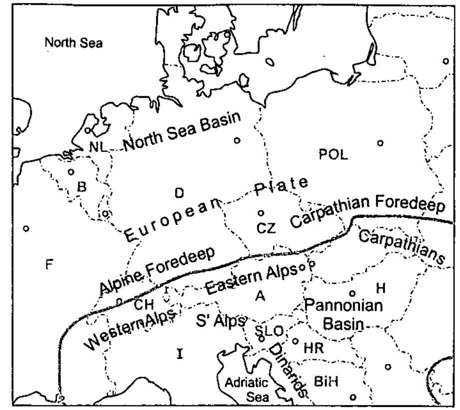


Fig. 17.2. Main physiographic and geological units of Central Europe. Grey line marks the border between the European Plate and the Alpine–Carpathian orogenic system. Abbreviations correspond to official country code plates.

Transition indicated by the cooling of surface waters and the expansion of the East Antarctic icesheet (Shevenell *et al.* 2004). A final trend reversal during the Early Pliocene is reflected by a gentle warming until 3.2 Ma (Zachos *et al.* 2001) when the onset of permanent Arctic glaciation heralded the Pleistocene ice ages (see Litt *et al.* 2008).

The Cenozoic history of Central Europe is chronicled in a dense pattern of Palaeogene and Neogene basins. In addition to the more stable North Sea Basin, the majority of these basins were strongly influenced by the Alpine compressive tectonics which caused a general uplift of Europe during the Cenozoic (see Froitzheim *et al.* 2008; Reicherter *et al.* 2008). The marginal position of the seas covering the area and the considerable synsedimentary geodynamic control resulted in incomplete stratigraphic sequences with frequent unconformities, erosional surfaces and depositional gaps.

This chapter deals with the Paleogene and Neogene ("Tertiary") geological development of Central Europe and its adjacent areas. It is structured according to the main geological regions relevant for the Cenozoic: (1) The European Plate; (2) the Alps and Alpine Foredeep; (3) the Carpathians, their foredeep and the Pannonian Basins System; and (4) the Southern Alps and Dinarides. Each subchapter is arranged from west to east, and north to south.

Palaeomagnetism and palaeogeography (E.M.)

Central Europe is composed of two tectonically contrasting regions: a northern Variscian region and a southern Alpine region (Figs 17.2 & 17.3). The former is called stable Europe in the palaeomagnetic literature, since the generally accepted view is that Europe, north of the Alpine front, behaved as a rigid plate during the Cenozoic Era. However, the number of palaeomagnetic results supporting this view is surprisingly few and most of them were acquired before 1980. The scarcity of good data is especially notable for stable Central Europe. The situation is best characterized by data sets used for calculating stable European Cenozoic palaeomagnetic poles by Van der Voo (1993) who included a single result from stable Central Europe, or more recently by Besse & Courtillot (2002, 2003) who used five (all representing the 16.5–34 Ma time interval). Thus, the Cenozoic apparent polar wander curve (APWC), which serves to represent the displacements of stable Europe with respect to the present-day north and its shift in latitude, is a synthetic one computed by using palaeomagnetic data from all continents and transferred to stable Europe through a plate tectonic reconstruction model (e.g. Besse & Courtillot 2002).

Palaeomagnetic directions computed from the synthetic APWC suggest that stable Europe travelled a few hundred kilometres northward during the Cenozoic, but did not change its orientation (within the limit of the resolution of palaeomagnetic observations, which in this case is *c.* ±5°). This implies that the expected stable European declinations are more or less aligned with the present meridians. The actually measured declinations at several points in stable Europe (e.g. France, Germany, Poland, Czech Republic; Global Palaeomagnetic Database 2005) deviate more from the present meridian than the coeval declination computed from the synthetic APWC. Nonetheless, these deviations cannot be interpreted in terms of tectonics, since the studied rocks, without exception, are igneous rocks, where secular variation of the Earth's magnetic field and poor control on local tectonics can cause bias from the direction of the geomagnetic field.

The stable European pattern of Cenozoic declinations breaks down as soon as we enter the North Alpine Foreland Basin at the southern margin of stable Europe and south of this. The pattern here suggests large-scale movements during the Cenozoic Era, accompanied by counterclockwise (CCW) rotations of variable angles and timings; observations for dominant clockwise rotations are known only from the Apuseni Mountains (Romania) and from the area between the Periadriatic and Sava faults.

Sediments lying directly on the stable European margin exhibit moderate CCW rotation (Scholger & Stingl 2004; Márton *et al.* 2003*a*). The mean angle of rotation is the same in every segment (Fig. 17.3), which is remarkable, since the age of the studied sediments varies between 20 and 7 Ma. The termination of the rotations is not constrained in any of the segments. Therefore, there are alternative solutions to explain these observations. One is that the rotations were simultaneous between longitudes of 12°E and 23°E. In this case, the Bohemian Massif (which is considered to be part of stable Europe) must have been involved in rotation. The other possibility is that the rotations are not coeval and the sediments were detached from the stable European basement. In either case, the observations would require dramatic modification of existing tectonic models.

As we move into the mobile part of Europe, the palaeomagnetic picture becomes fairly complicated. During the Palaeogene, there is no palaeomagnetic evidence for mobility, apart from a general northward travel for all tectonic units (Márton 2001). Rotations begin in the Miocene. Some of these can be related to tectonic escape from the East Alpine realm, others to subduction pull in the Carpathians and the rest to the motion of the Adriatic microplate (Adria).

Notable cases with rotations related to escape tectonics are known from the Eastern Alps and from the Periadriatic-Sava fault system (Fig. 17.3B, areas 10 and 11). In the first case, *c.* 30° of the CCW rotation observed in the sediments of the Miocene intramontane basins (Fig. 17.3B, area 10) are related to lateral extrusion (Márton *et al.* 2000a); in the second (Fig. 17.3B, area 11), the non-uniform, but dominantly clockwise rotations are attributed to right-lateral strike-slip movements (Fodor *et al.* 1998).

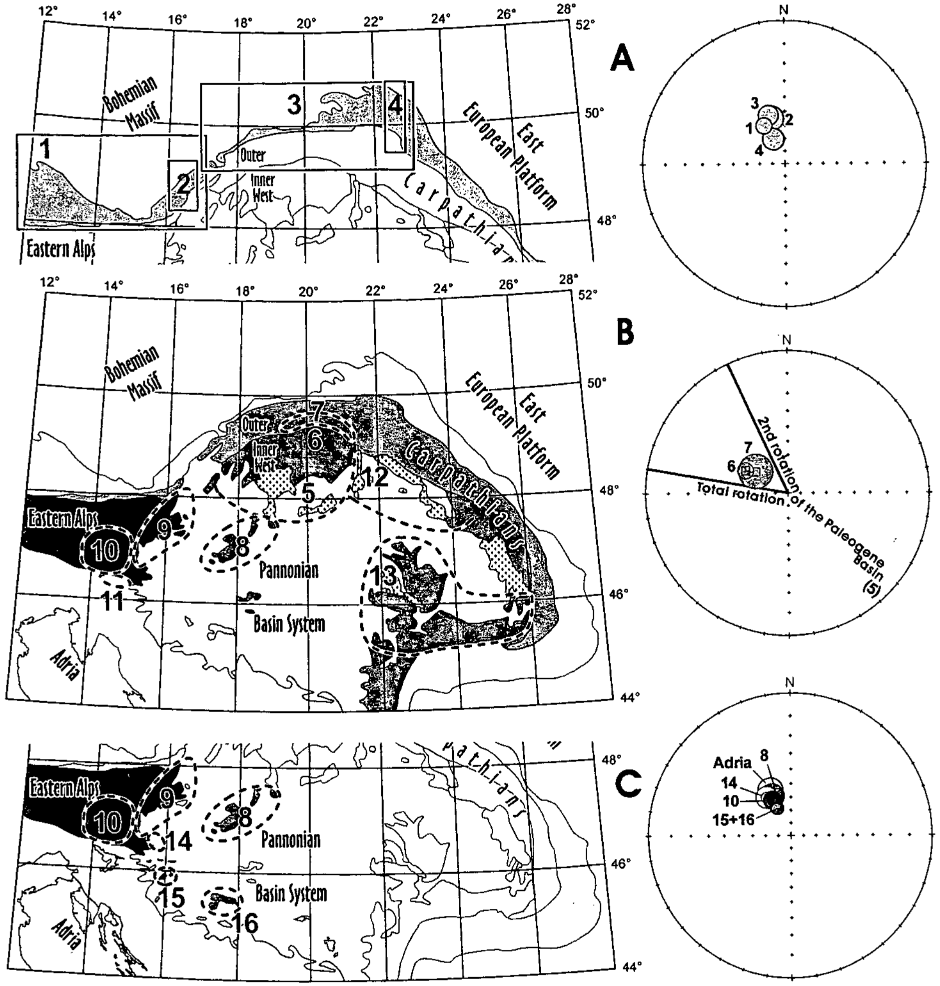


Fig. 17.3. Palaeomagnetics. **(A)** Alpine and Carpathian Foredeep. Age of sediments of the four numbered segments from west to east are: 27.5–17.2 Ma, 17.2-16.4 Ma, 16.4-13.0 Ma, 13.0-10.0 Ma (for Paratethyan geological stages refer to Rögl 1996). The stereonet (right side) shows the mean palaeomagnetic directions (declination and inclination) for the segments. The size of the dots corresponds to the size of the confidence circle for each palaeomagnetic direction. **(B)** Eastern Alps - Carpathians - Pannonian Basin. Numbered tectonic units are characterized by regionally coherent rotations, which are due to tectonic escape or subduction pull. Rotations connected to tectonic escape are the first-phase Miocene rotations in area 10 (intramontane basins of the Eastern Alps) and the rotations in area 11 (Periadriatic-Sava fault system). Rotations connected to subduction pull are those observed for areas 5 (North Hungarian-South Slovakian Paleogene basin), 6 (Central West Carpathian Paleogene Basin), 7 (Magura unit of the Outer Western Carpathians), 8 (Transdanubian Range), 12 (East Slovak basin), 13 (Apuseni Mountains). The stereonet shows the angle of the total rotation and that of the younger rotation for area 5, and the palaeomagnetic direction (declination and inclination) for areas 6 and 7. **(C)** Eastern Alps, Western Pannonian Basin. Rotations triggered by rotating Adria are observed for areas 8, 9 (border zone of the Eastern Alps and the Pannonian Basin), 10, 14 (Mura-Zala depression), 15 (Medvednica-Hrvatsko Zagorje), 16 (Slavonian inselbergs). The stereonet on the right side compares the palaeomagnetic direction characterizing the post-Eocene rotation of Adria with those reflecting the youngest rotation of the above listed areas.

Subduction (slab) pull accounts for most of the rotations observed for the Carpathians and for the Pannonian Basin. These rotations occurred during 18.5–17.5 Ma and 16–14.5 Ma (Magura unit of the Outer Western Carpathians (Márton *et al.* 2000*c*); Central West Carpathian Flysch basins (Márton *et al.* 1999*a*); North Hungarian–South Slovakian Palaeogene basin (Márton & Márton 1996; Márton *et al.* 1996; Márton & Pécskay 1998); Transdanubian Range (Márton & Fodor 2003)), close to the Palaeogene–Neogene boundary (e.g. Apuseni Mountains; Panaiotu 1998) and at 14–12 Ma (e.g. East Slovak Basin (Márton *et al.* 2000*b*) and Apuseni Mountains; (Panaiotu 1998)). The resulting rotations are CCW, except in the Apuseni Mountains.

As soon as subduction terminated in the Western Carpathians, at *c.* 14 Ma, a push from Adria and, subsequently, rotation of Adria with respect to Africa (Márton *et al.* 2003*b*) became the dominant driving force for displacements in the Alpine realm. A series of palaeomagnetic observations from the intramontane basins of the Eastern Alps (Márton *et al.* 2000*a*), from the Mura depression (Márton *et al.* 2002*a*) and from Northern Croatia (Márton *et al.* 2002*b*) indicate that *c.* 25° of CCW rotation occurred in the circum-Adriatic region, close to the Miocene–Pliocene boundary (Márton 2005). Moderate CCW rotations observed at the eastern margin of the Eastern Alps and in the bordering basins (Styrian and Vienna basins; Scholger & Stingl 2003, 2004; Scholger *et al.* 2003) and the youngest rotation of the Transdanubian Range (Márton & Fodor 2003) were probably also induced by the rotation of Adria.

Rotations affecting the same tectonic unit are of a cumulative nature. Repeated rotations in the same sense can cause large declination deviation from the present north. The largest declination deviations in the subject area were measured on Lower Miocene and Palaeogene rocks in the North Hungarian–South Slovakian Palaeogene basin (Fig. 17.3B, area 5), in the Transdanubian Range (Fig. 17.3B, area 8) and in the Apuseni Mountains (Fig. 17.3B, area 13), up to 90° in the counterclockwise and in the clockwise sense, respectively. Middle and Late Miocene rocks from the same areas are characterized by moderate rotations (*c.* 25–40°).

Age control on the termination of the youngest rotations is lacking, except for the North Hungarian–South Slovakian Palaeogene basin (14.5 Ma) and for the Apuseni Mountains (12 Ma). Thus, the possibility of large-scale neotectonic movements remains to be explored, especially in the southernmost part of Europe.

The European Plate: overview (M.W.R.)

The European Plate represented the so-called stable European continent during the Cenozoic northward drift of Africa. It comprises a large epicontinental sedimentation area extending from NW Europe towards Ukraine (Fig. 17.4). During most of the Cenozoic it was separated from the Alpine-Carpathian chain by the Alpine-Carpathian Foreland Basin – often referred to as the Molasse Zone or Molasse Basin – and its precursors as a part of the Palaeogene Tethys, or Oligocene–Miocene Paratethys.

The North Sea Basin (NSB) is a main element of the European Plate. It comprises a belt of Cenozoic deposits in the northern part of Central Europe, extending from England to northern Poland. The Polish Lowlands form its eastward continuation and served as an occasional connection between the NSB and the Eastern Paratethys. Further eastward lies the Volhyno-Podolian Plate forming the SW margin of the East European (= Russian) Platform. Towards the south, the NSB is bordered by the Belgian and German low mountain ranges as well as by the Bohemian Massif.

The Upper Rhine Graben represents an approximately 1100 km long structural element between the shores of the North Sea and the western Mediterranean. It formed a complex

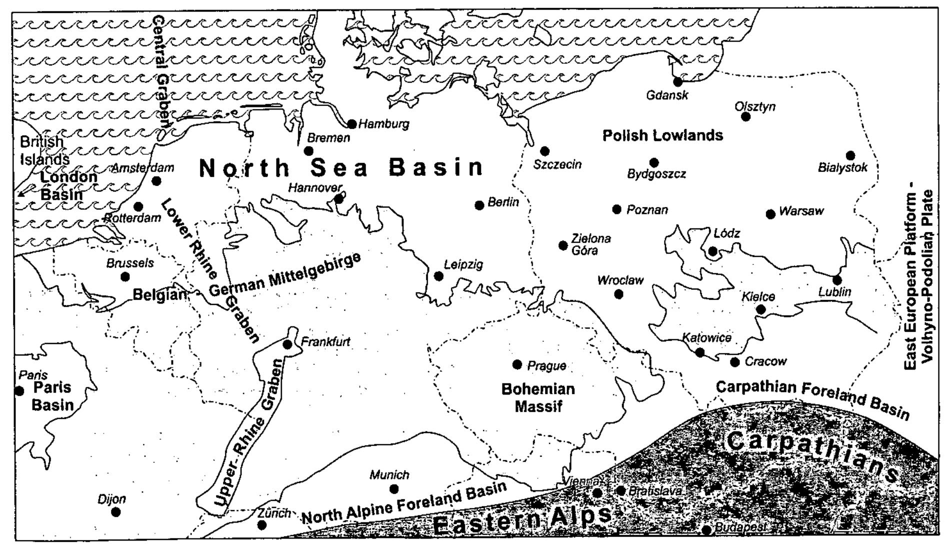


Fig. 17.4. Main Palaeogene–Neogene sedimentary basins of the European Plate.

Cenozoic rift system with sedimentation commencing during the Middle Eocene.

During the Paleocene and Eocene, the Helvetic units formed the southern margin of the European Plate, i.e. the northern margin of the Tethys Sea. Towards the south, the Helvetic units pass into the Ultrahelvetic Unit and the Rhenodanubian (Penninic) Flysch Zone. In late Palaeogene times, however, the Helvetic units are also considered by several authors to have constituted a transition to Ultrahelvetic and Penninic deep-marine clastic sedimentation as well as to the Northern Alpine Foreland Basin. Therefore, the Ultrahelvetic Unit is considered in both this and the following section on the Alps and Alpine Foredeep.

North Sea Basin: Palaeogene (K.G., F.W., G.S.)

The Cenozoic North Sea Basin (NSB) was at its most extensive in the Early Eocene, where it comprised the present-day North Sea as well as southern and eastern England, eastern Scotland, Belgium, the Netherlands, northern and eastern Germany, southern Sweden, Denmark, and northern Poland, and the Samland coast of Russia (Fig. 17.4). During the Palaeogene the NSB was a mainly marine semi-enclosed basin (Fig. 17.1) with paralic margins in the SE. It was separated from the Atlantic Ocean by the Thule landbridge at least until the Late Oligocene (Wold 1995). This volcanic barrier resulted from a magma chamber forming at 70 Ma showing at least four magmatic pulses until late Middle Eocene (O'Connor *et al.* 2000).

The NSB was the successor of the German Zechstein Basin, which developed as a result of Permian rifting (see McCann *et al.* 2008). It is bordered by the Belgian and German Uplands (= Mittelgebirge) to the south, by the Fennoscandian Shield to the north and Great Britain to the west. To the east, the NSB was episodically connected to the Eastern Paratethys via the Polish Lowlands. The palaeogeographical limitation to the north is unclear since the Fennoscandian Shield underwent significant uplift (up to 3000 m) from Oligocene times onward. The sediment cover of the shield was subsequently eroded, providing the main source of the NSB sediments younger than Oligocene (Overeem 2002).

Tectonic setting

The NSB is subdivided by a series of fault zones that originated in the Mesozoic rift phase (see Michelsen & Nielsen 1993; Kockel 1995; Kockel *et al.* 1996; Mulder *et al.* 2003). The most active of these is the Central Graben which trends in a north–south direction, curving to a SE direction in the offshore Netherlands area where it connects with the Lower Rhine Valley. It is connected to the Upper Rhine Graben via faults through the Rhenish Massif. Additional north–south trending fault zones, such as the Horn Graben and the Brande Trough within the Ringkøbing-Fyn High, the Sieverstedt Fault Zone, the Glückstadt Graben and the Hessian Depression, also terminate in the Upper Rhine Graben. Large areas of the basin are underlain by thick Zechstein salt deposits which influenced the sedimentation history of the basin (Weber 1977).

Sedimentary and stratigraphic development

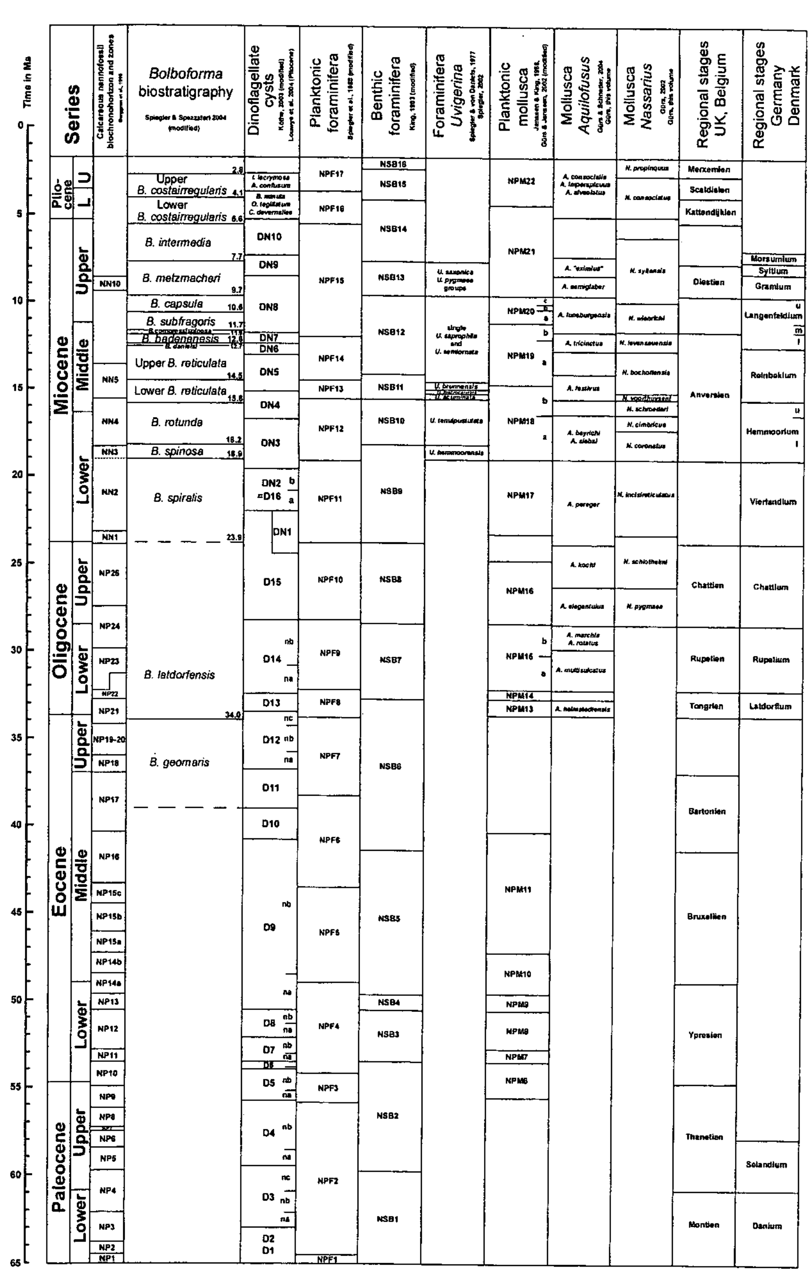
The Palaeogene development of the NSB can be divided into five phases (Figs 17.5, 17.6 & 17.7): the Early–Middle Palaeocene was a continuation of the tropical settings of the Cretaceous, as witnessed by widespread carbonate platform deposits in the entire NSB. It was also a major regressive phase. During the Early–Middle Palaeocene, carbonates were partly deposited in very shallow water as the Houthem, Ciply and Mons formations of Belgium. The Hückelhoven Formation of the Lower Rhine area, the Gödringen sands south of Hannover, and the Wülpen and Wasmansdorf formations of Brandenburg (Eastern Germany) represent clastic sedimentation generally with a high carbonate content (Lotsch *et al.* 1969). In the central part of the basin deep-water carbonates with *Lophelia* reefs (Bernecker & Weidlich 1990) or bryozoan limestones developed. During the Middle Palaeocene a dramatic sea-level fall led to paralic sedimentation spreading into the basin centre.

The Late Palaeocene to Early Eocene was a time of dramatic change. With the Thanetian transgression, the NSB sedimentation regime changed from predominantly calcareous to clastic. Clays and silts, partly with a carbonaceous microfauna (including agglutinated foraminifera and other non-carbonate microfossils) indicate a deep-water environment (e.g. Ølst Formation, Basbek Formation, Helle Formation). The macrofauna from the shallow-water deposits of the Thanet sands of Kent (UK) and Copenhagen (Denmark) indicates a colder climate. The Late Palaeocene Thermal Maximum is not reflected in the fauna of the NSB. Paralic or continental sedimentation is very rare at the southern margin of the basin apart from the estuarine Helmstadt-Halle embayment in Sachsen-Anhalt (Blumenstengel & Krutzsch 2008). The Thanetian transgression is also expressed in Belgium (e.g. Gelinden Formation). The overlying Landen Group begins with shallow-marine sands which were deeply eroded by the following continental Landen deposits. These are terrestrial in the south and lagoonal-brackish in the north (Tienen Formation). The regressive succession appears to be related to the uplift of the Artois axis rather than a sea-level fall within the basin (Dupuis *et al.* 1984). The Landen Group is covered by the partly deeper marine Ieper Group of Early Eocene age.

The marine diatomites of the uppermost Palaeocene to lowermost Eocene Fur Formation of northern Denmark (Homann 1991) are rich in vertebrate remains. They contain numerous (>200) ash layers and probably represent outer neritic depositional settings indicating upwelling on the shelf edge (Bonde 1979). The fish fauna suggests subtropical conditions.

The Eocene commenced with a major transgression that submerged large parts of the NSB. Depositional environments became dominated by dark clays of outer neritic to bathyal waters with ash layers from the Thule volcanism. These clays are rich in macro- and microfauna in the London and Hampshire basins and Belgium, but further to the east are almost unfossiliferous with rare agglutinating foraminifera and pyritized diatoms (*Coscinodiscus*) ('Untereozän 1 and 2', Røsnaes Formation, Zerben Formation). For these clays the term London Clay was originally used across the basin but has now been replaced by

Fig. 17.5. Chronostratigraphy and biostratigraphy of the North Sea Basin. The *Bolboforma rotunda*–Lower *Bolboforma reticulata* Zone boundary was lowered according to the match of the NN4–NN5 boundary with the *Uvigerina tenuipustulata*–*Uvigerina acuminata* Zone boundary (C. Müller, pers. comm.). The DN7–DN8 boundary and the NPF14–NPF15 boundary were lowered according to recalibration of ODP drillings in the North Atlantic by Müller & Spiegler (1993) and according to Gürs & Spiegler (2000). The NSB14 lower boundary is lowered due to the intra- Late Miocene age of the top of NSB13 (= upper boundary of the Gramian; pers. obs.). NPM18 is subdivided into a and b in contrast to Gürs & Janssen (2002), who named the upper part of NPM18 as NPM18a. The Nassariid zonation of Gürs (2002) is extended by the Late Oligocene *Nassarius pygmaeus* Zone as well as the Pliocene *N. consociatus* and *N. propinquus* zones.



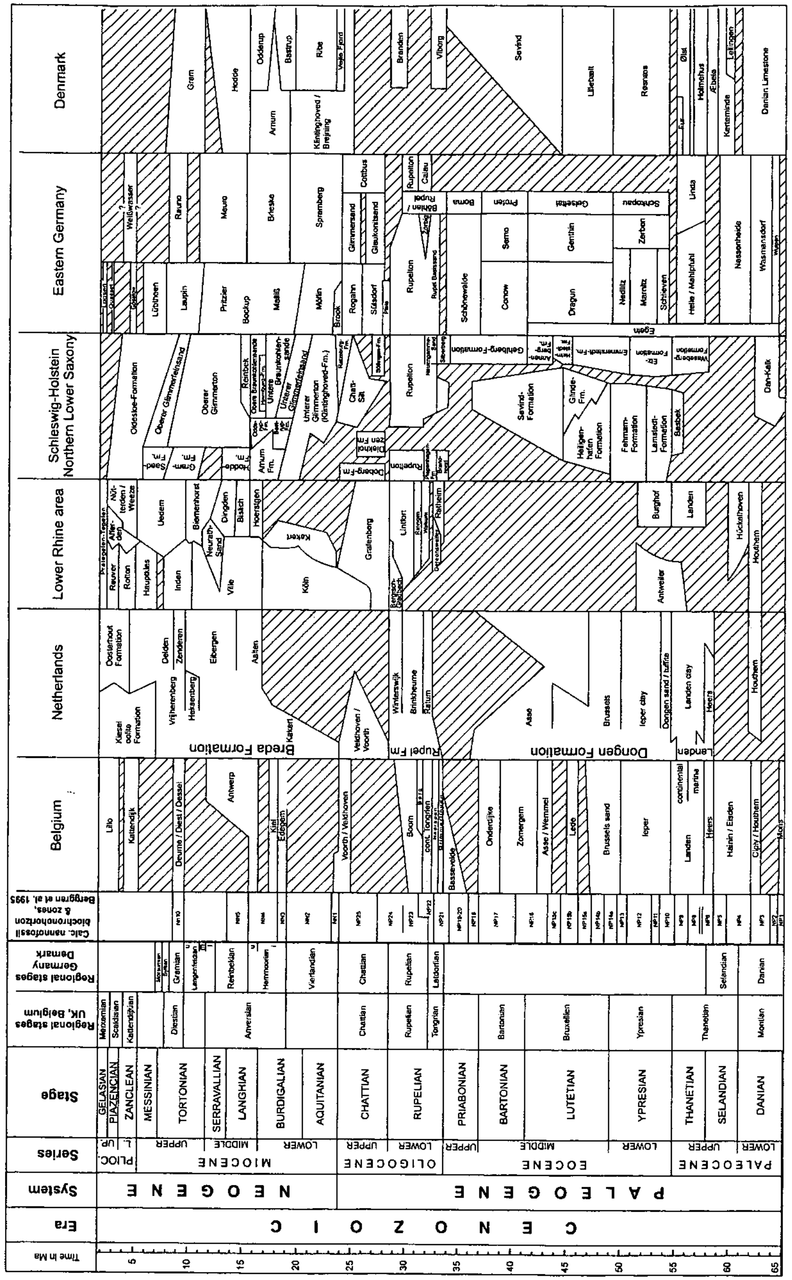


Fig. 17.6. Palaeogene–Neogene lithostratigraphic units of northern and eastern Germany.

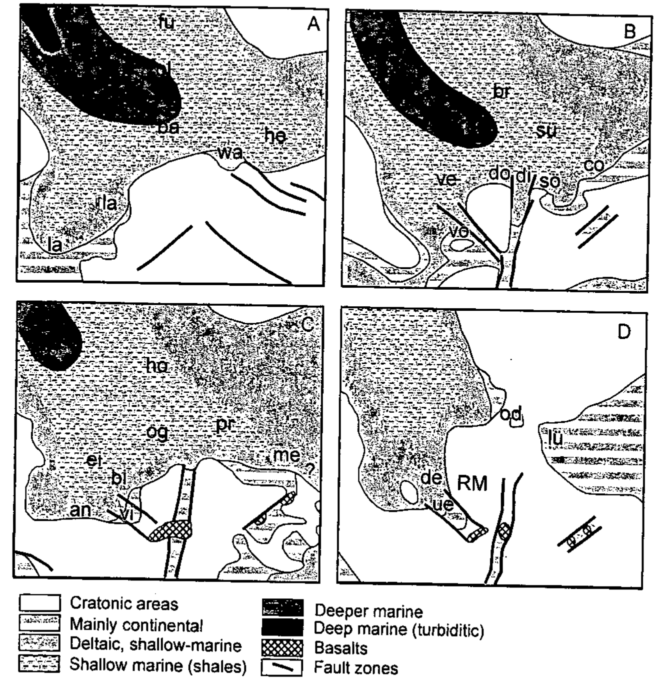


Fig. 17.7. Palaeogeography of the North Sea Basin: (**A**) Late Palaeocene; (**B**) Early Chattian; (**C**) Late Middle Miocene–Early Late Serravallian; (**D**) Early Messinian. Abbreviations: an, Antwerp Member (Mbr); ba, Basbek Formation (Fm); bi, Biemenhorst Fm; br, Brejning Fm; co, Cottbus Fm; de, Delden Mbr; di, Diekholzen Fm; do, Doberg Fm; ei, Eibergen Mbr; fu, Fur Fm; he, Helle Fm; ho, Hodde Fm; la, Landen Fm; lü, Lübtheen Fm; me, Meuro Fm; ne, Neureith Fm; od, Oldesloe Fm; og, Oberer Glimmerton; ol, Olst Fm; pr, Pritzier Fm; sö, Söllingen Fm; sü, Sülsdorf Fm; ue, Uedem Fm; ve, Veldhoven Fm; vi, Ville Fm; vo, Voorth Fm; wa, Waseberg Fm. Redrawn from the palaeogeographic Ziegler maps, modified according to new data of K. Gürs.

many regional formation names. In Germany and Denmark the overlying Early Eocene clays (Røsnaes Formation upper part, Fehmarn Formation) are intercalated by brown to reddish clays with a rich microfauna (Fehmarn and Lillebaelt formations). Water stratification, with reduced salinity in the surface layers, is inferred for the dark grey clay intervals deposited in Denmark (Schmitz *et al.* 1996). During deposition of reddish clays, water circulation was restored and a connection to the Atlantic, via the south of the UK, has been proposed by Bonde (1979).

Water depths during the Early Eocene were bathyal in the basin centre and outer neritic towards the south. Dill *et al.* (1996) demonstrated shallower environments in the area of Lower Saxony (Germany). However, their interpretation of the Late Palaeocene to Early Eocene sandy to clayey deposits, as tidal flats and marshes deposited in an estuary (Dill *et al.* 1996), is not followed here.

The Ypresian development of Belgium is one of the classic models of sequence stratigraphy (Vandenberghe *et al.* 1998). The Ieper Clay Formation reveals the strong cyclicity of outer neritic sedimentation, while in Late Ypresian time middle to inner neritic sands and clays indicate shallowing of the sedimentation area (e.g. Egemcappel clay, Egem sands, Merelbeke and Pittem clay).

During the Early Eocene, paralic conditions prevailed along the SE and southern basin margins (e.g. Saxony, Sachsen-Anhalt and Hannover area). In this environment, lignite seams interfinger with marine clays. Clays from the Hannover area have been radiometrically dated by Ahrendt *et al.* (1995) and range from 52.8 Ma to 46.0 Ma. Dinoflagellate cysts indicate zones D5b, D6b, D7a, D8nb and D9na, thus comprising nannoplankton zones NP10 to NP13/early NP14. The youngest unit (Emmerstedt Greensand) represents a substantial marine transgression dividing the lower and the upper seam complexes of this lignite district. In the Lower Rhine area, marine conditions predominated during the Early Eocene. However, the Antweiler Formation indicates paralic environments in the south of the region.

The climate of the Early Eocene in the NSB was subtropical to tropical (Krutzsch *et al.* 1992). Palm trees and mangroves bordered the sea which was inhabited by tropical faunas. Hooyberghs *et al.* (2002) postulated a climate optimum at the end of the Ypresian linked to the invasion of a southern fauna into the basin.

The Middle to Late Eocene witnessed a major sea-level fall and strong sea-level oscillations. A subtropical climate with rainforest vegetation (Mai 1995) and carbonate sedimentation in the shallower marine basin parts has been deduced from fossils.

Lignite formation in the south of the basin (Leipzig–Bitterfeld–Halle district) led to the development of economically important lagerstätten.

In the NE of the German part of the basin, a shelf edge developed at the Early–Middle Eocene transition (Heiligenhafen Formation, dated NP14; Martini 1991). The marked fossil contents (radiolaria and sponge spicules) of the slope clays and silts indicate upwelling ('Unter-Eozän 4'). The outer neritic sediments are replaced to the south and east by inner neritic, partly calcareous, glauconite sands and silts (Glinde Formation, upper part of Nedlitz Formation, Emmerstedt Formation).

During the successive Middle Eocene transgressions the bathyal Søvind marls were deposited over a much broader area than the Lillebaelt clays and neritic sedimentation shifted to the east, where carbonate or non-carbonate silts to fine sandstones of the Conow, Serno and Annenberg formations mark the shelf edge. These formations typically have an age of late NP15 to NP16 (Köthe 1988). Deposition of the Søvind marls continued with decreasing carbonate content until the Late Eocene (NP19/20). A decrease in carbonate content, as well as a coarsening-upward trend, can also be detected in the Upper Serno, Gehlberg and Lower Schönewalde formations. Thus, a more or less continuous shallowing- upward trend can be seen from Middle to Late Eocene times. During the Late Eocene, submarine erosion took place across vast areas of the eastern NSB. Only in the Kysing 1 borehole (Denmark) is a more or less continuous section across the Eocene–Oligocene boundary known (Heilmann-Clausen *et al.* 2001).

The Early Lutetian Brussel Sands of Brabant (Belgium) are marine incised valley fills showing a major hiatus with the underlying Ypresian clays (Vandenberghe *et al.* 1998). Many sand units disconformably overlie one another as a result of the strongly oscillating sea levels. The cyclic sedimentation patterns are also well preserved in the northern Belgian Kallo Complex where alternating glauconitic fine sands and clays can be traced over a vast area. The Late Eocene succession ends with the Bassevelde Sands dated to zone NP19/20 (Vandenberghe *et al.* 2003).

Middle Eocene paralic facies cover the famous Lutetian Geiseltal Lignite District (Germany), the Bartonian Leipzig–Bitterfeld–Halle Lignite District (Eissmann 1970; Eissmann & Litt 1994; Blumstengel *et al.* 1999; Rascher *et al.* 2005) and the Lutetian Helmstedt Formation. Late Eocene paralic areas include the Leipzig Lignite District (Standke 2002). The Helmstedt Formation is covered by the marine Middle Eocene Gehlberg Formation. The marine extension of the basin goes far to the NE where vast amounts of Baltic amber are now found in the marine clays, termed 'Blue Earths' of the Kalingrad area (Standke 1998).

The Oligocene began with a major transgression during NP21 which flooded the Egeln–Halle Lignite District, the Leipzig District (Standke 1997), and large parts of the German Mittelgebirge (Ritzkowski 1999*b*). After a short regressive phase (Vandenberghe *et al.* 2002) a minor transgression was observed in the NSB in zone NP22 (Gürs 2005), which flooded the Upper Rhine Valley via the Hessian Depression (Martini 1973). The most striking transgression took place in the early NP23 when the sea spread over vast areas of the low mountain ranges (Ritzkowski 1987) flooding the Leipzig Bay (Müller 1983; Eissmann & Litt 1994), the Upper Rhine Valley, the Lower Rhine Embayment and northern Belgium (Vandenberghe *et al.* 2002).

Neogene development of the NSB is already evident in the Late Oligocene following a severe regressive phase when the basin was clearly delimited to the east and river systems brought in vast amounts of sediment from the NE and east. Greater parts of the Fennoscandinavian Shield were lowlands and low mountains.

During the Chattian transgression, marine conditions were restored in large parts of the NSB. The NE part of Denmark and the Lower Rhine District were slowly covered by river systems and the coastline retreated inwards (Dybkjaer 2004*a*). Northern parts of eastern Germany were marine influenced throughout the entire Late Oligocene (Lotsch 1981; Standke 2001, 2008*a*). In the SW a semicontinuous connection to the Atlantic developed as shown by faunal evidence (R. Janssen 1979). Other pathways, such as the northern one around Scotland, have been discussed as the only connection (Gripp 1958) but single tropical molluscs such as *Morum*, *Perrona* and *Mitrolumna raulini septentrionalis* Janssen 1979 require a southern connection, the latter showing an affinity to the Chattian of the Aquitaine Basin. A warm climate led to the invasion of a nearly subtropical marine fauna. It is preserved in the Voort Sands of northern Belgium, the Grafenberg Sands of the Lower Rhine Embayment, the Kassel Sands (showing a strong ingression into the Hessian Depression in the early Late Oligocene) and the famous Sternberg erratic boulders of Mecklenburg (Bülow & Müller 2004) and Glaukonitsand in the Lusatia area (Standke 2006). Late Oligocene sediments in northern Belgium are restricted to the Roermond Graben and a small portion of the Central Graben area.

Deep water conditions prevailed in the basin centre. In the Late Chattian a coast with lagoons and barrier islands developed in middle Jutland and this was related to the Ringkøbing-Fyn High (Dybkjaer 2004*a*, *b*).

Significant stratigraphic boundaries

The famous Stevns Klint site in the eastern part of Denmark is situated where a continuous section across the Cretaceous–Palaeogene (K/Pg) boundary is developed (Birklund & Bromley 1979).

Volcanism

The Late Palaeocene and Early Eocene were times of significant volcanic activity at the Thule ridge. This volcanism provided the NSB with enormous amounts of volcanic ash. Ash layers can be traced as far south as northern Germany. Volcanism commenced in this region in the Late Cretaceous and ended in the late Middle Eocene (O'Connor *et al.* 2000). In the Middle Eocene volcanism began in the Eifel region (western Germany). On the SE margin of the NSB in the Erzgebirge and in the Lusatia (eastern Germany) areas, volcanism started in the lowermost part of the Oligocene (Standke & Suhr 1998).

North Sea Basin: Neogene (K.G., F.W., G.S.)

Tectonic setting and development

The Neogene was the time of infilling of the NSB. Tectonic development was controlled by intraplate stress regimes resulting from variable spreading rates along the Mid-Atlantic rift and Alpine compression. As a result, the NSB underwent basin subsidence, some inversion and salt tectonism (Clausen *et al.* 1999). Strong subsidence was restricted to the Central Graben, the Lower Rhine area, the Elbe Basin and rim sinks of younger salt structures. Salt tectonism was especially common in northern Germany (Weber 1977; Lange *et al.* 1990). Glacioeustacy-controlled tectonism occurred from 12.6 Ma onwards in the northern NSB and became widespread after the Piacenzian–Gelasian boundary (*c.* 2.5 Ma) (Overeem 2002).

The Thule landbridge drowned during the Late Oligocene and a permanent connection to the North Atlantic was established (Wold 1995). There is strong faunal evidence for a direct connection to the Atlantic in the SW for the entire Early Miocene to early Middle Miocene (Figs 17.1 & 17.7) (Janssen 2001; Janssen & Gürs 2002) and for the late Late Pliocene (Marquet 2004). The Dover Strait opened at 4 Ma (Van Vliet-Lanoë *et al.* 2002). A marine connection to the northern Paratethys cannot be entirely excluded for the late Middle Miocene as there are strong similarities in plankton successions (pteropods and bolboforms) with those of the Carpathian foredeep of the Central Paratethys (Gürs & Janssen 2002; Janssen & Zorn 1993). Fennoscandinavian uplift in the Oligocene narrowed the basin from the north. The centre of the shield experienced some 3000 m of uplift (Overeem 2002) and became a major sediment source for the NSB. Uplift of the Erzgebirge and the Lusatia Block (Brause 1990), as well as enlargement of catchment areas following the retreat of Paratethys in Central Europe, led to a more marked fluvial influence from the SE from Late Miocene times. Another main sediment source was the enlarged Rhine catchment in the south (Zagwijn & Hager 1987; Schäfer *et al.* 2004). Uplift of the Rhenish Massif accelerated in the Pliocene.

The bathyal water depths in the basin centre prevailed until the Early Pliocene. Subsequently the basin was mainly filled with eroded clastics from the uplifting hinterland.

Palaeomagnetism

Palaeomagnetic results from the area have been presented by Kuhlmann (2004) for the Pliocene of the central North Sea.

Sedimentary and stratigraphic development

The depositional system of the NSB during the Neogene was dominated by the westward progradation of at least three river systems from the Baltic Shield. These rivers were integrated in the Eridanos river system (Overeem 2002) which is a complex of rivers, whose number, source areas and courses varied considerably over time. Neogene development can be subdivided into four phases (Figs 17.5, 17.6 & 17.7).

During the Early to early Middle Miocene, major progradation of the Scandinavian and Baltic rivers took place. During the earliest Miocene a transgression led to the deposition of the lagoonal Vejle-Fjord clays and sands of Denmark (Dybkjaer 2004*a*, *b*; Rasmussen 2004). This lagoon was soon infilled by the fluvial Ribe sands. Marine conditions prevailed in southern Denmark and Schleswig (Klintinghoved Formation). In southeastern Holstein and Lower Saxony (northern Germany) the middle neritic Lower Mica Clay (Aquitanian to Burdigalian) overlies the Ratzeburg Formation of latest Chattian age (Hinsch 1986*a*, 1994). During the Aquitanian peat formation took place in the Lusatian Lignite District (Standke 2008*b*). Rich macrofloras from these lignites indicate warm climate conditions (Mai 1999, 2000). In the Lower Rhine Embayment the sediment load of the river system balanced subsidence.

A short interruption or slowing down of fluvial input occurred in the NE and eastern parts of the NSB, after which rapid infilling continued (e.g. Bastrup and Odderup sands; Rasmussen 2004). The Miocene transgression in northern Belgium commenced with the deposition of the Edegem sands in the Antwerp region in the Mid- to Late Burdigalian (Janssen 2001; Steurbaut & Verbeek 1988). A second transgression in the Antwerp area is represented by the Kiel Sands with a stratigraphic position between the Edegem Sands and the Antwerp Sands. At the end of the Early Miocene, sediment input from the east ceased and, at a time of low sea level around the Early–Middle Miocene boundary, a large brackish lake developed in the northern German Hamburg Trough (Hinsch 1988) represented by about 100 m of the Hamburg Clay.

This lake was replaced by the sea at the beginning of the Middle Miocene. Shortly afterwards an easterly derived fluvial system was once more established (Upper Lignite Sands). Coevally, the Moorken Seam developed in the Lower Rhine Embayment (Zagwijn & Hager 1987). In the Lusatian Lignite District the paralic and shallow-marine Brieske Formation was deposited (Standke *et al.* 2005; Standke 2006). This transgression is known as the 'Hemmoor transgression' (Anderson 1964) since the Hoerstgen Beds of northern Westphalia (upper Hemmoorian age) disconformably overlie Oligocene strata. The Miste Beds and the Breda Formation in the Peel region (the Netherlands) show the same transgressive history. In northern Belgium the Antwerp and Zondershot sands disconformably rest on older strata. The first could be dated as NN4 (Steurbaut & Verbeek 1988) and DN4 (Louwye *et al.* 2000*a*). The geometry of the prograding lower to middle Miocene sediment bodies indicates water depths of more than 400 m in the central North Sea.

During the early Middle to middle Late Miocene a major transgression restored marine conditions in the NSB. This major Miocene transgression (Reinbek transgression; Anderson 1964) began at the NN4–NN5 boundary (Spiegler 2002). In Denmark the deeper marine Hodde Clay (Rasmussen 1966) overlies the fluviatile Odderup Formation and extends to the NE beyond the Ringkøbing-Fyn High. To the east the predominantly marine Meuro Formation was deposited in Lusatia (Standke *et al.* 2005). To the south the transgression extended into the northern German low mountain range (Hinsch *et al.* 1978). The maximum flooding of this transgression occurred during the late Middle Miocene, when the marine Neurath sands were deposited in the Lower Rhine Embayment. At the same time the Eibergen and Biemenhorst beds, the anoxic Tostedt clays of northern Lower Saxony and southern Schleswig-Holstein, the glauconite bed between the Hodde and Gram formations in Denmark, and the 'Heller Horizont' (Bülow 2000) in Mecklenburg (11.8 Ma: Gürs & Spiegler 2000) were deposited. Deposits from this maximum flooding event have only recently been documented from the Antwerp sands (Louwye *et al.* 2000*a*). Despite the widespread flooding, basin inversion in the Channel region resulted in the simultaneous closure of the southwestern connection with the Atlantic Ocean. The exact timing of the maximum flooding is placed between 12.3 Ma (van Leeuwen 2001) and 11.8 Ma (Gürs & Spiegler 2000).

In Late Miocene times, a continuous sea-level fall (Michelsen *et al.* 1998) and lower subsidence rates in eastern Germany led to the emergence of vast areas. Sedimentation was limited to areas of increased subsidence. While in the early Late Miocene (Tortonian) large parts of Mecklenburg, Schleswig-Holstein, Lower Saxony and Denmark were covered by the sea, during the late Late Miocene (Messinian) the rivers bypassed these regions and sedimentation was displaced towards the delta fronts to the west. Burger (2001) distinguished three heavy mineral facies in the Nieder-Ochtenhausen borehole (Lower Saxony, Germany) belonging to different river systems with different source areas, of which only Norway with the Oslo fan can be specified. Uplift and deep erosion, with the formation of valleys of up to 100 m depth, has been described from northern Belgium (Vandenberghe *et al.* 1998). These structures are filled with the marine Diest and Dessel sands of early Late Miocene age. In the Antwerp area, the coeval Deurne Sands disconformably overlie the older strata.

In Early Pliocene times peat formation took place in the central part of the Danish North Sea Sector (Rasmussen 2004). Sediment infill in the Lower Rhine Embayment increased (Schäfer *et al.* 2004) and shallow-marine clastic environments prevailed across much of the Netherlands (e.g. Breda Formation; Sliggers & van Leeuwen 1987). However, in the northern part of the Netherlands the *Astarte trigonata* assemblages demonstrate a deeper environment, probably continuing until late Early Pliocene times. Early Pliocene sea-level highstands are represented by the Kattendijk Formation (northern Belgium), the uppermost Breda and basal Oosterhout formations (the Netherlands), and the Coralline Crag Formation (UK). Late Pliocene transgressions are manifested in the Lillo Formation (northern Belgium), the upper Oosterhout and the Maassluis formations (the Netherlands) and in the Red Crag Formation (UK). During the latest Pliocene the sea retreated from most of the basin and did not return to the southern and eastern margins of the NSB until the late Middle Pleistocene.

Evidence of glacial conditions, including the presence of deep permafrost and icebergs in the southern North Sea off the Dutch coast, have been recorded at marine isotope stages (MIS) 100–96 (Kuhlman 2004). These initial glacial intervals produced widespread deposition of fine-grained clastics, derived from enhanced erosion of the deep regolith that formed over the NW European continent in the warm-temperate climates of the earlier Neogene. In the latest Pliocene the rates of progradation and infill increased dramatically as a result of the increasing input of erosion products during glacial intervals. Three glacial and three interglacial periods are detected in northern Germany at this time (Litt *et al.* 2002).

Several biozonation schemes have been elaborated for the Neogene NSB. For open-marine settings, biozonations based on planktonic organisms (e.g. bolboforms: (Spiegler 1999), dinoflagellates (Louwye *et al.* 2000*a*, *b*; Köthe 2003) and pteropods (Gürs & Janssen 2002; Janssen & King 1988)) allow a fine-scaled subdivision of the strata. In nearshore marine settings, biozonation schemes are mainly based on benthic molluscs (Hinsch 1986*b*; Gürs & Weinbrecht 2001; Gürs 2002; Gürs & Schnetler 2004), benthic foraminifera (King 1982; Gramann & von Daniels 1988), bolboforms and dinoflagellates. For terrestrial and paralic deposits, pollen zonations (Krutzsch 2000*a*, *b*; Schäfer *et al.* 2004, and references therein) are available.

Volcanism

Some Neogene volcanism developed in the Lusatia and Erzgebirge regions from Early to later Miocene times (Suhr 2003). Early Miocene volcanism is known from the Hessian Depression (Kassel and Vogelsberg) and the Siebengebirge in Nordrhein-Westfalia. In Pliocene times the Eifel volcanism recommenced.

Polish Lowlands (M.Pi., J.R.K.)

The Polish Lowlands Basin (PLB) forms the eastward continuation of the North Sea Basin (Fig. 17.4). Towards the north it was bordered by the Fennoscandinavian Shield while to the south it was bordered by the Bohemian Massif including the Sudetes Mountains, and by the South Polish Uplands including the Holy Cross Mountains (Meta-Carpathians Rampart). Towards the east it was connected to the Ukrainian Massif.

Palaeogeography and tectonic setting

The Polish Lowlands are framed by two structural units separated by the Teisseyre-Tonquist lineament. The NE part of the Polish Lowlands belongs to the Precambrian European platform and the SW part to the Variscan Palaeozoic platform. Palaeogene and Neogene sediments developed on the Precambrian platform are rather thin, whereas slow continuous subsidence of the Palaeozoic platform led to the formation of a much thicker sedimentary unit during Palaeogene and Neogene times. In addition, some basins developed along major active structural lineaments and were later filled by thick continental deposits (up to 500 m).

During the Early Palaeocene, the PLB showed a marine connection with the Denmark Basin, the East European Basin, and with western Europe across northern Germany (Fig. 17.1). After a regressive phase, a Middle/Late Eocene transgression from the NW created a connection with cold oceanic waters across the Denmark area. At the same time, a connection with the Tethys was established. During the Oligocene, this connection ceased. Central and eastern parts of the PLB were connected with the Atlantic Ocean. The PLB was separated from the Ukrainian and Belorussian basins by an area of shallow shoals and islands. By the end of the Oligocene, brackish environments became widespread. The Neogene is characterized by increasing continental development with the establishment of large floodplains and lakes. Minor marine ingressions continued in the Pliocene.

Sedimentary and stratigraphic development

Neogene deposits are mainly continental, whereas the Palaeogene is mainly marine (Fig. 17.8). The onset of Palaeogene sedimentation, however, is not synchronous in the Polish Lowlands. In the eastern part it began during the Palaeocene, while in the west it commenced during the Middle Eocene or Late Eocene/Early Oligocene (Rupelian). The thickness of the Palaeogene deposits is *c.* 50–100 m but can be up to 300 m in basinal areas.

During the Palaeocene, marine deposition occurred across the entire NE part of the PLB and, locally (Szczecin Depression), in the NW. The Palaeocene PLB is littoral and shallow (<100 m), warm-temperate (17–20°C on average) with normal salinity. Palaeocene sediments mostly overlie Upper Cretaceous rocks (Maastrichtian, Campanian), separated by an erosional unconformity. Their mean thickness is 35 m, but areas with increased subsidence (Baltic Basin and Mazury High) can have successions of up to 100 m. The Danian Puławy Formation comprises marine sediments, including marls and limestones with intercalations of glauconite and phosphorite. Absolute dating gave an age of 60.2 ± 0.9 Ma (Pożaryska & Kreutzer 1978). Upper Palaeocene (Selandian, Thanetian) sediments include marls, sands and siltstones (Słodkowska 2000). The Thanetian Odra Formation contains brackish/continental lignitic sands, siltstones and clays. Authigenic deposits are weathered crust with residual clays. Kaolinite covers of plutonic and metamorphic rocks can be up to 74 m thick. These sediments are mostly abundant during the Palaeogene and rare during the Neogene.

The PLB Palaeocene can be correlated with the planktonic foraminifera zone NPF1, the benthonic foraminifera zone B1 and the nannoplankton zones NP1 and NP3 (Pożaryska & Szczechura 1968; Piwocki *et al.* 1996). Deposits from the Danian–Thanetian boundary and the Thanetian were dated using phytoplankton assemblage of the D3 and D4 zones (Słodkowska 2000), which are correlated with the NP4 and NP8 nannoplankton zones (Costa & Manum 1988).

Following a regression, marked by an erosional unconformity between the Palaeocene and the Eocene, the palaeogeography of the Polish Lowlands did not change substantially during Early Eocene times. A local transgression from the NW is restricted to the Szczecin Depression. A new transgressive influx from the Ukraine and subsequently from the NW created a connection with cold oceanic waters across the Denmark area. Another

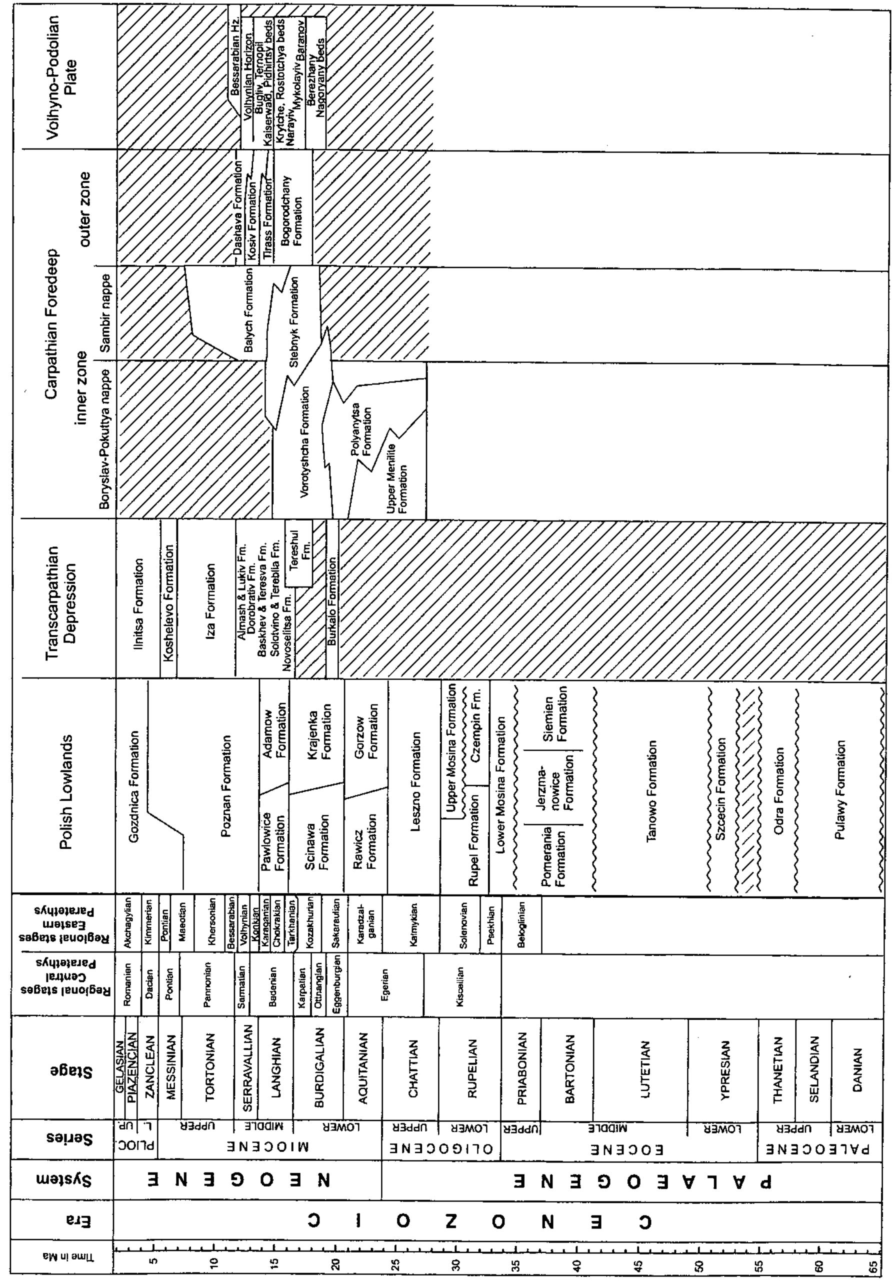


Fig. 17.8. Palaeogene–Neogene lithostratigraphic units of NE Europe.

transgression from the Tethys reached the Polish Lowlands from the west, and probably also from the south, during Middle/Late Eocene times. The pelagic Eocene basin was *c.* 100–150 m deep with cold-temperate and normal-salinity water. Average annual temperatures ranged between 16 and 20°C and salinity was not lower than 31–36‰. Around the Eocene–Oligocene boundary, amber-bearing deltas developed along the coasts (Piwocki & Olkowicz-Paprocka 1987; Kasiński & Tołkanowicz 1999).

The oldest Eocene deposits (Lower Eocene), belonging to the Szczecin Formation, include quartz-glauconite sands with gravel and clay intercalations (Ciuk 1973, 1974, 1975). The Middle Eocene is represented by the Tanowo Formation, consisting of lignitic clays, siltstones, and sands with thick lignite seams (Ciuk 1974, 1975). Sediments of the Pomerania Formation (in the north of the PLB) (Ciuk 1974) as well as its lateral equivalents, the Jerzmanowice Formation (west and central part) and the Siemień Formation (in the east) have a Middle/Late Eocene age. The 20-m-thick Pomerania Formation consists of quartz-glauconite sands in the lower part and calcareous sands, siltstones and claystones in the upper part. The deposits of the Jerzmanowice Formation consist of calcareous quartz-glauconite sands with gravel and phosphorite grains, and limestone intercalations within the uppermost part including molluscs, corals and bryozoans. The deposits of the Siemień Formation comprise calcareous quartz-glauconite sands and sandstones with phosphorites and calcareous siltstones, partly with a rich fauna of molluscs and corals as well as amber.

Planktonic foraminifera assemblages of zones NPF6 and NPF7 and the nannoplankton zones NP16–20 are documented (Piwocki *et al.* 1996). The benthic molluscan assemblages of the Pomerania and Siemień formations are located within the BM11A and BM13A zones (Piwocki 2002), which are characteristic for the Upper Eocene of NW Europe, and also for the Latdorfian facies (Hinsch *et al.* 1988).

Around the Eocene–Oligocene transition, a marine transgression encroached from the NW European Basin (Kockel 1988; Gramann & Kockel 1988) towards the Polish Lowlands (Piwocki & Kasiński 1995; Ciuk & Pożaryska 1982). The connection with the Tethys was severed during this time. A shallow epicontinental sea connected with the Atlantic Ocean covered the central and eastern parts of the Polish Lowlands. The seawater was relatively cold (10–20°C) with normal salinity, but slightly oxidizing conditions (Buchardt 1978; Odrzywolska-Bieńkowa *et al.* 1978). The Oligocene Polish Lowland sea was separated from the Ukrainian and Belorussian basins by an area of shoals and islands. The presence of quartz-glauconite sands records a regression at the end of the Early Oligocene. A subsequent short transgression extended as far as the west of the PLB. By the end of the Oligocene, the marine development ceased. The last regression was connected to the uplift of the Sudetes Mountains.

Oligocene deposits are common on the Polish Lowlands with thicknesses ranging from several metres up to >100 m. They commence with the Lower Mosina Formation comprising green quartz-glauconite sands, gravels, phosphorites, siderite concretions and rare amber grains (Ciuk 1970, 1974). The Eocene–Oligocene boundary is difficult to determine but may be placed within the lower Mosina Formation (Odrzywolska-Bieńkowa *et al.* 1981; Piwocki *et al.* 1985, 1996). The marine Rupelian (100 m) crops out in the western part of the Polish Lowlands. It consists of siltstones and claystones with sphaerosiderites and mollusc remains. Similar deposits are common in Germany and are named the Septaria Clays. These deposits interfinger with the brackish/continental Czempin Formation and the marine upper Mosina Formation towards the SE. The Czempin Formation (*c.* 25 m) comprises siltstones and silty sands with lignites. The deposits of the Upper Mosina Formation (*c.* 25 m) include quartz-glauconite sands with silt and clay intercalations. These deposits are overlain by the Leszno Formation in the west, which consists of silty sands.

Dating of the deposits is based on the occurrence of planktonic foraminifers of the Rupelian NPF8 zone and of the benthic B6 zone spanning the Early–Late Oligocene boundary. Nannoplankton assemblages of zones NP21–22 and NP24 have been documented.

Following the regression at the end of the Oligocene, continental and brackish sediments were deposited across the entire basin. This led to the development of widespread coastal plains related to the North Sea Basin (Standke *et al.* 1993). The continental deposits developed mostly as floodplains, meandering river systems (Osijuk 1979; Kasiński 1989) and in residual lakes with extensive lignite deposits.

Miocene lithostratigraphic units include the Lower Miocene Rawicz/Gorzów and Ścinawa/Krajenka, the Middle Miocene Adamów/Pawłowice, and the Upper Miocene/Lower Pliocene Poznań formations. These are continental lignite-bearing sediments of fluviatile and limnic facies. Most of the fluvial deposits were related to the southern tributaries of the Baltic fluvial system. Units of shallow marine and brackish deposits with glauconite and marine microfauna and phytoplankton occur in the western part of the PLB. Lower and Middle Miocene deposits comprise rather monotonous sandy/silty sediments with clay and lignite, while the Upper Miocene is represented by clays with silty/sandy intercalations. Some upper Middle Miocene deposits in SW Poland originated from a brackish marine embayment connected with the Paratethys (Łuczkowska & Dyjor 1971).

Pliocene deposits are represented by the upper part of the Poznań and the Gozdnica formations together with their lithostratigraphic equivalents (e.g the Ziębice Group; Dyjor 1966; Czerwonka & Krzyszkowski 2001). The Poznań Formation comprises clays and silts with sandy intercalations and thin lignites deposited mainly in fluvial (floodplain), lacustrine and limnic settings. Continental sedimentation was interrupted by marine ingressions, which periodically transformed nearshore lakes into brackish lagoons (Kasiński *et al.* 2002). The Pliocene deposits of the Poznań Formation were deposited under temperate and periodically cold-temperate and humid conditions (Stuchlik 1987). During sedimentation, the climate became increasingly arid. The Gozdnica Formation comprises sandy/gravelly clastic deposits with fine silt and clay intercalations developed under conditions comparable to the Poznań Formation.

Volcanism

The Lower Silesian volcanics comprise basic volcanic and pyroclastic rocks (up to 200 m) which belong to the eastern part of the Mid-European Volcanic Province. This extends over 700 km in front of the Alpine-Carpathian system. Most of the localities are situated on the Fore-Sudetic Block along the margin of the Sudetes Mountains. Volcanic rocks include basalts, trachytes, tephrites, basanites, phonolite basanites, quartz latites, phoideferous basalts, nephelinites, dolerites, basanitoides, limburgites, ankarites, trachyandesites and trachyphonolites (Wierzchołowski 1993).

Volhyno-Podolian Plate (O.Y.A., V.V.A.)

The Volhyno-Podolian Plate (VPP) (Fig. 17.4) represents the SW margin of the East European (= Russian) Platform. The most significant structures there are the Volhynian and Podolian highs which are oriented in a NW–SE direction between the upper parts of the Western Bug and the valley of the Southern Bug river. In the north a terrace zone of 30–50 m separates the Volhynian High from the Polessye Lowland.

Tectonic setting

The basement of the East European Platform in the western Ukraine consists of deformed Archean to Lower Proterozoic rocks. To the west the crystalline basement is covered with a thick platform nappe, comprising up to 2700 m of Mesozoic–Cenozoic sediments. From latest Early Miocene to latest Middle Miocene times, the Podolian-Subdniester Block subsided resulting in the formation of accommodation space. With the beginning of the late Middle Miocene the western part was uplifted (up to 450 m) resulting in the erosion of Middle Miocene deposits (Palienko 1990).

Sedimentary and stratigraphic development

Sedimentation of the VPP commenced during the late Early Miocene (Fig. 17.8). The basal Nagoryany beds (2 m) consist of quartz-glauconitic sand and sandstones, with a diverse marine mollusc fauna. They discordantly overlie Upper Cretaceous limestones. The Nagoryany beds are presumably of Karpatian age (= late Burdigalian). The discordantly overlying Berezhany beds (4.5 m) are freshwater limestones and marls, containing ostracods, characean algae and numerous freshwater gastropods.

The Middle Miocene (Badenian regional stage) begins with the Baranov Formation (1 m) comprising glauconitic sands and sandstones with a rich molluscan fauna as well as coralline algae. The Mykolayiv Formation (14.5 m) consists of quartz-glauconitic sandstones rich in bryozoans and serpulids, foraminifers, bivalves, brachiopods and rare echinoderms (Vialov 1986). The Narayiv Formation (25 m) represents the most continuous horizon. Dense corallinacean limestones with huge rodoliths, pectinids, oysters and an impoverished foraminifer fauna are typical (Vialov 1986). The Rostotchya Formation consists of quartz sands and sandstones with molluscs comparable to the Baranov Formation (Vialov 1986; Kulchytsky 1989). A thin layer (10–15 cm) of limestones or calcareous sandstones constitutes the Kryvtche Formation and represents an important marker horizon.

The upper Badenian Pidhirtsy Formation is restricted to the east of the VPP. It comprises sandstones with molluscs, foraminifers and bryozoans (Pishvanova *et al.* 1970; Vialov 1986). The upper Badenian Ternopil beds (6 m) consist of corallinaceans (mainly rhodoliths), grading into clayey glauconitic sand and limestone. The general character of the fauna is the same as in the underlying units, but additionally a biohermal limestone (Podolian Toltry or Miodobory), dominated by coralline algae and vermetids, is found (Vialov 1986).

The uppermost Badenian is represented by the Bugliv Formation (Grischkevich 1970) representing an offshore facies. It includes an impoverished marine fauna with the endemic bivalve *Parvivenus konkensis media* (Sokolow) and can be correlated with the Konkian regional stage in the eastern Paratethys (Vialov 1986).

The late Middle Miocene sediments of the Sarmatian regional stage are transgressive attaining a thickness of 25–100 m (Vialov 1986). In the NE of the VPP the Sarmatian can be clearly correlated with the Volhynian and Bessarabian regional substages of the eastern Paratethys (Grischkevich 1970; Pishvanova *et al.* 1970). The lower part of the the Sarmatian deposits consists of sand (1.2 m) and oolitic sandstone containing a rich endemic mollusc fauna. Up-section, a coarse-grained sand with a huge number of molluscs and foraminifers occurs (Pishvanova *et al.* 1970). In the east of the VPP, biohermal and oolitic limestones developed. The Early Sarmatian bryozoan bioherms cover the tops of the Badenian 'Toltry' limestone.

Upper Middle Miocene and lower Upper Miocene sediments (= Bessarbian regional substage) are restricted to the east of the VPP. Oolitic limestones, nubeculariid-foraminifera limestones and bryozoan bioherms formed parallel to the Badenian 'Toltry' ridge (Vialov 1986; Paramonova 1994; Anistratenko 2000; Anistratenko & Anistratenko 2005). The intensive uplift of the Carpathian Mountains during the latest Middle and early Late Miocene and tectonic reduction of the Subcarpathian Basin cut off the marine connections to the VPP.

Upper Rhine Graben (J.P.B.)

The Upper Rhine Graben (URG) is part of the complex Cenozoic rift system of western and Central Europe that extended from the shores of the North Sea over a distance of some 1100 km into the western Mediterranean (Fig. 17.4). The stratigraphic development of the URG varies in its southern and its northern part (Figs 17.9 & 17.10). Berger *et al.* (2005*a*) subdivided the southern URG into three different units: (1) Southern URG (Basel Horst, Dannemarie Basin, Mulhouse Horst, Mulhouse Basin potassique, Sierentz-Wollschwiller Basin, Rauracian Depression); (2) Southern-Middle URG (Colmar, Sélestat, Erstein, Zorn or Strasbourg Basin, Saverne Fault Zone, Ribeauvillé & Guebwiller); and (3) Northern-Middle URG (Haguenau, Pechelbronn, Rastatt, Karlsruhe, Landau, Bruchsal-Wiesloch Fault Zone). The North Upper Rhine Graben includes the Mainz and Hanau basins.

A general synthesis on the Upper Rhine Graben was recently published in the frame of the EUCOR-URGENT project ('Upper Rhine Graben Evolution and NeoTectonics', Behrmann *et al.* 2000). The stratigraphy and palaeogeography, together with a detailed reference list, were published by Berger *et al.* (2005*a*, *b*). This section presents an overview of these articles. Another study on the Rhine Graben examines the correlation of eustatic sea-level changes, rifting phases, palaeogeography and sedimentary evolution (Sissingh 2003).

Palaeogeography and tectonic setting

Palaeogeographic reconstructions of the URG were published by Kuhlemann & Kempf (2002), Sissingh (1998, 2003), Becker & Berger (2004) and Berger *et al.* (2005*b*). Following an initial fluviolacustrine phase of sedimentation (Middle Eocene), the URG was affected by an initial rifting phase that was responsible for the development of large conglomeratic fans along the eastern, western and southern graben margins. The axis of the evolving URG was occupied by salt basins in its southern areas with brackish (with local salt) to lacustrine facies developing in its central parts, and fluviatile to lacustrine facies in its northern part and in the Mainz Basin.

During the Early Rupelian, a general transgression came from the North Sea; following a regressive phase marked by lacustrine (north) and salt (south) deposition during the middle Rupelian, the sea again invaded the entire basin during the Late Rupelian, resulting in a possible connection with the Paratethys Sea via the North Alpine Foreland Basin. During the Late Oligocene, fluviolacustrine sedimentation prevailed, with localized marine ingression in the northern part of the URG.

During the Early Miocene, the southern URG was affected by uplift and erosion. In the northern URG, brackish (rarely marine) sedimentation prevailed. In the Late Burdigalian, the basin was

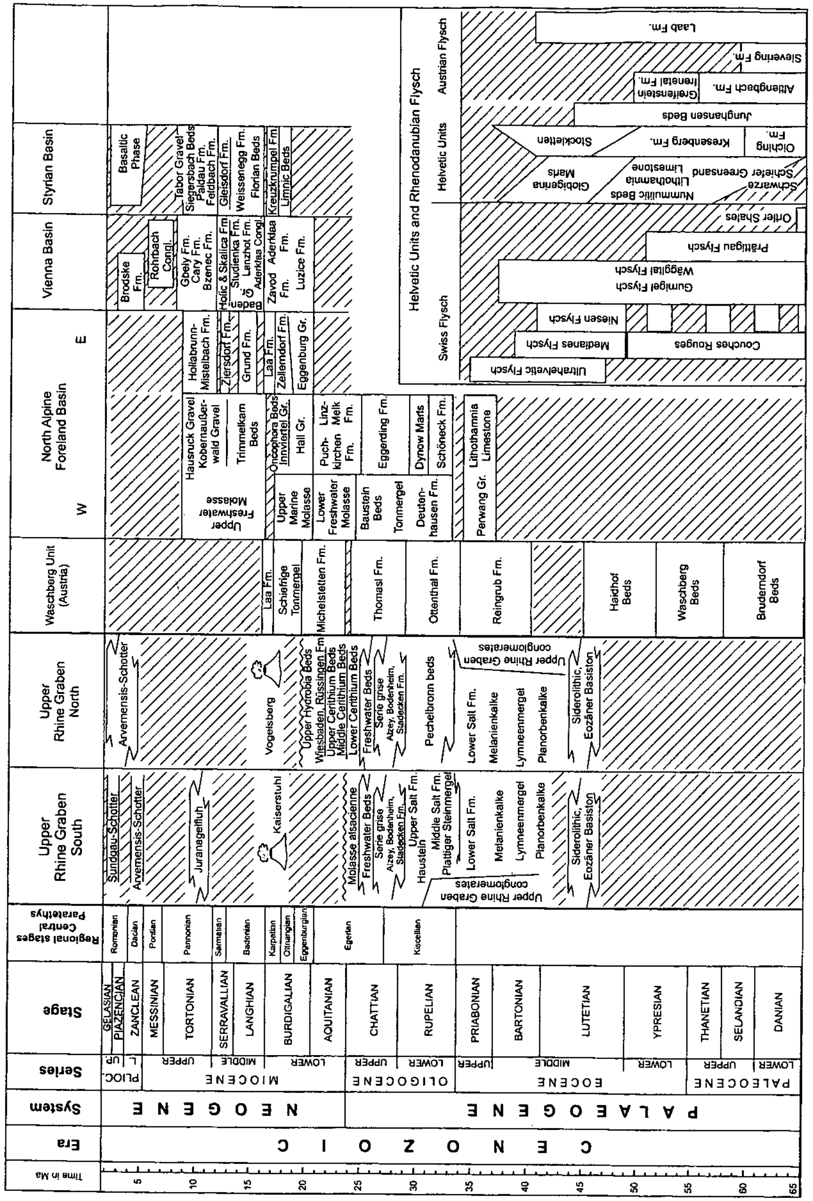


Fig. 17.9. Palaeogene–Neogene lithostratigraphic units of the southern part of the European Plate and the North Alpine Foreland Basin.

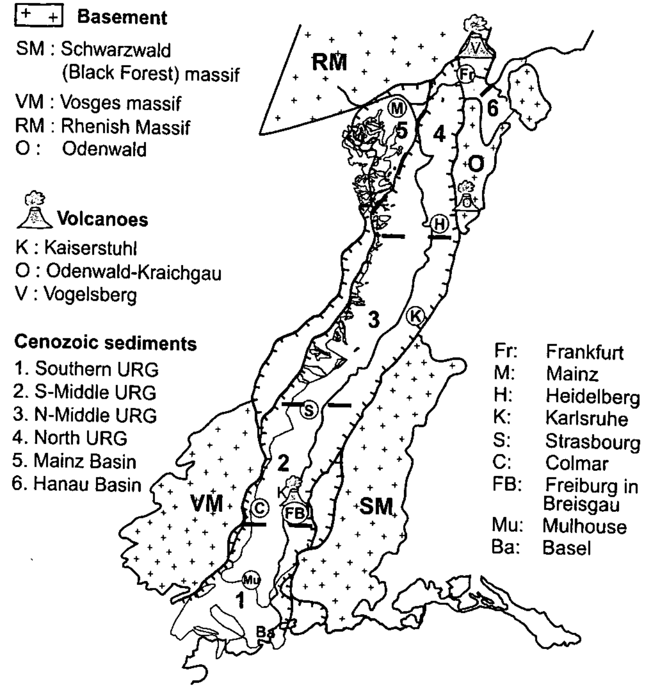


Fig. 17.10. Geology of the Upper Rhine Graben.

affected by two important volcanic events (Kaiserstuhl and Vogelsberg events).

Reflection seismic data, calibrated by wells, indicate that south of the city of Speyer a regional erosional unconformity of mid-Burdigalian age cuts progressively deeper southwards into Burdigalian and Oligocene sediments. This unconformity is related to the uplift of the Vosges–Black Forest arch, including the southern part of the URG (Roll 1979; Dèzes *et al.* 2004). During the Late Miocene, the sedimentation was essentially continental, brackish and lacustrine. In the Mainz Basin Langhian-age sediments are absent. In conjunction with progressive uplift of the Vosges–Black Forest arch, the central and southern parts of the URG were subjected to erosion.

Sedimentary and stratigraphic development

Eocene sedimentation in the southern Upper Rhine Graben generally commences with Middle to Late Eocene siderolithic deposits (MP19). These are overlain by various sedimentary facies whose deposition was influenced by tectonic processes in the URG. Accumulation of brackish sediments (Lymnaeenmergel) and salt (Lower Salt Formation) took place. Their ages probably range from the Lutetian to the Priabonian (NP12 to NP17). However, it is not clear if sedimentation was continuous or if a gap occurred during the Bartonian (Sissingh 1998). In the west, some undated conglomerates are tentatively assigned to the Lutetian and the Priabonian (Duringer 1988).

On the Mulhouse Horst, brackish marls and lacustrine limestones, probably ranging from Lutetian to Priabonian in age, are present. The Melanien Limestones are of Priabonian age (MP18) at Brunstatt (Schwarz 1997; Tobien 1987). In the eastern part of the basin, some conglomerates (locally dated by mammals (MP18; Tobien 1987)) probably accumulated during the Priabonian.

In the Colmar-Haguenau area a similar development occurs, and includes lacustrine limestones, Lymnaea Marls and salt formation. A brackish complex (Zone Dolomitique) generally overlies these deposits, but it has not been dated. Conglomeratic facies are present at both basin margins (particulary from the Priabonian onwards). Towards the north, the Lower Pechelbronn Beds are partly dated by mammals as Priabonian (MP20; Tobien 1987).

Early Rupelian sedimentation is represented by: (1) the Middle and Upper Salt Formation (Dannemarie Basin); (2) the Middle and Upper Pechelbronn beds (Colmar-Haguenau area) (NP22; Griessemer 2002); (3) the Zone fossilifère and the Streifige Mergel (= Marnes en plaquettes) as well as the Haustein – the latter is dated as NP21–22 (Schuler 1990) and MP21 (Altkirch (see Storni 2002; Hinsken 2003), Mulhouse Horst); and (4) the conglomeratic facies at both basin margins.

The Zone fossilifère and the Middle Pechelbronn Beds represent the first Rupelian transgression from the North Sea (partly corresponding with the global sea-level rise following the Ru1 sequence at 33 Ma; see Hardenbol *et al* 1998). In the Middle Rupelian, the most important marine transgression (second marine Rupelian transgression from the North Sea, corresponding with the global sea-level rise between sequences Ru2 and Ru3; see Hardenbol *et al.* 1998) flooded the entire URG. The sediments associated with this transgression constitute the classical succession from the Foraminiferenmergel to the Fischschiefer and partly to the Meletta-Schichten. These sediments range from NP23 to NP24 (Grimm 2002). They are combined with the Late Rupelian brackish-marine Cyrenenmergel into the so-called Graue Serie (= Série grise), representing an important seismic marker. During the Early Chattian, the fluviatile sediments of the 'Molasse alsacienne' interfingered to the north (Tüllinger Berg) with lacustrine or brackish marls and with the fluviolacustrine Niederroedener Beds. During the Middle and Late Chattian, the lacustrine limestones of the 'Calcaires delémontiens' and the Tüllinger Kalk were deposited. The 'Calcaires delémontiens' are well dated as MP29 to MN1 in the Jura Molasse (Becker 2003; Picot 2002), and the Tüllinger Kalk can be correlated by charophytes with the Middle Chattian *Stephanochara ungeri* Zone. These deposits are overlain by the brackish-lacustrine *Cerithium* Beds.

From the Aquitanian onwards, the southern and middle parts of the URG were subject to erosion. Sediments may have been deposited up to the onset of the uplift of the Vosges and the Black Forest (Burdigalian), or the non-deposition may already have commenced during Middle and Late Aquitanian times, as proposed for the Jura Molasse (see Picot 2002; Becker 2003). Sedimentation locally resumed during the Early Tortonian, with the deposition of the *Dinotherium* and *Hipparion* sands (dated by mammals as MN9). No Messinian sediments are known. In the northern part of the URG, sedimentation continues during the Aquitanian with the deposition of the brackish Upper *Cerithium* Beds, which are overlain by the Rüssingen Formation (formerly *Inflata* Beds), and are followed by the Wiesbaden Formation (formerly Lower *Hydrobia* Beds). According to its correlation with the Mainz Basin (MN1 and MN2; see Reichenbacher 2000), this succession appears to be Aquitanian in age. The Upper *Hydrobia* Beds are still present and may be attributed to the Early Burdigalian (MN3). During the Late Burdigalian an important volcanic event took place at Kaiserstuhl. Following a significant gap (Burdigalian to Messinian), sedimentation was resumed during the Pliocene.

An important sedimentary event is represented by the Sundgau Gravel, which may be Middle Pliocene (MN15–16) in age, based on comparison with the Bresse Graben mammal localities (Petit *et al.* 1996). Locally, the Arvernensis Gravel is present (MN14–MN16; Tobien 1988). Pliocene deposits (dated by spores and pollens) are also known from the vicinity of Colmar and Strasbourg. Following a gap (Burdigalian to Messinian), sedimentation in the northern part resumed in Pliocene times (e.g. Haguenau area).

During the Eocene and Early Oligocene, the northern Upper Rhine Graben was not yet developed and the main structures were influenced by pre-Palaeogene tectonics. From north to south, pre-rift structures include the Rüsselsheim Basin, the Palatinate-Stockstadt Ridge, and Marnheim Bay (Grimm & Grimm 2003). In the Rüsselsheim Basin, Marnheim Bay and the southerly connected northern URG, sedimentation commenced with the terrestrial 'Eozäner Basiston' and 'Basissand'. On the Palatinate-Stockstadt Ridge (particulary on its eastern part, i.e. the Sprendlingen Horst), the famous Messel Formation accumulated in a maar lake (Harms 2001). These sediments are dated by mammals as MP11 (Tobien 1988). The Priabonian Lower Pechelbronn Beds (Rüsselsheim Basin) and 'Green marls-Rote Leitschicht' (Marnheim Bay) overlie the Eocene basal sediments. The 'Green marls-Rote Leitschicht' correlate with the Lutetian *Embergeri* Zone and the Priabonian *Vectensis* Zone (Schwarz 1997). Along the margins of these basins, terrestrial alluvial-fan deposits accumulated during the Lutetian to Early Rupelian (Grimm & Grimm 2003). Eocene sediments are absent in the Hanau Basin.

During the Rupelian, the sedimentary history was similar to that of the Southern URG, with the deposition of the Middle Pechelbronn Beds (first marine Rupelian transgression), the Upper Pechelbronn Beds, the Alzey, Bodenheim, Stadecken and Sulzheim formations (second and third marine Rupelian transgression, corresponding with the 'Serie grise' in the Southern URG), the Niederroedener Beds and the brackish *Cerithium* beds (Grimm *et al.* 2000; Reichenbacher 2000; Grimm & Grimm 2003).

The Upper *Cerithium* Beds cross the Chattian–Aquitanian boundary. The brackish-lagoonal sediments of the Oberrad Formation (= upper part of the Upper *Cerithium* Beds; see Schäfer & Kadolsky 2002) are of Aquitanian age (MN1, MN2; Engesser *et al.* 1993; Försterling & Reichenbacher 2002). The overlying brackish-lacustrine marls and limestones of the Rüssingen Formation (= *Inflata* Beds) still belong to the Aquitanian (MN2a; Engesser *et al.* 1993). In the uppermost part of the Rüssingen Formation, biota indicate a further brackish-marine ingression from the south (Reichenbacher 2000), which was again followed by an ingression from the North Sea (Martini 1981). This ingression characterizes the base of the Wiesbaden Formation (= Lower *Hydrobia* Beds; see Reichenbacher & Keller 2002), which mainly consists of bituminous marls and limestones and is of Aquitanian age, except perhaps in its uppermost part.

During the Burdigalian, brackish sedimentation was progressively reduced and replaced by lacustrine and fluviatile deposits. During the Langhian and the Early Serravallian, sedimentation was essentially continental, brackish or lacustrine in the Northern URG and the Hanau Basin (e.g. Staden Formation, Bockenheim Formation; see Grimm & Hottenrott 2002). A plateau-basalt layer ('Maintrapp'), with a radiometric age of 16.3 Ma, is found between the Staden and Bockenheim formations (Fuhrmann & Lippolt 1987). In the Mainz Basin, no Upper Burdigalian to Upper Serravallian sediments are known. The presence of limnofluviatile Tortonian deposits is represented by the Lauterheim Formation, the *Dinotherium* sands and the Dorn Dürkheim Formation, which are dated at several localities within the Mainz Basin (MN9 and MN11). The presence of several unconformities (Lower Serravallian–Tortonian, Tortonian–Lower Messinian as well as below the Piacenzian Arvernensis Gravels) is still controversial; field and borehole observations indicate three breaks in sedimentation, whereas seismic lines do not show any unconformity during the Mio-Pliocene in the northern part of the URG (Dèzes *et al.* 2004).

Uppermost Miocene to Pliocene sediments are known from some areas of the Mainz Basin and the western part of the Northern URG. They are represented by the 'White Mio-Pliocene' and Bohnerz clays (Rothausen & Sonne 1984; Grimm & Grimm 2003). Piacenzian-age fluviatile sediments are represented by the Arvernensis Gravels and the Weisenau Sands of the northernmost URG and the Mainz Basin and dated by magnetostratigraphy and heavy mineral associations (Fromm 1986; Semmel 1983). Late Pliocene sediments probably exist in the Heidelberg and Frankfurt areas (Hottenrott *et al.* 1995).

Helvetic units: western part (J.P.B.)

Prior to their tectonic incorporation into the Alpine orogenic system during the Eocene, the Helvetic units were part of the stable European continent and formed its southern margin. Towards the south, the Helvetic units pass into the Ultrahelvetic Unit and the Penninic Flysch Zone (see below) (Figs. 17.11 & 17.12). Due to the Alpine Orogeny causing overthrusting and increased subsidence of the Helvetic units, deeper marine *Globigerina*-rich marls and shales were deposited during the Late Eocene. This is, herein, considered to be a transition to the foreland sedimentation of the North Alpine Foreland Basin (see below).

Sedimentary and stratigraphic development

The sedimentary history of the Helvetic units can be summarized in a number of stages (Fig. 17.9): (1) continental residual fissure-filling (= siderolithic) and freshwater limestones of Middle Eocene age; and (2) shallow-marine sediments (i.e. nummulitic limestones) indicating the beginning of subsidence during the Early Eocene (perhaps Palaeocene) in eastern Switzerland and during the Middle Eocene in the west; subsidence was accompanied by prominent normal faulting creating different structural blocks. The timing of this faulting, associated with eustatic sea-level changes, was responsible for the sedimentation in the Helvetic domain, as demonstrated by Menkfeld-Gfeller (1995).

From the Lutetian to the Priabonian, these shallow-marine units were covered by micaceous *Globigerina* marls and shales (Stad Schiefer) with an anoxic facies occurring in the NW (*Meletta* Shales). Lutetian to Priabonian deep-marine sediments are present in the Helvetic nappes (Einsiedeln Formation, Steinbach-Gallensis Formation, Bürgen Formation, Globigerinenmergel, Flysch sudhelvétique, Marnes à Foraminifères, Hohgant Formation). In addition, coastal facies (Klimsenhorn Formation, Wildstrubel Formation) and brackish environments (Couches à Cerithes, Couches des Diablerets, Sanetsch Formation) of the same age are also recognized (see Herb *et al.* 1984; Herb 1988; Menkfeld-Gfeller 1994, 1995, 1997; Weidmann *et al.* 1991).

The western part of the Palaeogene Ultrahelvetic domain is characterized by deep marine sedimentation. This is described and discussed below.

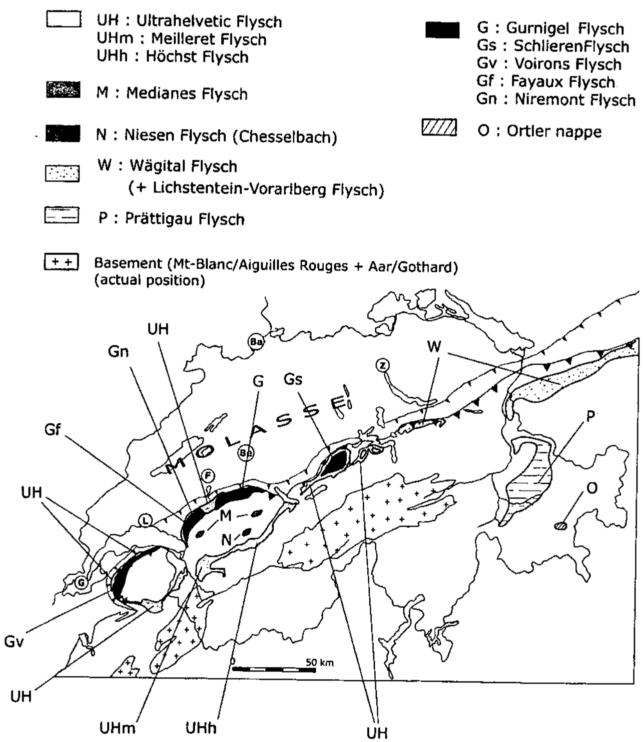


Fig. 17.11. Flysch deposits in Switzerland.

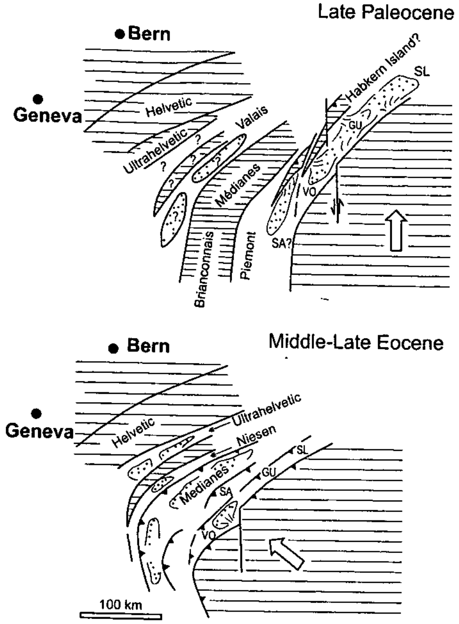


Fig. 17.12. Palaeogeography of the Swiss Flysch after Caron *et al.* (1989). Abbreviations: GL, gurnigel; SA, Sarine Slices; SL, Schlieren; VO, Voirons.

Helvetic units: eastern part (M.W.R.)

The Helvetic units in Austria and Bavaria are part of the Eastern Alpine Foreland, and form a narrow, east–west trending belt north of the Alpine orogenic wedge. Sedimentary successions in the Vorarlberg area (westernmost Austria) are comparable to those of Switzerland (see above). Therefore, this section focuses on the occurrences in Bavaria, Salzburg and Upper Austria.

Overviews have previously been published by Richter (1978), Prey (1980), Tollmann (1985), Oberhauser (1991, 1995), Kurz *et al.* (2001) and Rasser & Piller (2001). Most occurrences of the Helvetic Zone in western Austria are tectonically isolated; all of them are considered to be part of the southern Helvetic facies. Classic studies based on the Bavarian occurrences, such as the Kressenberg (Gümbel 1861; Schlosser 1925; Hagn 1967, 1981; Hagn & Wellnhofer 1973; Ziegler 1975, 1983; Rasser & Piller 2001), have been published. However, only rare publications deal with other Bavarian localities, such as Grünten (Reis 1926; Ziegler 1983) and Neubeuern am Inn (Hagn & Darga 1989). Occurrences in Salzburg and Oberösterreich (Austria) are dominated by sediments attributed mostly to the Southern Helvetic facies. They are best developed in the Haunsberg area, north of Salzburg (Traub 1953, 1990; Gohrbandt 1963; Vogeltanz 1970; Prey 1980, 1984; Kuhn & Weidich 1987; Kuhn 1992; Schultz 1998; Rasser & Piller 1999*a*, *b*, 2001).

Palaeogeography

From north to south, the central and eastern part of the Helvetic Unit (HU) passes into the Ultrahelvetic Unit (UU), the Rhenodanubian (= Penninic) Flysch Unit (RFU), and the Austro-Alpine units including the Gosau Basin.

Most authors suggest that the Palaeogene depositional environment of the HU as well as the adjacent UU and RFU was morphologically determined by the presence of several submarine highs and island chains. The northernmost high is the Intrahelvetic High (Hagn 1954, 1981; Vogeltanz 1970), which separates a northern (Adelholzen facies) and a southern (Kressenberg facies) Helvetic facies unit. Emersion of this high from latest Maastrichtian to Middle Eocene times, resulted in the development of a sedimentary gap in the northern HU, while sedimentation continued in the southern HU (Hagn 1981). The Pre-Vindelician island chain (Traub 1953) marks the transition from the HU to the UU. The UU can be subdivided into a northern Ultrahelvetic facies, characterized by marls, and a southern Ultrahelvetic facies, characterized by carbonate-free clays (Hagn 1981). The southern Ultrahelvetic facies was separated from the adjacent turbiditic RFU by the Cetic Island Ridge (Hagn 1960, 1981).

The UU represents a depositional realm to the south of the HU. It is characterized by the Cretaceous to Eocene 'Buntmergelserie', comprising marls which contain olistoliths and turbidites probably derived from the Cetic Island Ridge (Faupl 1977).

Tectonic setting

The Palaeogene development of the Helvetic shelf was influenced by the Alpine Orogeny. Collisional processes led to the subduction of the European Plate below the African-Adriatic Plate within the Penninic Realm. Subduction resulted in the formation of the Alpine Foreland Basin with a more or less east–west striking basin axis and a corresponding east–west striking shoreline (Prey 1980; Wagner 1998). Sedimentation in the HU ceased in the Late Eocene, when it was completely overthrusted by the northward-moving Alpine nappe system (Hagn 1981).

Sedimentary and stratigraphic development

Sediments in Austria and Bavaria show a more or less continuous development from the Jurassic to the Late Eocene (Fig. 17.9). A generalized section of the Palaeogene Helvetic Unit comprises Danian to Upper Thanetian siliciclastic pelagic sediments, followed by uppermost Thanetian algal limestones. These are overlain by Ypresian to Lutetian ferruginous larger foraminiferal limestones and siliciclastic sediments. Deep-marine conditions occurred from the Bartonian to the Priabonian; they are characterized by the marly 'Stockletten". Sedimentation ceased at the Eocene–Oligocene boundary.

The Palaeogene succession begins with Palaeocene marls and marly sands of the Olching Formation. They developed directly above Upper Cretaceous sediments (Traub 1953, 1990). The Danian–Thanetian Olching Formation (P1a–P5; Kuhn & Weidich 1987; Kuhn 1992) represents one of the most fossiliferous Palaeocene units with respect to macrofossils (e.g. Schultz 1998). Its depositional depth was *c.* 50–150 m (Kuhn 1992). The abundance of sands and sand beds intercalated with the marls increases upward. The Olching Formation is overlain by the Thanetian Kroisbach Member, which represents the basal part of the Kressenberg Formation. This member is characterized by quartz sandstones rich in iron concretions, iron ooids, small brachiopods and pycnodont bivalves, as well as coarse-grained sands particularly rich in pycnodont bivalves (formerly known as

'Gryphaeenbank'). Glauconite is generally abundant and increases up-section. The coarsening-upward trend and the termination by ooid-bearing sandstones suggest a shallowing of the depositional system from the Danian to Thanetian. This relative sea-level fall corresponds to a long-term eustatic sea-level fall, which indicates that subsidence during the Palaeocene was low.

The glauconitic sands of the Kroisbach Member pass upwards into red-algae-dominated limestones (compare Kuhn (1992) and modifications by Rasser & Piller (2001)) of the Thanetian Fackelgraben Member. This member is characterized by bedded rhodolith and red algal detritic limestones with thin marly intercalations. It develops gradually from the pycnodont sands below and still contains detritic glauconite.

Near the Palaeocene–Eocene boundary a deepening of the environment is indicated in western Austria by the shift from the red-algae-dominated limestones towards a globigerinid facies (NP9, after Oberhauser 1991). Due to an angular unconformity, these deeper-water sediments are absent in Salzburg. This would suggest that tilting and erosion took place close to the Palaeocene–Eocene boundary. Rasser & Piller (2001) argued that the angular unconformity was related to a tectonic pulse within the Alpine system to the south ('Laramide 3' of Tollmann 1964) which corresponds to a rapid westward shift of the Alpine Nappes (Oberhauser 1995). Consequently, the Alpine Orogeny began to affect the sedimentation in the HU close to the Palaeocene–Eocene boundary.

The Thanetian to Lutetian Kressenberg Formation is characteristic for the earlier Eocene of the HU. The Ypresian commences with the mainly ferrugenic Frauengrube Member (= 'Roterzschichten'). This is characterized by massive, iron-ooid bearing, nummulite limestones and coarse-grained sandstones to fine-grained conglomerates. Quartz grains are frequently coated by iron, while most bioclasts are completely ferruginized. Flat and unfragmented larger foraminifera indicate very low energy conditions. However, the occurrence of high-energy iron ooids and ferruginized allochems in the same sediments suggest the opposite. Consequently, this facies has been interpreted by Rasser & Piller (2001) as representing a mixture of two different environments (e.g. input of iron ooids from ooid bars into the relatively deeper water where the larger foraminifera lived). The Ypresian-age ferrugenic sedimentation ceased with the formation of sand bodies of the Sankt Pankraz Member. In the Haunsberg area, it is characterized by up to 32 m of matrix-free, fine-grained sandstones without fossils. Its development in Bavaria is characterized by a coarse-grained sandstone to fine-grained conglomerate, rich in larger foraminifera. In the Haunsberg area, these sands represent subtidal sandbodies which show a marked fluvial influence, as well as a high degree of reworking (Vogeltanz 1970). The facies succession indicates a shallowing-upward from ooid bars (Frauengrube Member), surrounded by deeper foraminifera-rich sediments, up to the sandbodies of the Sankt Pankraz Member. The Middle Ypresian age of these ferruginous foraminiferal limestones (P7?, after Kuhn 1992) coincides with a long-term eustatic sea-level fall beginning during the Middle Ypresian. The rates of siliciclastic input can be correlated with a short-term eustatic sea-level fall during the latest Ypresian. The accommodation space necessary for the deposition of the ferruginized limestones and the 32-m-thick sandstones requires an increased subsidence rate at that time, suggesting that the subsidence rate was much higher than during the Palaeocene and the Early Ypresian.

Lutetian limestones are represented by the Kressenberg Member and the Weitwies Member (Rasser & Piller 2001). The Kressenberg Member (= 'Schwarzerz Schichten') is characterized by coarse-grained, poorly sorted, limestones containing iron ooids and larger foraminifera. In Bavaria, this member had been mined for its iron (Gümbel 1861; Hagn & Wellnhofer 1973). The Weitwies Member (= 'Fossilschicht') comprises glauconitic bioclastic limestones rich in larger foraminifera. Towards the top these grade into glauconitic marls with nummulitids and molluscs. In the Haunsberg area it also contains phosphate nodules (Traub 1953).

The entire Lutetian succession represents a period of eustatic sea-level fall. Since the described facies pattern indicates a relative sea-level rise, Rasser & Piller (2001) have suggested that the increased subsidence rate continued during the Lutetian. This coincided with the flooding of the northern Helvetic Unit, which represents a backstepping typical for transgressions on carbonate ramps (cf. Dorobek 1995). This feature is typical for foreland basins and was related to the ongoing subduction of the European Plate below the African-Adriatic Plate.

Lutetian shallow-water carbonate deposition was terminated by Bartonian to Priabonian deep-water conditions, and the so-called 'Stockletten' (globigerinid marls, >100 m thick) overlay the Kressenberg Formation. This succession represents an ongoing backstepping of facies belts of the ramp towards the north. The autochthonous subsurface of the later North Alpine Foreland Basin was covered by Bartonian–Priabonian shallow-water carbonates, which were reworked as olistoliths into the 'Stockletten' (Buchholz 1989; Darga 1992; Rasser 2000). Further Eocene–Oligocene shallow-water carbonates of the circum-Alpine area, include occurrences of the HU, were summarized by Nebelsick *et al.* (2003, 2005). The long-term eustatic sea-level fall continued until the Early Priabonian and was followed by a continuous rise up to the Eocene–Oligocene boundary.

Alps and Alpine Foredeep: overview (M.W.R.)

The Alps are a mountain chain formed during the multiphase Alpine Orogeny (see also Froitzheim *et al.* 2008; Reicherter *et al.* 2008). They are separated into the Western Alps situated west of the Rhine Valley, and the Eastern Alps between the Rhine Valley and the Vienna Basin (Fig. 17.13). The Carpathians represent their eastward continuation. During the Palaeocene and Eocene, the Alpine system formed an archipelago including the Gosau basins. The Penninic tectonic units together with the Rhenodanubian Flysch Unit, formed the northern continuation of the Gosau basins and represent the main element of the Alpine Foredeep. The Helvetic Unit (see above) formed the northern continuation of the Penninic units. Towards the east, the Flysch Zone continues into the Carpathian Foredeep. During the Eocene, the Alpine Orogeny led to subduction of the Penninic oceanic crust and related units beneath the Alps. Together with the Helvetic Unit these were incorporated into the Alpine orogenic wedge, resulting in the termination of sedimentation in the Penninic and Helvetic units, and the Gosau basins during the Eocene. Sedimentary environments and processes changed remarkably during Oligocene and Miocene times.

Penninic units: flysch zones in Switzerland (J.P.B.)

Caron *et al.* (1989) presented a detailed overview of the palaeogeographic development from the Turonian to the Late Eocene subdividing the sucession into three areas (Figs. 17.11 & 17.12): (1) Ultrahelvetic Unit and Niesen- and Medianes Flysch with a north-Tethyan character; (2) the Gurnigel Flysch of Central Tethyan origin; and (3) the south Tethyan Flysch, restricted to the Cretaceous. During the Late Eocene, the

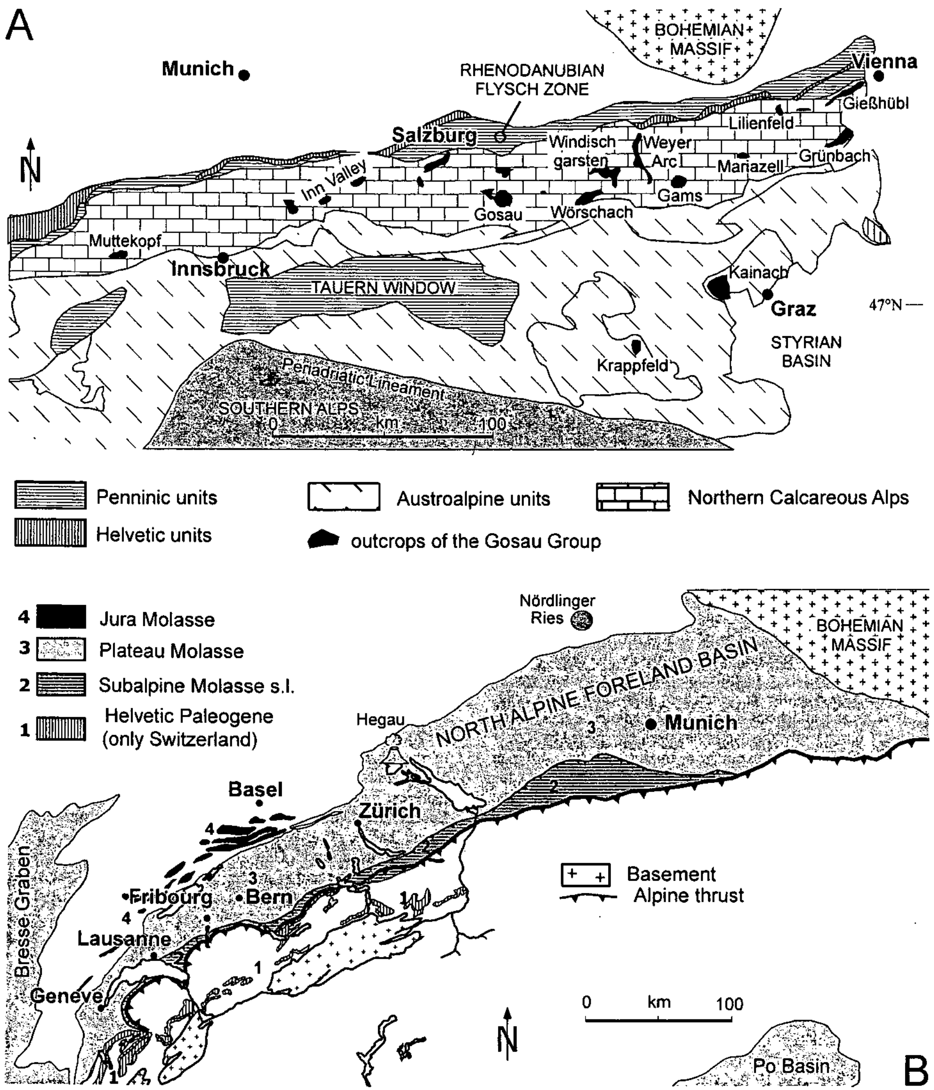


Fig. 17.13. Geology of (A) the Eastern Alps and adjacent areas and (B) the western part of the North Alpine Foreland Basin and the Swiss part of the Western Alps.

thrusting of the Penninic nappes onto the European Foreland marked the end of the deep-marine clastic sedimentation and the creation of the North Alpine Foreland Basin. Increased subsidence rates are reflected by the presence of globigerinid marls in the Helvetic Unit.

The term 'flysch' was defined in Switzerland as a rock stratigraphic unit in the Alpine front ranges (Studer 1827). From a geodynamic point of view, the formation of 'flysch' sediments requires subaerial exposure and high erosion rates in the source areas, short transport distances, and sufficient relief and slope for

gravity flow processes to occur. Petrographical and structural analyses clearly demonstrate the role of compressive tectonics on the 'flysch' sedimentation (Caron *et al.* 1989). However, as postulated by Stuijvenberg (1979), eustasy was also responsible for the accumulation of the Gurnigel Flysch.

Sedimentary and stratigraphic development
The various structural units which make up the Pre-Alps (from the Arve River in France to Lake Thun) may be situated palaeogeographically in a series of basins extending from the European margin of the Tethys across an oceanic area and onto the southern margin (Adria); from NW to SE these are (Figs. 17.9, 17.11 & 17.13): (1) the distal Helvetic shelf with the Ultrahelvetic Unit; (2) the Valais Trough as an oceanic basin including the Niesen Nappe and North Penninic Mélange; (3) the Briançonnais Microcontinent containing the Pre-Alpes Medianes and Breche Nappes; (4) the Piedmont-Ligurian Ocean including the Gurnigel Flysch and Dranses Nappes; and (5) the South Alpine margin with the Simme Nappe. The orogenic facies of the Pre-Alps (i.e. flysch) were first deposited during the Cretaceous, extending to Middle–Late Eocene times. Olistostromes and mélange formations, which are related to tectonic phases, constitute chaotic units which have been termed 'wildflysch'. Some of these important nappe systems are presented below.

The **Ultrahelvetic nappes** (Lutetian–Bartonian) occur in tectonically complex units, together with Triassic to Late Eocene formations. These units occur on and between the Helvetic nappes, between the Niesen and Wildhorn nappes, and between the Gurnigel Nappe and the sub-Alpine Molasse. The Meilleret Flysch (Homewood 1974) is composed of pelitic and arenitic turbidites, which are cut by coarse clastics. The successions terminate with microbreccias and conglomerates containing abundant shallow-marine fauna (algae, foraminifera, bryozoans and corals). The Höchst Flysch (Ferazzini 1981) contains thick-bedded coarse arenites and conglomerates. According to benthic and planktonic foraminifera as well as nannofossils, the age of these deposits is Middle to Late Eocene. The conglomerates are composed of clasts derived from the Variscan basement and the Triassic–Cretaceous cover. The heavy mineral association comprises zircon, tourmaline, barite and TiO_2. The depositional environment was a moderately deep-water environnement, with a tectonically active source in the south, and a short transport distance.

The **Niesen Nappe** is essentially of Maastrichtian age, with a Lutetian upper part (Chesselbach Flysch; Ackermann 1986). This unit is thrust directly over the Ultrahelvetic Unit and is overlain by a mélange (zone submediane; Weidmann *et al.* 1976). The Chesselbach Flysch is composed of arenitic turbidites and shales, with north–south palaeocurrents. Its maximum thickness reaches 200 m. The top of this unit is dated as Middle Eocene by benthic foraminifera. Deposited in a deep-marine environment, this flysch contains a heavy mineral association composed of tourmaline, zircon, TiO_2 and apatite.

The **Medianes nappes** are of ?Early to Middle or Late Eocene age (NP15 is recorded), probably even younger (Caron *et al* 1980). This flysch is in stratigraphic continuity with the hemipelagic marls of the Couches Rouges Group (Palaeocene–Middle Eocene; Guillaume 1986). It is highly deformed and commonly associated with (and covered by) chaotic wildflysch deposits (flysch à lentilles de Couches rouges). The heavy minerals are dominated by garnet, zircon, tourmaline and TiO_2.

The **Gurnigel Nappe**, known from Bonneville (France) to central Switzerland (Caron 1976; Stuijvenberg 1979), is Maastrichtian to Eocene (Bartonian) in age. Its basal thrust is thought to override the Ultrahelvetic units and Pre-Alpes Medianes, but later reverse faulting has emplaced the front of the Pre-Alpes Medianes over the rear of the Gurnigel Nappe (Caron *et al.* 1980). Several local names have been proposed within this unit, such as the Voirons Flysch (Stuijvenberg 1980), the Fayaux Flysch (Stuijvenberg *et al.* 1976), the Niremont Flysch (Morel 1980) and the Schlieren Flysch (Winkler 1983).

This flysch is mainly composed of turbiditic shales and thin sandstones with fine-grained carbonate turbidites, hemipelagic shales with bentonite layers as well as rare thick-bedded sandstone channels and conglomerates. The thickness ranges from 1.5 km (Gurnigel-Schlieren area) to 2.5 km (Voirons). The original substratum is unknown. The upper limit corresponds to an erosional surface. This flysch comprises classical Bouma-type turbidites and bears the entire facies spectrum of Mutti & Rici Lucchi (1972). The depositional environment was an abyssal, deep-sea fan. The dominant palaeocurrent trend is from west to east (Wildi 1985). The conglomerate clasts are composed of 20% metamorphic, 50% magmatic and 30% sedimentary rocks with two dominant populations: (1) tonalite, quartzite, and dolomicrites (Triassic) with various other Mesozoic and basement rocks; (2) granite, granodiorites and cherts, together with other sedimentary, metamorphic and crystalline rocks. The basement is of South Alpine origin, the Mesozoic clasts were derived from platforms to deep basins situated in the South- and Austro-Alpine units. The heavy mineral association comprises garnet, apatites and staurolite (Wildi 1985). Several bentonite layers (Winkler *et al.* 1985) have their acme in the Late Maastrichtian and the Early Eocene.

The **Wägital Flysch** is recognized in central and eastern Switzerland. It is very similar to the Gurnigel Flysch, and probably represented the geographic transition from the Gurnigel-Schlieren area to the Rhenodanubian Flysch Unit (Trümpy 1980).

From flysch to wildflysch and to foreland basin sedimentation
Until the 1980s, many geologists equated the terms 'flysch' and 'turbidites'. Therefore, 'flysch' was used for each turbiditic unit found in the peri-Alpine realm. Thus, several sedimentary units studied in the Helvetic nappes and the sub-Alpine area have been named according to this philosophy, as for example the 'North Helvetic Flysch' (with the well known locality of Engi, rich in fish remains) or the 'Subalpine Flysch'. Incorporating the geodynamic control between the earlier and later stages of the Alpine Orogeny, Homewood & Lateltin (1988) published an exhaustive review of these units and placed them into a molasse (foreland-basin style sedimentation) geodynamic context, with the erection of the Taveyannaz and Val d'Illiez formations replacing the North Helvetic Flysch and the Subalpine Flysch, respectively. The Val d'Illiez Formation passes upward to the littoral facies of the Lower Marine Molasse. Thus, the original conception of the Lower Marine Molasse extended to the Val d'Illiez and Taveyannaz formations.

Rhenodanubian Flysch Unit (M.W.R.)

The Rhenodanubian Flysch Unit (RFU) is part of the Alpine-Carpathian Foreland and received its name from the equivalent occurrences in the Swiss Western Alps. The RFU forms a narrow zone along the northern front of the Eastern Alps (Fig. 17.13). With a length of 520 km it strikes in a west–east direction from the Rhine Valley towards the east and dips below the Neogene of the Vienna Basin and continues into the Carpathians (e.g, Magura Nappe).

Palaeogeography

During the Palaeogene, the Northern Calcareous Alps (NCA) were part of the Austro-Alpine Microplate (for summary Faupl & Wagreich 2000; Wagreich 2001; see also Froitzheim *et al.* 2008) situated between the European Plate to the north and the Adriatic Plate to the south (Channel *et al.* 1992). The RFU was part of the Alpine Foredeep to the north of the NCA. Slope deposits of the NCA Gosau Group formed the southern margin of the RFU (Wagreich 2001; Trautwein *et al.* 2001). To the north, it was bordered by the Ultrahelvetic Unit. The RFU was part of the Northern Alpine Foredeep and was formed from eroded material of the rising Alpine mountain chain. In this respect, it was functionally replaced by the North Alpine Foreland Basin during the Oligocene and Neogene. The palaeogeographic origin of the RFU as part of the Valais palaeogeographic domain is a matter of discussion (e.g. Kurz *et al.* 2001, and references therein). Wagreich (2001) assumed that the depositional area of the RFU was situated within the (northern) Penninic domain (compare Faupl & Wagreich 2000).

Tectonic setting

The Alpine Orogeny and the development of the Gosau basins were controlled by ongoing convergence between Africa and Europe in late Mesozoic and Cenozoic times. The northward drift of the African-Adriatic Plate led to subduction of the European Plate within the Penninic realm and subsequent uplift of the Austro-Alpine nappe stack. Northward tilting and basin subsidence of the NCA occurred during the latest Cretaceous to Palaeocene. This was caused by tectonic erosion of parts of an accretionary wedge to the north of the NCA (Wagreich 1993, 1995).

Sedimentary and stratigraphic development

Sedimentation of the RFU (Fig. 17.9) commenced during the Early Cretaceous with carbonate-dominated deep-marine deposits passing into siliciclastic sediments. Upper Cretaceous turbidites are intercalated with pelitic intervals related to sea-level changes. Palaeogene deposits are characterized by stable heavy minerals (e.g. Wagreich & Marschalko 1995; Trautwein *et al.* 2001). Sedimentation ceased during the Eocene, when the RFU was subducted below the Alpine nappe stack. Cretaceous deep-marine sediments are known from western Austria (Egger 1990, 1995). Palaeogene deep-marine sediments are best developed in SE Bavaria (Freimoser 1972), Salzburg and Lower Austria (Egger 1995). Overviews on the tectonics, stratigraphy and palaeogeography have been published by Egger (1992, 1995), Oberhauser (1995), Wagreich & Marschalko (1995), Adamova & Schnabl (1999) and Faupl & Wagreich (2000). According to these papers, the stratigraphic and sedimentary development of the area can be subdivided into various stages (see below).

The Maastrichtian (CC25) to Palaeocene (NP9) Altlengbach Formation is known from both the central (Salzburg) and the western (Wienerwald, Lower Austria) parts of the RFU (Egger 1995). It extends from the Campanian–Maastrichtian boundary through to the uppermost Palaeocene. The Palaeocene parts of the Altlengbach Formation are restricted to the Acharting Member, the deposition of which began during the Maastrichtian. The Acharting Member is characterized by rhythmic alternations of fine-grained sandstones to siltstones and turbiditic mudstones. The sand- and siltstones have a high carbonate content and include foraminifers and coralline algal fragments. As revealed from heavy minerals (Egger 1990), palaeocurrents from the east prevailed up to the early Palaeocene, changing to a westerly direction during the later Palaeocene.

In Salzburg, the Palaeocene–Eocene boundary interval is present within the Anthering Formation (NP9/P6 to NP11; Egger 1995). Egger *et al.* (1997) subdivided the Anthering Formation into mud turbidites, hemipelagites and bentonites. The mud turbidites are dominated by silty marls and marls, while sandstones and siltstones typical of the Altlengbach Formation are subordinate. Turbidite beds are up to 2 m thick. Hemipelagic claystones are usually bioturbated, contain abundant agglutinated foraminifera, and have relatively high organic carbon content. In contrast to the turbidites, the hemipelagic claystones generally lack planktonic organisms, benthic foraminifera and bioturbation. These are, therefore, interpreted as having formed in an oxygen-deficient environment. Several bentonite layers were found in the type area of the Anthering Formation and were interpreted as volcanic ashes. Egger *et al.* (1997) suggested a short period of marked explosive volcanic activity. The reduced abundance of siliciclastics in the Anthering Formation, when compared to the Altlengbach Formation, is interpreted to be related to a eustatic sea-level rise rather than to tectonic subsidence (Egger *et al.* 1997). The high abundance of bentonites in the pelitic rocks suggests that increased volcanic activity accompanied this transgression.

During the Early Eocene, sedimentation ceased in the Salzburg part of the RFU. In the east Austrian Wienerwald (Vienna Woods) area, the eastern equivalent of the Anthering Formation is represented by the Greifenstein Formation, whose deposition commenced during the Late Palaeocene (NP8; Egger 1995). While the earlier Palaeocene development is comparable between Salzburg and the Wienerwald area (Acharting Member), different sediment types were deposited from the latest Palaeocene onwards. The Greifenstein Beds (e.g. Hösch 1985, cited in Egger 1990) are characterized by thick-bedded, coarse-grained sandstones. In part they are rich in glauconite and nummulites (e.g. Oberhauser 1980). Another Late Palaeocene–Eocene equivalent, but located on a different nappe, are the Laab Beds. Based on fision track data, Trautwein *et al.* (2001) suggested the existence of two basins in the eastern part of the RFU: a main basin close to the NCA nappe stack, and a Laab Basin towards the north existing since Early Cretaceous time.

North Alpine Foreland Basin: western part (J.P.B.)

The North Alpine Foreland Basin (NAFB) – often referred to as the Molasse Basin – was part of the Alpine-Carpathian Foredeep forming a west–east trending basin in front of the prograding nappes of the Alpine orogenic wedge. It can be traced from the Rhône Basin in the west via Switzerland and Bavaria to Austria (Fig. 17.13).

Palaeogeography

Several palaeogeographic reconstructions of the NAFB have been published (Berger 1996; Sissingh 1998; Schlunegger & Pfiffner 2001; Kuhlemann & Kempf 2002; Becker & Berger 2004; Berger *et al.* 2005*b*). During the Lutetian, the sea was located in the area of north Italy and southern Switzerland, with its northern shoreline about 70 km south of Bern (Sissingh 1998). The Alpine front was probably situated *c.* 300 km south of its present position (Dèzes *et al.* 2004). Most of the area of present-day Switzerland was affected by erosion (marked by local siderolithic deposits).

Subsequently, the sea transgraded along the northern front of the Alps. During the Late Rupelian, the Paratethys regressed towards the east. It is most likely that during this time a marine connection existed between the NAFB and the Upper Rhine Graben (URG) (perhaps during NP23; see Berger 1996; Kuhlemann & Kempf 2002; Diem 1986; Berger 1995; Picot 2002). A marine connection between the URG and the NAFB was only possible via the central and eastern parts of the Swiss Foreland Basin, since its western part was covered by fluviatile sediments (Berger 1996) while in its eastern part (Bavaria), open marine conditions occurred. Following a regression during the Early Chattian, fluvial sedimentation prevailed in the Swiss part of the NAFB. Alluvial conglomeratic fans, derived from the Alps, were drained by a SW–NE orientated fluviatile system referred to as the 'Genferseeschüttung'. The fluvial drainage joined the marine development still present in Bavaria. Part of this system continued into the southern URG through the Jura Molasse, as evidenced by the heavy minerals of the 'Molasse Alsacienne' (Picot 2002).

The Late Chattian shows an important decrease in clastic supply to the western Swiss NAFB, resulting in the development of lacustrine and brackish conditions, as attested by the accumulation of lacustrine limestones and gypsiferous marls (Molasse à charbon, Grès et marnes gris à gypse). During the Aquitanian, the accumulation of alluvial clastic sediments continued in the Swiss Molasse Basin, with local lacustrine and brackish deposits dominating in its distal parts. The occurrence of a marine ingression in its western parts is evidenced by the presence of mammal faunas (MN2a) which were trapped in tidal deposits (Berger 1985). In Bavaria, the fluvial drainage entered the sea in the vicinity of Munich (Kuhleman & Kempf 2002). A marine ingression, originating from west and east, covered the entire NAFB in the Burdigalian. During the Late Burdigalian, the sea probably regressed towards the distal part of the basin, where brackish conditions prevailed (Marnes rouges, Helicidenmergel; see Becker 2003), and then towards the west (i.e. France). An important estuary is evident in NE Switzerland and SW Germany, the so-called 'Graupensandrinne' (see Reichenbacher *et al.* 1998). In the German part, brackish and lacustrine sedimentation prevailed.

During the Langhian a NE–SW fluviatile drainage system (= 'Glimmersand') drained the NAFB from Germany to Switzerland. Alluvial fans derived from the Alps (Napf, Hörnli and Bodensee-Pfander, Hochgrat, Nesselburg) are evident in central and eastern Switzerland and Bavaria (see Kuhlemann & Kempf 2002). All of these deposits form the so-called Upper Freshwater Molasse (Obere Süsswassermolasse) and contain several bentonite levels. Equivalent sediments are unknown for the western part of the Swiss NAFB, where 2000 m of Miocene sediments were probably eroded during the Plio-Pleistocene (Kuhlemann & Kempf 2002). Only relicts of marine middle Miocene deposits occur in the western Jura Molasse (NN4–NN5; see Kälin *et al.* 2001). During the Serravallian, fluviatile sedimentation continued, with an important clastic source located in the Vosges and Black Forest massifs, confirming continued uplift during this time.

The Tortonian drainage pattern is very difficult to reconstruct in Switzerland; a general west–east drainage of the Swiss NAFB has been proposed (Giamboni *et al.* 2004; Kuhlemann and Kempf 2002; Liniger 1966; Petit *et al.* 1996; Schlunegger *et al.* 1998). This pattern is based on the uplift of the western and central parts of the NAFB in conjunction with the folding of the Jura Mountains. In Germany, fluviatile sedimentation persisted until the Early Tortonian (11 Ma; Kuhleman & Kempf 2002).

Sedimentary and stratigraphic development

The NAFB is traditionally subdivided into four depositional groups (Fig. 17.9): (1) the Lower Marine Molasse (= 'Untere Meeresmolasse', UMM) commenced with the increase of subsidence leading to the creation of the foreland basin; (2) the Lower Freshwater Molasse (= 'Untere Süsswassermolasse', USM) is composed of freshwater and rare brackish sediments; (3) the Upper Marine Molasse ('Obere Meeresmolasse', OMM) represents general marine sedimentation occurring in the entire basin; and (4) the Upper Freshwater Molasse (= 'Obere Süsswassermolasse', OSM) represents freshwater sedimentation extending to the Tortonian. The NAFB is also traditionally subdivided into a Sub-Alpine (thrusted and folded units, originally deposited further to the south) and a 'Plateau' or 'Foreland' molasse, which is relatively autochthonous. The distal part of the Swiss NAFB is generally called 'Jura Molasse', representing the connection with the Upper Rhine Graben.

In the **Sub-Alpine Molasse**: the UMM deposits comprise the Meletta shales, the Taveyannaz and Aldorf sandstones, the Val d'Illiez Formation and the Vaulruz Formation (see Diem 1986; Kuhlemann & Kempf 2002; Lateltin 1988; Schlunegger *et al.* 1997). In the western part of the Swiss NAFB, marine sedimentation stopped earlier (not before NP22) than in the eastern part, where marine sediments are known until NP24 or even NP25 (Doppler *et al.* 2000). Marine sedimentation continued in Germany during the Rupelian (i.e. Bausteinschichten) and even until the Aquitanian in Bavaria.

Due to the general sea-level fall (corresponding with the Ch1 sequence; see Hardenbol *et al.* 1998) these deposits are overlain by the freshwater deposits of the USM, which consist of conglomeratic alluvial fan deposits, fluviatile sediments, and palustrine-lacustrine deposits, such as the Molasse à Charbon (see Berger 1998; Engesser *et al.* 1984; Fasel 1986; Schlunegger *et al.* 1996, 1997). In Germany, the Chattian sediments were essentially deposited in freshwater environments (Granitische Molasse), passing towards the east into brackish (Cyrenenschichten) and later fully marine conditions (Prombergerschichten; Schwerd *et al.* 1996). No Miocene sediments have been recorded in the Sub-Alpine Molasse of Western Switzerland. Some conglomeratic fans persist in eastern Switzerland until the Middle Aquitanian (Schlunegger *et al.* 1997). In Germany, the Aquitanian is characterized from west to east by the freshwater–marine transition (i.e. 'Aquitan Fischschiefer'; Schwerd *et al* 1996). The OMM and the OSM are only rarely present.

In the **Swiss Plateau Molasse**, continental deposits, rich in lateritic products (so-called Siderolithic, with 'Bohnerz' = iron ore nuggets and/or 'Hupper' = quartz sands), were deposited, some of them dated by mammals (Engesser & Mödden 1997; Hooker & Weidmann 2000). Few lacustrine deposits of Rupelian age are known from the Swiss Plateau Molasse (Calcaires inférieurs), which are generally dated by charophytes (Berger 1992). In the Early Chattian, fluviatile sediments (Untere Bunte Mergel) were deposited in the entire foreland basin. During the Middle and Late Chattian, lacustrine and brackish limestones, dolomites and gypsiferous marls occur. In the NE distal part of the basin, sedimentation probably did not begin before the middle Chattian (Müller *et al.* 2002; Schlunegger & Pfiffner 2001). All mentioned units are relatively well dated by mammals, charophytes, otolithes and magnetostratigraphy; see Berger *et al.* 2005*a*).

In Germany, the UMM is represented by the distal parts of the Tonmergelschichten and Bausteinschichten (Rupelian). During the Chattian, a freshwater to marine transition occurred (Doppler *et al* 2000; Schwerd *et al* 1996). A fluviatile facies characterizes the Aquitanian deposits of the Swiss Plateau Molasse (e.g. Molasse Grise de Lausanne, Obere Bunte Mergel, Granitische Molasse; Berger 1985; Keller *et al.* 1990). A marine transgression flooded the North Alpine

Foreland Basin during the late Aquitanian (MN2) and Burdigalian. Its typical deposits comprise the OMM (Berger 1985; Homewood & Allen 1981; Keller 1989; Kempf *et al.* 1997, 1999; Schoepfer 1989; Strunck & Matter 2002). In Germany, a freshwater–marine transition prevailed during the Aquitanian. Following reduced sedimentation during the early Burdigalian, the sea invaded the entire basin. In Switzerland, marine sediments were confined during the Langhian to the north and NE (Graf 1991); fluviatile sediments associated with alluvial fans are present in the south (OSM) and persisted until the Serravallian (MN7, ?MN8) (Bolliger 1992; Kälin 1993; Kälin & Kempf 2002; Kempf *et al.* 1999). The OSM is very well represented in Germany, with conglomerates alternating with fluvial and lacustrine sediments. The meteorite of the Nördlinger Ries impacted during the Early Serravallian (14.6 Ma). In Switzerland, no Upper Serravallian, Tortonian or Messinian sediments are recorded. However, several authors suggested that an additional 700 m and up to >2000 m of sediment were deposited in the east and west Swiss NAFB (see Kuhleman & Kempf 2002).

The Palaeogene to Neogene **Swiss Jura Molasse** is preserved in a number of synclines in the Jura Mountains (Picot 2002; Becker 2003). Eocene deposits are represented by siderolithic units. Even though it has not been precisely dated, a Lutetian, Bartonian and Priabonian age is suggested based on correlation with other siderolithic occurrences. The Swiss Jura Molasse was subdivided by Berger *et al.* (2005*a*) into five different areas. (1) SW Jura Molasse (Valserine, Joux, Auberson, Travers, Val de Ruz): deposits of the Lower Marine Molasse are absent. The USM is represented by the Chattian Calcaires inférieurs and the Aquitanian Calcaires de La Chaux (MN2b). OMM sediments are preserved as well, but the OSM is absent. (2) NW Jura Molasse (Verrières, Pont de Martel, Locle-Chaux de Fonds): only deposits of the OMM and of the OSM are present; the latter is represented by the famous Oeningian facies. (3) Central-South Jura Molasse (St. Imier, Pery-Reuchenette, Tramelan-Tavannes, Balstahl, Moutier): thick units of fluviatile and lacustrine sediments accumulated during the Oligocene (Calcaires inférieurs, Molasse alsacienne, Calcaires delémontiens). After a gap during the Aquitanian, the OMM commenced with classic tidal sandstones. The OSM is essentially lacustrine (Oeningian facies) and conglomeratic. (4) Central-North Jura Molasse (Soulce, Delémont, Laufen, Porrentruy, Liesberg): the UMM is represented by marine marls dated from NP22 to NP25, alternating with Oligocene freshwater deposits (e.g. conglomerates, Molasse Alsacienne, Calcaires delémontiens). A sedimentary gap during the Aquitanian is followed by the sedimentation of the OMM and lacustrine OSM deposits. (5) East Jura Molasse (Mummliswil, Waldenburg): essentially USM and OMM are developed. OSM is only partly preserved (Glimmersand).

The Pliocene is principally represented by conglomerates (Ältere Deckenschotter, Graf 1993) in the NE of Switzerland. The only dated Pliocene deposits in this area are the karstic filling of the Vue des Alpes (MN15; Bolliger *et al.* 1993), and the Höhere Deckenschotter from Irchels (MN17, Bolliger *et al.* 1996). In Germany, Pliocene sediments have been reported from the 'Hochflächenschotter' and the 'Hochschotter' (e.g. Schwäbische Alb, Naabtal).

North Alpine Foreland Basin: eastern part (M.H., R.R.)

The Austrian part of the North Alpine Foreland Basin (NAFB) is bordered to the north by the passive margin of the Variscan Bohemian Massif and by the overriding Rhenodanubian Flysch Unit and Helvetic Unit of the Alpine-Carpathian thrust front in the south. The width of the *c.* 300 km long Austrian part of the NAFB in Lower and Upper Austria ranges from 5 to 50 km. The present basin is only a narrow remnant of the original basin (Fig. 17.13).

Tectonic setting

During the Oligocene, progradation of the Alpine-Carpathian nappe system resulted in the passive margin of the Bohemian Massif being transformed into a foreland molasse trough. In the south, emplacement of the Helvetic and Rhenodanunian Flysch units formed the southern slope of the basin. During the Late Oligocene, these units became integrated into the Alpine thrust front. The Calcareous Alps formed the shelf on which debris from the Central Alps accumulated. In the north a fault system of conjugate NW–SE and NE–SW trending faults in the southern part of the Bohemian Massif resulted in the formation of a triangular crystalline peninsula, which extended far south into the NAFB, separating the basin in the area into western and eastern parts. Tectonic activity within the thrust sheets and lateral movements of the basement along the eastern flank of the Bohemian Massif are still ongoing.

Sedimentary and stratigraphic development

The depositional history and stratigraphic development of the eastern NAFB differs considerably from that of its western part (see above) and may be subdivided into several stages. The oldest Cenozoic deposits in the NAFB are Upper Eocene fluvial and shallow-marine sandstones, shales and carbonates of the Perwang Group (Buchholz 1989; Rasser 2000) (Fig. 17.9). As a result of subduction of the European plate beneath the Peri-Adriatic plate and the weight of the advancing Alpine nappe system, the downwarping of the foreland crust accelerated during the Early Oligocene and the NAFB subsided rapidly into a deep pelagic area. Black shales were deposited in Switzerland, Bavaria and western Austria (= Fischschiefer). Towards the south, these shales graded into the deep-marine deposits of the Deutenhausen Formation. An upwelling current system might have been established during the Late Eocene, affecting deposition along the northern slope throughout the Early Oligocene, as suggested by the presence of nannoplankton ooze.

The Southern Bohemian Massif acted as the northern margin of the NAFB. During the Oligocene it was drained towards the south and east by fluvial systems, represented by the sands and gravels of the Freistadt-Kefermarkt Beds and the St. Marein-Freischling Formation. During the Early Oligocene (Kiscellian regional stage) vast mudflats and lagoonal embayments developed along the coast formed by the Bohemian Massif (e.g. mollusc-rich pelites, lignites of the Pielach Formation; Harzhauser & Mandic 2001). Sandy sediments of the coeval and overlying Linz-Melk Formation reflect the marine transgression during the Oligocene (Kiscellian and Egerian regional stages). These comprise lagoon, rocky shore, sandy shoreface and tide-influenced open-shelf environments. The interruption of this transgression by the marked Lower Egerian regression is evidenced by erosion surfaces, redeposition and intercalations of lagoonal dark coaly pelites (Roetzel 1983).

Along the western part of the northern margin of the NAFB, the Oligocene shallow-water deposits interfinger towards the south with offshore pelites of the Ebelsberg and Eferding formations. Due to the progressive Upper Oligocene transgression, these offshore pelites overlap the shallow-water deposits to the north. Along the southern slope of the NAFB, the Egerian Puchkirchen Formation was deposited in a deep marine channel, comprising slumps, conglomeratic debris flows and turbidites, derived from both slopes of the NAFB (DeRuig 2002). Frequently strong bottom currents reworked the turbidites into contourites (Wagner 1996, 1998).

Continuous deep-marine sedimentation in the eastern NAFB is contrasted with the deposition of the Lower Freshwater Molasse in Vorarlberg (western Austria) and Bavaria during the Egerian (see section above). Here, brackish environments formed in the Late Oligocene and overlie the shallow-marine deposits of the Baustein Beds. A westward prograding limnic-fluvial facies of the 'Lower Coloured Molasse' developed and this grades into the brackish swampy environments of the 'Lower *Cyrena* beds'. A short-lived marine ingression (Promberg Beds) is followed by the limnic-fluvial conditions of the 'Upper Coloured Molasse' in the latest Oligocene and early Early Miocene. In the Early Miocene (Eggenburgian regional age) marine sedimentation was re-established, as evidenced by the deposition of the shallow marine sediments of the Upper Marine Molasse. In eastern Austria, however, deep-water sedimentation continued during the Late Egerian and Eggenburgian, and is typified by the debris flows, slumps and contourites of the Eggenburgian Hall Group. During the Eggenburgian and Ottnangian (Early Miocene), deltaic gravels and sands were transported from the south into the NAFB, forming the conglomerates of the Buchberg and Wachtberg, today partly located in the imbricated nappes of the NAFB.

Shallow-marine deposits are mostly absent in the western, Upper Austrian part of the NAFB due to both submarine and later Alpine erosion. Erosional relicts of the Eggenburgian are best preserved along the eastern margin of the Bohemian Massif in Lower Austria. There, the Lower Eggenburgian successions reflect a stepwise landward shift of the palaeoshoreline (Mandic & Steininger 2003). The marine sediments onlapped either directly onto a relief formed by crystalline rocks of the SE Bohemian Massif, as shown by the Fels Formation, or prograded towards the NW into the estuarine-fluvial systems of the St. Marein-Freischling Formation. Initial brackish-estuarine biotopes of the Mold Formation were gradually replaced by fully marine conditions, typified by the sandy Loibersdorf Formation, which includes the historical holostratotype of the Eggenburgian stage at Loibersdorf (Steininger 1971). Continuing transgression onto the Bohemian Massif in the late Eggenburgian led to the deposition of the mollusc-rich nearshore sands of the Burgschleinitz and Gauderndorf formations. Following a prominent regressional phase, the sandy bryozoan limestones of the Zogelsdorf Formation were deposited at the beginning of the new Ottnangian transgressional cycle. Towards the east these shallow-water deposits interfinger with the offshore clays of the Zellerndorf Formation. During the Ottnangian, due to further westward transgression of the sea onto the Bohemian Massif, these offshore clays were deposited above the nearshore sediments. Towards the west a transition of the marine clays to the brackish clays of the Weitersfeld Formation can be noted. In marginal areas brackish-estuarine sediments with coal seams were deposited (Langau Formation) during the Ottnangian and were followed by marine nearshore sands (Riegersburg Formation). In several of these marginal Ottnangian units, acidic volcaniclastics, derived from the Carpatho-Pannonian region, are present (Nehyba & Roetzel 1999).

In the Late Eggenburgian and Ottnangian the reopening of the NAFB towards the Rhône Basin in the west led to a distinct change in the depositional environments. Consequently, in the western part of the NAFB sands and silts of the Innviertel Group show evidence of strong tidal influence, ubiquitous reworking, and submarine erosion (Faupl & Roetzel 1988). Close to the margin of the Bohemian Massif, reworking of Egerian and Eggenburgian sediments is reflected in the presence of the phosphorite-bearing Plesching Formation (Faupl & Roetzel 1990). In the eastern part of the NAFB, basinal clays, silts and sands of the Robulus Schlier and the so-called Sandstreifenschlier represent Eggenburgian to Ottnangian offshore deposits. In the Ottnangian, at the margin of the Bohemian Massif, submarine debris flows occur (Mauer Formation). These interdigitate with tidal channel sands of the Prinzersdorf Formation (Krenmayr 2003*a*, *b*).

In the Late Ottnangian, the presence of the widespread *Rzehakia* ('*Oncophora*') beds reflects a major regression. These beds yield an endemic mollusc fauna with bivalves such as *Rzehakia* and *Limnopagetia*, which thrived in shallow brackish lakes. In the western NAFB, a pronounced phase of erosion took place up to the latest Early Miocene. Subsequently, deposition of limnic-fluvial sediments of the Upper Freshwater Molasse commenced. Its basal parts comprise lignite-bearing clays, sands and gravels (Trimmelkam and Munderfing beds), which are of Middle Miocene age. After another gap the 'Lignite bearing Freshwater beds', the Kobernausserwald gravels and the Hausruck gravels were deposited in limnic-fluvial environments during the Late Miocene.

In the eastern part of the NAFB shallow-marine conditions also prevailed in late Early Miocene times and are represented by the pelitic and sandy sediments of the Laa and Nový Přerov formations (Roetzel 2003). In the early Middle Miocene, similar sediments were deposited (Grund and Gaindorf formations). Sandy-gravelly and shelly intercalations within the predominantly pelitic Grund Formation are interpreted as storm-induced event deposits (Roetzel & Pervesler 2004). On topographic highs corallinacean limestone developed (Mailberg Formation). To the south these sediments are correlated with submarine deltaic conglomerates (Hollenburg-Karlstetten Formation). Foraminiferal data document an Early Badenian age (= Langhian) for these deposits and suggest that marine deposition in the eastern NAFB continued until *c.* 14.5 Ma, when Middle Miocene uplift caused the sea to retreat.

The very last marine ingression into the already dry basin took place during the Early Sarmatian (= late Serravallian). The marine Sarmatian is confined to a rather narrow, *c.* 40 km long west–east trending trough extending from the Bohemian Massif in the west to the Vienna Basin in the east. The location was controlled by an older incised valley which became flooded during the Early Sarmatian. Clays, silts and sands were deposited suggesting the formation of extended sandy to muddy tidal flats. The dating of the Ziersdorf Formation as Early Sarmatian is based on the occurrence of several species of the endemic gastropod *Mohrensternia* (Kowalke & Harzhauser 2004) and rare *Elphidium reginum* (Papp *et al.* 1974). Upper Sarmatian deposits are absent from the NAFB due to the final retreat of the Paratethys Sea from that area.

The early Late Miocene of the NAFB in Lower Austria is characterized by a huge fluvial system often referred to as the Palaeo-Danube. The corresponding deposits of the Hollabrunn-Mistelbach Formation indicate a dominant gravel-rich fluvial depositional environment, which grades into a braid-delta environment towards the east at the entrance to the Vienna Basin (Nehyba & Roetzel 2004). The mostly coarse-grained clastic fluvial to deltaic sediments, extend, on the surface, in a WSW–ENE direction from Krems in the NAFB towards the Steinberg Fault in the Vienna Basin over a distance of >86 km. The width of this sediment body is between 3 and 14 km, and up to 20 km in the delta area.

Austro-Alpine Gosau basins (M.W.)

Palaeogene strata in the Austro-Alpine parts of the Eastern Alps (Northern Calcareous Alps (NCA) and Central-Alpine Zone (CAZ)) (Fig. 17.13) are partly continuous from the Upper Cretaceous successions. Following the work of Kühn (1930), Palaeogene-age strata have been recognized from the Gosau Group of the NCA. Most of the NCA Palaeogene deposits are deep-marine (Wagreich & Faupl 1994; Wagreich 2001); shallow-water sediments have only been reported from the southeastern part of the NCA (Kambühel and Hochschwab areas; Tragelehn 1996), and from the CAZ (Krappfeld area; Wilkens 1989; Rasser 1994). Shallow-water carbonates are also found as olistoliths in deep-water strata (Lein 1982; Moussavian 1984).

Outcrops of the Palaeogene Gosau Group comprise only erosional remnants of the widespread Palaeogene cover of the NCA and the CAZ as evidenced by the widespread redeposition of Palaeogene sediments into younger formations (e.g. Hagn 1981). The Palaeogene deposits of the Gosau Group record the geodynamic evolution of the Eastern Alpine orogenic wedge from a phase of deep-water sedimentation to renewed thrusting, which culminated in the meso-Alpine Orogeny related to compression between the European and the African-Adriatic plates during the Palaeogene (e.g. Dewey *et al.* 1989; see also Froitzheim *et al.* 2008).

Palaeogeography

The Palaeogene deposits of the Gosau Group of the NCA record deep-water sedimentation on top of the early orogenic wedge of the Eastern Alps, which evolved during Early to early Late Cretaceous Alpine deformation (Faupl & Wagreich 2000) (Fig. 17.14). Palaeogeographic reconstructions indicate a generally northward-deepening slope, dissected by depocentres and structural highs forming slope basins along an active continental margin. To the south of the NCA, exhumation of metamorphic complexes of the Austro-Alpine basement formed a rising source area for Palaeogene siliciclastics. This rising hinterland separated the depositional area of the Gosau Group of the NCA from southern CAZ basins such as the Krappfeld (Neubauer *et al.* 1995; Wagreich & Siegl-Farkas 1999). Palaeogene deep-water sediments, comparable to the Gosau Group, are also known from the eastward continuation of the NCA into the Western Carpathians (Wagreich & Marschalko 1995). Palaeogene deposition within the CAZ can be interpreted as a continuation of the 'Central Carpathian Palaeogene' deep-water trough from Slovakia (Wagreich 2001; Kázmér *et al.* 2003) into the Eastern Alps. Farther to the south, a marine seaway from the 'Central Alpine' Gosau Basin into the Southern Alps can be assumed, due to the occurrence of Palaeocene–Eocene deep-water strata, in, for example, the Lombardian Basin in northern Italy, where an Early Palaeocene hiatus is followed by the deposition of turbidites and marlstones of Palaeocene–Middle Eocene age (Bersezio *et al.* 1993).

Sedimentary and stratigraphic development

Formally defined lithostratigraphic subdivisions of Palaeogene deposits of the Austro-Alpine units have been established in the Gosau area (Fig. 17.15) (based on Weigel (1937) and Kollmann in Plöchinger (1982): Nierental Formation, Zwieselalm Formation), in the Gießhübl syncline (Gießhübl Formation; Plöchinger 1964; Wessely 1974; Sauer 1980) and in the Grünbach-Neue Welt area (Zweiersdorf Formation; Plöchinger 1961). Paleocene shallow-water carbonates of the Kambühel Formation, as defined by Tollmann (1976), were investigated by Tragelehn (1996) who distinguished two members (St. Lorenzen Member, Ragglitz Member). Biostratigraphic data for the deep-water successions are mainly based on planktonic foraminifera and calcareous nannoplankton (e.g. Hillebrandt 1962; Wille-Janoschek 1966; Wille 1968; Kollmann 1964; Wagreich & Krenmayr 1993; Hradecká *et al.* 1999; Egger *et al.* 2004).

The Palaeogene in the CAZ is subdivided into the Holzer Formation (Paleocene), the Sittenberg Formation (Ypresian/Lutetian) and the Dobranberg Formation (Lutetian; Wilkens 1991; Rasser 1994) which includes several members.

Four generalized facies associations have been distinguished within the Palaeogene part of the Gosau Group of the NCA (Wagreich 2001): siliciclastic and mixed siliciclastic/carbonate turbidites, hemipelagites and pelagites, debrites, and shallow-water carbonates of reef, lagoonal and forereef facies. The gravity-flow-dominated facies association suggests deposition on small submarine fans. Proximal fan areas are characterized by channels filled with conglomerates and pebbly sandstones, whereas classical turbidites indicate interchannel and distal fan depositional environments.

During the Early Palaeocene, turbidites and hemipelagites dominated the Gosau Group of the NCA, from the Tyrol in the west (Muttekopf area, Ortner 1994) to the Gießhübl Syncline in the east (Sauer 1980). Several localities (Gosau, Gams, Lattengebirge, Gießhübl Syncline) display a conformable succession of Maastrichtian to Palaeocene deep-water sediments, without major facies changes around the K/Pg boundary (Lahodynsky 1988; Krenmayr 1999). Several hundred metres of thick turbiditic successions are known from Gosau, Gams and the Gießhübl Syncline. Palaeocurrent data point to southern source areas for the turbidites (Faupl 1983), although basin-parallel, east–west trending flows are also recorded. Turbiditic sandstones can be classified as lithic arenites and display mixing of siliciclastic (mainly low- to medium-grade metamorphic clasts) and carbonate debris. Carbonate clasts include both extraclasts from the underlying strata of the NCA, and bioclasts from a contemporaneous carbonate shelf to the south. Ar/Ar dating of micas from pebbles suggests significant erosion of pre-Alpine, Permian metamorphic crystalline units of the Austro-Alpine basement to the south (Frank *et al.* 1998). Hemipelagites and pelagites (Nierental Formation; Krenmayr 1996, 1999) occur either as packages several tens to hundreds of metres thick or as thin intervals within turbidite-dominated successions. Thin- to medium-bedded marly limestones, marls and shales with a generally high degree of bioturbation (*Zoophycus* ichnofacies) predominate. The carbonate content consists mainly of calcareous nannoplankton and planktonic foraminifera. Intercalations of thin sandstone and debrite beds or slump deposits are common (e.g. Krenmayr 1999; Wagreich & Krenmayr 2005).

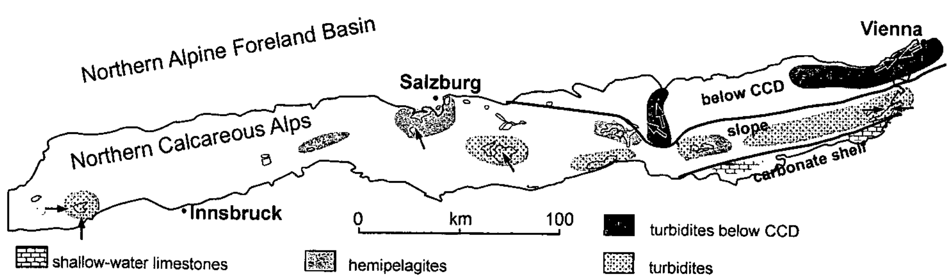


Fig. 17.14. Palaeogeographic sketch map for the Early Palaeocene (nannoplankton zones NP1–4) of the Northern Calcareous Alps with the main Gosau Basins. Arrows indicate palaeocurrent directions.

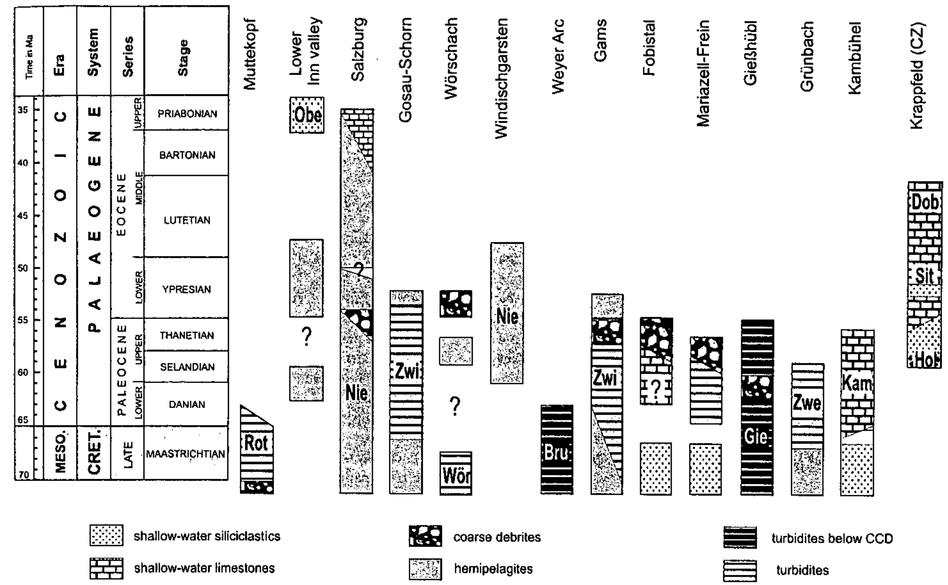


Fig. 17.15. Lithofacies, lithostratigraphy and chronostratigraphic correlation of main successions of Palaeogene sediments of the Gosau Group of the Northern Calcareous Alps and the Palaeogene of the Central Alpine Zone. Abbreviations: Bru, Brunnbach Formation; Dob, Dobranberg Formation; Gie, Gießhübl Formation; Hol, Holzer Formation; Kam, Kambühel Formation; Nie, Nierental Formation; Obe, Oberaudorf Formation; Rot, Rotkopf Formation; Sit, Sittenberg Formation; Wör, Wörschachberg Formation; Zwe, Zweiersdorf Formation; Zwi, Zwieselalm Formation.

Within the Weyer Arc area, carbonate-free hemipelagic clays within the turbiditic succession of the Brunnbach Formation record deposition below the local carbonate compensation depth (CCD) (Faupl 1983). This deep-water basin received material from two sources, building a sand-rich and a sand-poor submarine fan (Faupl 1983). To the south and SE of this deep basin, the Gosau Group of Gams, deposited in a slope basin–trench–slope basin setting, records deposition in bathyal depths during the Early Palaeocene (Wagreich & Krenmayr 1993; Krenmayr 1996; Egger *et al.* 2004). South of Gams, in the Hochschwab area, shallow-water carbonates and limestone olistoliths are present. These remnants of shallow-water facies (Kambühel Formation) continue up to the eastern margin of the Alps (Kambühel near Ternitz). Palaeocene reef carbonates comprise bafflestones/boundstones/rudstones with abundant corallinacean and dasycladacean algae and corals. Forereef facies include packstones rich in corallinacean algae. Third-order transgression–regression cycles, marked by emersion horizons, have been reported by Tragelehn (1996). Reef growth stopped during the Middle Thanetian, probably due to a sea-level fall and tectonism (Tragelehn 1996).

During the Thanetian to Lower Ypresian coarse debrites, including olistoliths of Palaeocene shallow-water carbonates, are widespread in the NCA. They occur either as debrite layers several metre thick associated with mixed siliciclastic-carbonate turbiditic successions (e.g. Gießhübl Formation), or as isolated breccias including olistoliths, up to 50 m in diameter, in the Hochschwab area. Olistostromes, including metamorphic clasts, were reported from the lower 'Ilerdian' (P5) of the Salzburg-Reichenhall area and the Kaisergebirge/Tyrol (Moussavian *et al.* 1990). The southern carbonate platform was probably dissected by canyons, enabling the siliciclastics to bypass the carbonate environment.

The occurrence of bentonites, originating from airfall, in the Upper Palaeocene (NP10) of Salzburg and Gams (Egger *et al.* 1996, 2004) suggest that there was a close connection of the depositional areas of the Gosau Group and parts of the Rhenodanubian Flysch, where similar bentonites of basaltic composition are known (Egger *et al.* 2000).

In the CAZ a hiatus marks the top of the Late Maastrichtian to Early Palaeocene sequence. This is followed by a (Late?) Palaeocene succession of terrestrial conglomerates, sandstones and clays including coals (Holzer Formation; Wilkens 1991).

Eocene deposits of the Gosau Group are known from Gams (Kollmann 1964; Egger & Wagreich 2001), Windischgarsten, the Schorn area near Gosau and Abtenau (Wille 1968), the Salzburg-Untersberg area (Hillebrandt 1962, 1981), and the Lower Inn Valley (Hagn 1981). An Early to Middle Eocene age (NP12–15) for the end of sedimentation has been reported from most of these areas; only the succession in the Salzburg–Untersberg area displays a significantly younger interval up to the Late Eocene (P15–16, NP19). This unusually high stratigraphic range may be explained by a transition from the Gosau Group sedimentary cycle to the Late Eocene/Oligocene sedimentary cycle of the 'lower Inn valley Tertiary' (Ortner & Sachsenhofer 1996; Löffler & Nebelsick 2001; Nebelsick *et al.* 2001; Ortner & Stingl 2001), a complex pull-apart-piggyback basin which was connected to the foreland Molasse Basin. Major facies types of Eocene–Oligocene shallow-water carbonates of the circum-Alpine area were recently summarized by Nebelsick *et al.* (2003, 2005).

Due to renewed northward thrusting onto the Rhenodanubian Flysch Zone and Helvetic Units during the Middle/Late Eocene, large parts of the NCA were subaerially exposed and marine sedimentation terminated in most of the Gosau basins. At the tip of the NCA wedge, marine sedimentation continued within piggyback basins, which subsided during ongoing thrusting of the NCA.

In the CAZ, a transgression around the Palaeocene–Eocene boundary resulted in the deposition of marginal marine siliciclastics and carbonates. Late Palaeocene to Early Eocene clastics and nummulite marls are represented by the Sittenberg Formation. The overlying Dobranberg Formation marks a transition to pure larger foraminiferal limestones. Basal parts of this formation are characterized by intercalations of alveolinid and nummulitid limestones with local intercalations of orthophragminid foraminifera (Wilkens 1989, 1991; Hillebrandt 1993). The nummulitid foraminifera are Early Eocene in age (Hillebrandt 1993). Coralline algae and encrusting foraminifera dominate the middle part of the succession where they form huge accumulations of rhodoliths and acervulinid macroids (Rasser 1994) which are thought to be Ypresian to Lutetian in age (Wilkens 1989; see also Moussavian 1984).

Vienna Basin and its satellite basins (M.H., M.K., R.R.)

The Vienna Basin covers large parts of eastern Austria (Lower Austria, Vienna and Burgenland) and extends into the Czech Republic in the north and the Slovak Republic in the east (Fig. 17.16). It is about 200 km long and 55 km wide, striking roughly SW–NE from Gloggnitz (Lower Austria) in the SSW to Napajedla (Czech Republic) in the NNE. As a classic Neogene basin it has been the subject of hundreds of geoscientific studies since the early nineteenth century.

Tectonic setting and development

The Vienna Basin is a rhombic Neogene-age pull-apart basin. Its SW border is formed topographically by the Eastern Alps and to the NW by the Waschberg and Ždánice units. To the east it is bordered by the Rosalia, Leitha and Hainburg hills, and the Male Karpaty Mountains, all four of which are part of the Alpine-Carpathian Central Zone. The Pieniny Klippen Belt represents an internal boundary of the Outer Carpathian Flysch Belt; sediments of the Magura Unit form the northern margin of the basin. The basement of the basin is formed by Alpine-Carpathian nappes. The maximum thickness of the Neogene basin fill is 5500 m. Since the basin is subdivided by a morphological high, the Spannberg Ridge, into a northern and a southern part, marine sedimentation was restricted to the north (north of the Danube) during the Early Miocene and extended into the south only during the Middle and Late Miocene. Due to the complex fault system the basin was internally subdivided into a series of horst and graben systems. The uplifted blocks at the margins of the basin are separated from the deeper areas by major faults (e.g. Mistelbach Block and Steinberg Fault, Moravian central depression and Bulhary Fault in the northern basin, Mödling Block and Leopoldsdorf Fault in the southern basin; Láb-Malacky High and Leitha and Láb fault zones).

A detailed overview including all relevant literature used in the present section was presented by Kováč *et al.* (2004). The formation of the Vienna Basin began in the Early Miocene as an east–west trending piggyback basin on top of the Alpine thrust belt. It was initiated during the Eggenburgian and was active until the late Early Miocene (Early Karpatian). In the late Early Miocene thrusting was replaced by the lateral extrusion of the Western Carpathian lithospheric fragment from the Alpine Realm and depocentres originated by pull-apart processes. During the latest Early Miocene, NE–SW orientated deep sinistral strike-slip faults were formed along the eastern margin of the basin, together with north–south orientated normal faults. The Middle Miocene subsidence of the synrift stage of the Vienna Basin was controlled by a palaeostress field with NE–SW orientated compression (NW–SE extension). The development of the basin during the Middle Miocene was influenced by NE–SW orientated normal faults. A second phase of more rapid tectonic subsidence during the late Middle Miocene (Early Sarmatian) is related to ENE–WSW sinistral strike-slip faults and NE–SW orientated normal faults. These faults induced subsidence of the Zistersdorf–Moravian Central Depression. Synrift extension in the northern part of the Vienna Basin was enhanced by active elongation of the Western Carpathian Orogen during the Sarmatian due to subduction in front of the Eastern Carpathians. The Late Miocene represents the post-rift stage in basin evolution. In the Late Miocene (Pannonian) and Pliocene, the Vienna Basin was inverted and subsequently only minor amounts of sediment were deposited. Fault-controlled subsidence in grabens at the eastern margin of the basin (Zohor-Plavecký Mikuláš and Mitterndorf grabens) documents a sinistral transtensional regime of this zone, lasting up to recent times, and accompanied by seismic activity.

Sedimentary and stratigraphic development

The initial phase of Miocene deposition in the present Vienna Basin (Fig. 17.9) was related to the Eggenburgian transgression and to the tectonic opening of depocentres in its northern part. Piggyback basin depocentres developed in the Outer Carpathian Flysch Belt zone, while in the Central Western Carpathians wrench fault basins opened. Deposition commenced with the clays and sands of the fluvial Stráže Formation. The onset of marine transgression is reflected by the boulder-sized Brezová conglomerates which pass into fine-grained conglomerates and shoreface sands. Upwards, sandy deposits containing a rich pectinid fauna are found. Laterally, to the south and east, the coastal facies passes into open-marine conditions marked by the upper part of the Lužice Formation. Neritic sands and clays (= 'Schlier') contain a rich deep-water foraminiferal assemblage, yielding taxa that tolerated low-oxygen bottom conditions. The

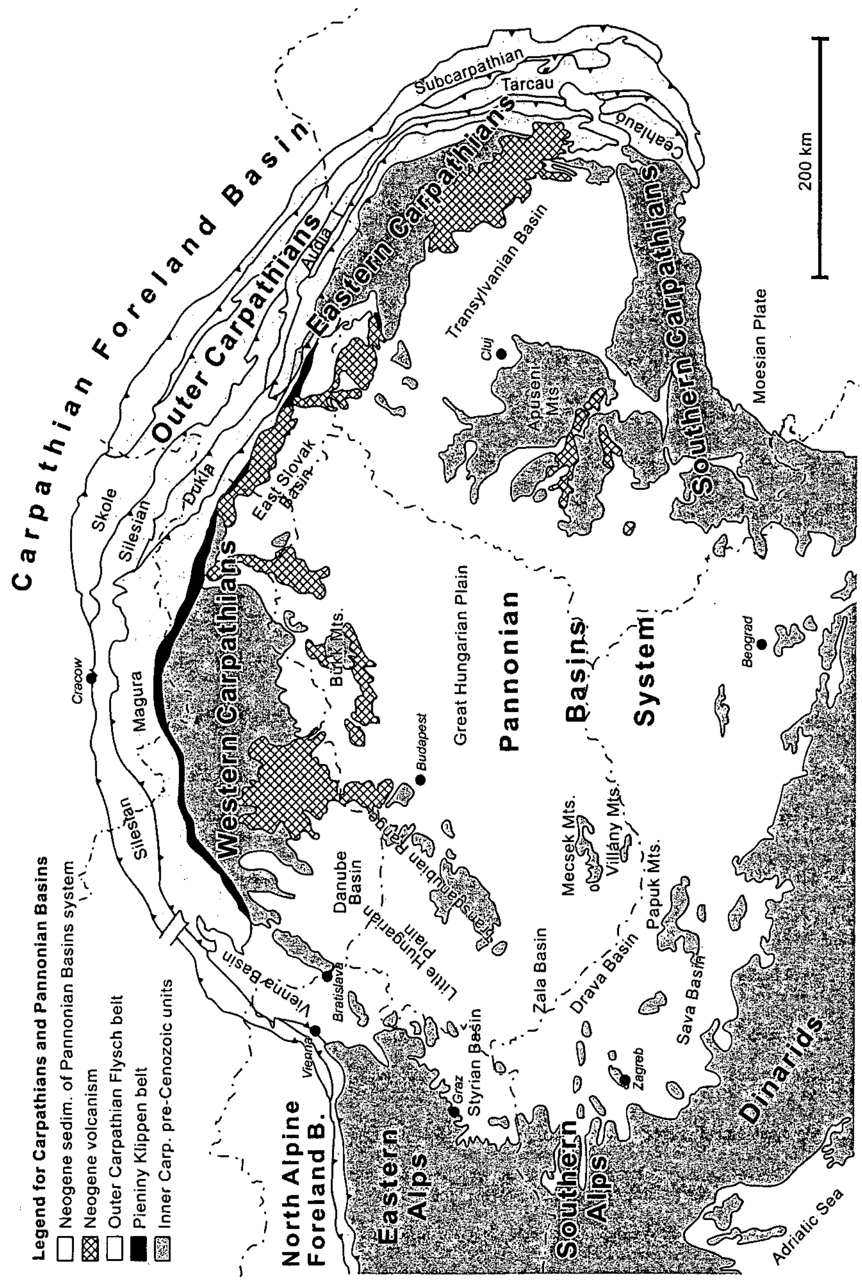


Fig. 17.16. Geology of the Carpathians and the Pannonian Basins System.

Eggenburgian–Ottnangian boundary is marked by a relative sea-level fall and recorded by the basinward progradation of shallow-water sandy facies at the base of the Ottnangian. In the northern Vienna Basin, the sandy Štefanov Member (150 m) represents a deltaic body entering the basin from the SW during this regression. The overlying Ottnangian transgression is represented by the silts and silty sands of the upper Lužice Formation, which onlaps the slopes of topographic highs.

During the subsequent Karpatian a pull-apart basin began to open. Rapid subsidence and sea-level rise led to the development of offshore settings in the northern Vienna Basin, as reflected by the pelites of the Laa and Lakšáry formations. Sedimentation in the southern part of the basin differed significantly, due to the presence of a topographic barrier formed by the Spannberg Ridge in the central Vienna Basin. Thus, in the southern Vienna Basin, sedimentation started during the Early Miocene with the deposition of the alluvial Bockfliess Formation comprising lacustrine to brackish-littoral environments. After a regressive phase at the end of the Early Karpatian this formation was discordantly overlain by the lacustrine-terrestrial facies of the Gänserndorf Formation. At the same time, up to 400 m of sandy deltaic deposits (Šaštín Member) prograded into the Slovak part into the Vienna Basin. The top of the Gänserndorf Formation grades into the overlying Aderklaa Formation without a major unconformity. Sandstones, interbedded pelites and rare conglomerates characterize the deposits of the 1000 m thick Aderklaa Formation. Deposition took place in a limnic/fluvial environment as part of a meandering river system. The Gänserndorf and Aderklaa formations in the southern Vienna Basin can be correlated with the marine, brackish to freshwater Závod Formation in the northern Vienna Basin.

A major regressive event at the Lower–Middle Miocene (Karpatian–Badenian) boundary, found in many Paratethyan nearshore settings, is also recorded in the Vienna Basin by erosional truncations of up to 400 m. The Vienna Basin became subaerial during this time, as reflected by palaeosol formation. In the south, sedimentation recommenced during the early Middle Miocene with the deposition of the Aderklaa Conglomerate, which formed part of a braided river system. Similar conditions are documented by the Jablonica Conglomerate in the north. During the subsequent Badenian transgression offshore pelites (Baden Group) were deposited. This group consists of several formations, including the Lanžhot Formation in the Slovak part of the basin. Corallinacean limestones ('Leitha limestone') frequently formed in areas with little clastic input. The first occurrences of *Praeorbulina* and of *Orbulina suturalis* within the Baden Group are important biostratigraphic markers (Rögl *et al.* 2002).

During the early Badenian at *c.* 14.2 Ma a major sea-level fall occurred. Several small deltaic bodies developed (e.g. Andlersdorf, Zwerndorf and Auersthal members). A second Badenian sequence followed. Deltaic, fluvial and lagoonal settings developed in the northern part of the Vienna Basin (Žižkov Formation). This mostly freshwater and brackish formation consists of calcareous clays, containing lenses of cross-bedded sand bodies (*c.* 1200 m in the Moravian Central Depression). This deltaic succession passes upward into littoral sandy clays with a marine fauna. The following relative sea-level rise is expressed by the deposition of transgressive shelf sand bars and a backstepping of the deltaic sediments. The littoral sands were subsequently overlain by the calcareous clays of the neritic offshore Jakubov Formation.

A third Badenian succession includes littoral and sublittoral shoreface sands with coralline algal biostromes at their base. A distinct flooding surface is preserved near Devínska Nová Ves in Slovakia. In the central part of the basin, layers of Upper Badenian corallinacean limestones represent a shallow-marine shoal along the Spannberg Ridge that extended for more than 150 km^2 (Kreutzer 1978). This topographic high acted as a Late Badenian platform from which a bird-foot delta spread into the southern basin (Weissenbäck 1996). This delta was supplied with sediments from the continental North Alpine Foreland Basin. The offshore sediments consist of marine calcareous clays (Studienka Formation). Oxygen-depleted bottom conditions commonly occurred in basinal settings but also on the carbonate platforms (Schmid *et al.* 2001).

The Badenian–Sarmatian boundary is characterized by a major fall in sea level. Badenian-age corallinacean limestones were exposed and significant erosion took place along the margins of the basin. Renewed transgression in the Early Sarmatian filled incised valleys. This transgression is represented mainly by calcareous clays, silts, and rare acidic tuffs (Holíč Formation; Vass 2002). The lowermost Sarmatian deposits are recorded from the Kúty and Kopčany grabens in the northern Vienna Basin. Fluvial gravel was derived from the Northern Alpine Foreland Basin. In coastal settings unique bryozoan–serpulid–algal bioconstructions flourished, forming reefoid structures several metres high. These are well preserved along the Malé Karpaty and Leitha mountains in the eastern and southern Vienna Basin and along the Steinberg Ridge. A subsequent regressive phase in the Middle Sarmatian was characterized by the progradation of a huge delta complex in the Matzen area in the central Vienna Basin and by erosion of Lower Sarmatian bioherms. The renewed flooding during the Late Sarmatian is reflected by the various mixed siliciclastic/carbonate deposits of the Skalica Formation (Vass 2002). Extended oolite shoals (up to 30 m thick) and coquina shell beds formed in the entire Vienna Basin (Harzhauser & Piller 2004).

At the Sarmatian–Pannonian boundary, the Paratethys Sea retreated from the Pannonian Basin area and the brackish Lake Pannon was established. The Vienna Basin was dry and Middle Miocene deposits were eroded. Consequently, lowermost Pannonian fluvial facies penetrated far into the basin, reworking older Sarmatian strata. A rather uniform, 50–100 m thick, monotonous unit of marl and sand of prodelta- and basinal facies follows and includes a 20–50 m thick marker unit of ostracod-bearing marly clay at the top. The marly prodelta facies is followed by a sandy succession up to 200 m thick with rare gravels representing a prograding delta of the palaeo-Danube in the northwestern part of the Vienna Basin. The occurrence of the three-toed horse *Hippotherium* within those deposits is an important biostratigraphic marker.

A major transgression followed and the elevated highs along the western margin of the basin were flooded by Lake Pannon (Harzhauser *et al.* 2004). The Middle Pannonian deposits consist of *c.* 440 m of clays and sands (Bzenec Formation = Inzersdorf Tegel). The Upper Pannonian is characterized by the ubiquitous occurrence of thin lignite seams in its basal parts and by a 200 m thick sandy/marly upper part (Čáry Formation). During the Late Pannonian, the margin of Lake Pannon had retreated from the Vienna Basin. Floodplain deposits and freshwater lakes developed which were not connected to Lake Pannon. The basin fill terminates with a 450 m thick succession of marls, clays and silts with intercalations of sands, gravels, rare lignites and freshwater limestones (Pannonian Gbely and Pliocene Brodské formations). In the southern Vienna Basin, Miocene sedimentation terminates with the fluvial Rohrbach Conglomerate.

The Korneuburg (Sub-)Basin formed due to pull-apart activity within the Alpine-Carpathian thrust belt. This asymmetric SSE–NNE orientated basin is *c.* 20 km long and attains a maximum width of 7 km, but is strongly narrowed in its northern extension. A central high separates the southern part (*c.* 880 m deep) from a shallower northern depocentre (*c.* 530 m deep).

The basin margins are formed in the north by the Waschberg Unit and towards the south by the Rhenodanubian Flysch Unit. These Alpine-Carpathian nappes are underlain by the autochthonous basement formed mainly by Upper Cretaceous and Jurassic units and by the crystalline basement of the Bohemian Massif. Sedimentation began during the early Miocene (Eggenburgian) and comprised shallow-marine marls and sands (Ritzendorf Formation). The main phase of deposition, however, began in the late Early Miocene (Karpatian), and is represented by marly silts and fine to medium sands (Korneuburg Formation).

A connection to the Paratethys was only warranted along the northern basin margin, where the sea extended into the Alpine-Carpathian Foredeep. This situation is also reflected in the internal facies patterns. Thus, the small, elongated basin was divided into a southern, estuarine part with extensive tidal mudflats and *Crassostrea* bioherms, and a northern shallow-marine part with depth of 20–30 m. No Middle or Upper Miocene deposits are known from the Korneuburg Basin (all data from Harzhauser *et al.* 2002; Harzhauser & Wessely 2003).

The Eisenstadt-Sopron (Sub-)Basin is more or less triangular in shape and measures about 20 km. In the north it is bounded by the NE–SW trending Leitha Mountains and the associated SE-dipping Eisenstadt Fault. In the east, the basin is bordered by the north–south trending Köhida Fault. The Rust-Fertörakos Mountains separate the basin from the Danube Basin in the east. A crystalline ridge, covered by Lower Miocene gravels, extending from the Rosalia Mountains to the Brennberg, defines the southern margin. This topographical barrier also separates the Eisenstadt-Sopron Basin from the Styrian Basin. The development of the Eisenstadt-Sopron Basin is closely linked with that of the Vienna Basin, although the thickness of the basin fill is much less (*c.* 1500 m).

The oldest Neogene deposits in the present Eisenstadt-Sopron Basin are of Early Miocene age. They comprise terrestrial, fluvial and lacustrine deposits which are genetically related to the fluvial system in the southern Vienna Basin. This suggests that the Leitha Mountains did not exist as a barrier at that time and that basin development was initiated not before the Middle Miocene. The onset of subsidence in the Early Badenian led to an initial marine ingression and the Leitha Mountains became a peninsula connected with the Alpine mainland in the east. Nearshore deposits of this transgression are represented along the SE margin of the Leitha Mountains by the Hartl Formation which grades from reworked gravel, through marine sandwaves into corallinacean debris. During periods of high sea level in the Middle and Late Badenian, the Leitha Mountains were completely covered by water allowing the growth of thick corallinacean limestones and coral carpets.

Following a sea-level fall at the Badenian–Sarmatian boundary, the Leitha Mountains and their Badenian sedimentary cover became exposed and the mountain ridge once again became an island again until the withdrawal of Lake Pannon during the Late Miocene. Incised valleys developed and intensively eroded the Middle Miocene limestone platforms. Lower Sarmatian deposits from the Eisenstadt-Sopron Basin consisting of pelites, sands and serpulid limestones can be correlated with the Holíč Formation of the Vienna Basin. Correspondingly, the Upper Sarmatian Skalica Formation of the Vienna Basin extends into the Eisenstadt-Sopron Basin. A mixed siliciclastic/carbonate succession comprising gravels, sands, oolitic sands and marls is typical throughout the basin. Lake Pannon covered the Eisenstadt-Sopron Basin during the Late Miocene, as shown by the presence of clayey marls and sands. During the middle Pannonian a small river formed on the Leitha Mountains supplying reworked Lower Miocene gravels into the basin. As in the Vienna Basin, the withdrawal of Lake Pannon allowed the establishment of floodplains and swamps during the latest Pannonian (all data from Schmid *et al.* 2001; Kroh *et al.* 2003).

Styrian Basin and Neogene intra-Alpine basins (M.H.)

Although situated along the eastern margin of the Eastern Alps the Styrian Basin (SB) is a sub-basin of the Pannonian Basin System (Fig. 17.16). The SB, situated in SE Austria, is *c.* 100 km long and 60 km wide. It is bordered in the west, north, and east by Alpine mountains such as the Koralpe, the Gleinalpe and the Wechsel. The Southern Burgenland High, consisting on the surface of a SW–NE trending range of low hills, represents the southern border. In the Early and Middle Miocene, the eastern part of the basin was covered by the Paratethys and by Lake Pannon during the Late Miocene. In the western part of the basin, swamps formed during the Early Miocene, giving rise to thick lignites, which have been exploited up to recent times.

Tectonic setting and development

The SB is a small extensional basin located on top of an eastward-moving crustal wedge, which contains *c.* 4 km of Neogene sediments. Basement comprises crystalline and low-grade metamorphic Palaeozoic rocks of the Austro-Alpine nappe system. The tectonic evolution is divided into an Early Miocene (Ottnangian to Karpatian) synrift phase and a subsequent post-rift phase. The SW–NE trending South Burgenland High separates the SB from the Pannonian Basin. Internally, the SB is subdivided by the Middle Styrian and Auersbach highs into several small sub-basins, including the shallower Western Styrian Basin and the deeper Eastern Styrian Basin complex, which consists of the Mureck, Gnas and Fürstenfeld basins.

Sedimentary and stratigraphic development

Sedimentation commenced during the Early Miocene (Ottnangian) (Fig. 17.9). In the Western Styrian Basin fault-controlled limnic-fluvial deposits of the Eibiswald Member formed thick lignite deposits. In the Western Styrian Basin Miocene sedimentation ended with the Middle Miocene limnic-fluvial Stallhofen Formation.

In the Eastern Styrian Basin, Ottnangian floodplain and coastal-plain deposits were overlain by a thick Karpatian-age succession deposited following a marine ingression from the Pannonian Basin. Due to the extremely high subsidence rate (30 cm/100 a), related to synsedimentary fault tectonics, the shallow-marine setting evolved rapidly into a deep-marine one (Sachsenhofer 1996). The corresponding 'Steirscher Schlier', a deep-marine shale succession with intercalated turbiditic sandstones and tuffs, formed in the upper bathyal zone under dysaerobic conditions (Spezzaferri *et al.* 2002). Coastal areas along the emerging Alps were affected by strong fluvial input of coarse clastics. Synchronous with the synrift phase, andesitic island-arc volcanism commenced and formed huge shield volcanoes. The andesitic and shoshonitic volcanism continued into the early Middle Miocene.

Towards the end of the Early Miocene, uplift (i.e. Styrian Tectonic Phase) led to basin shallowing and finally to the tilting of older deposits. In marginal areas considerable erosion took place and Middle Miocene deposits are separated by a distinct unconformity from Lower Miocene ones. A renewed marine ingression during the Early Badenian led to the establishment of shallow-marine conditions with widespread development of patch-reefs and corallinacean limestones (Weissenegg Formation) (Friebe 1990). Corallinacean platforms developed along the shallow swells, and sublittoral to fairly deep-water marly and pelitic sediments were deposited in the deeper parts of the SB. According to Friebe (1993), the Badenian of the Styrian Basin can be subdivided into three marine sequences which are related to global sea-level cycles.

A major drop in the relative sea-level occurred at the Sarmatian–Badenian boundary. Fault-controlled subsidence rates increased during the Sarmatian with the extensive deposition of marls. On topographic highs (e.g. South Burgenland High) and along the coasts, bryozoan–serpulid bioconstructions formed isolated carbonate bodies (Grafenberg Formation). Marine sedimentation was interrupted in the middle Sarmatian by the deposition of up to 100 m of fluvial sands and gravels (Carinthian gravel). The overlying Upper Sarmatian units consist of the mixed siliciclastic/carbonate Gleisdorf Formation, which contains several oolitic beds. The upper part of the Upper Sarmatian is characterized by repeated intercalations of thin lignites and by a marker horizon with peneroplid foraminifera.

In the Late Miocene, the SB became flooded by Lake Pannon, and marls and silty sands of the Feldbach Formation were deposited. The overlying fluvial-limnic Paldau Formation represents a second Pannonian sequence. The Upper Pannonian follows discordantly and comprises fluvial and limnic deposits reflecting a total separation from Lake Pannon. A second magmatic phase is represented by tuff pipes and basaltic lava flows of Plio/Pleistocene age.

An overview of the tectonic evolution of the Styrian Basin has been provided by Sachsenhofer (1996); detailed overviews of stratigraphy and depositional environments have been published by Kollmann (1965), Friebe (1993), Gross (2003), Kosi *et al.* (2003) and Harzhauser & Piller (2004).

The **Fohnsdorf Basin**, 22 km long and 11 km wide, formed to the NW of the Styrian Basin at the junction of two strike-slip fault systems (Sachsenhofer *et al.* 2000; Strauss *et al.* 2003). These fault systems, the sinistral east–west trending Mur-Mürz-Fault System and the dextral NNW–SSE trending Pöls-Lavanttal-Fault System, form the border of the escaping crustal wedge which hosts the Styrian Basin. The evolution of the basin, with a Neogene basin-fill of about 3 km, was subdivided by Strauss *et al.* (2003) into three stages. The initial pull-apart phase (stage 1) lasted from the Late Karpatian to the Early Badenian and commenced with the deposition of the Fohnsdorf Formation. This is characterized by up to 800 m of alluvial sediments which terminate in a 15 m thick lignite seam which yields a typical '*Congeria*'-coquina. Lacustrine to brackish prodelta and fan sediments were deposited in the late pull-apart phase and constitute the overlying 2000 m thick Ingering Formation. The brackish conditions in the lower part of the Ingering Formation suggest a connection to the marine flooding of the Lavanttal Basin (see below) in the Early Badenian. During the Badenian, the basin experienced a half-graben stage (stage 2) and was covered by floodplain and lacustrine fan delta deposits. These immature conglomerates and sandstones constitute the Apfelberg Formation (Strauss *et al.* 2003). The final compressive phase (stage 3) began in the Late Miocene and was heralded by inversion and movement along the Pöls-Lavanttal Fault System.

The fault system controlling the Fohnsdorf Basin extends south to the **Lavanttal Basin**. This basin, situated west of the Styrian Basin, is a pull-apart basin located between the crystalline of the Sauaple and the Koralpe. Its development began in the Early Miocene with the formation of the *c.* 12 km long west–east trending Granitztal Sub-basin. Fluvial clastics and limnic clays of Ottnangian and Karpatian age were deposited. At the Early–Middle Miocene boundary the basin geometry changed considerably due to activation of the Pöls-Lavanttal Fault System resulting in the formation of a 27 km long NNW–SSE trending basin. A diverse mollusc and foraminifera fauna in the marls of the Lower Badenian Mühldorf Beds is indicative of a marine ingression. This short-lived connection to the Paratethys ended during the Middle and Upper Badenian when fluvial-lacustrine and continental environments became dominant. A final transgression from the east took place during the Early Sarmatian. Shallow-marine to paralic conditions prevailed; rare lignite seams within marls are typical deposits. Following a Middle Sarmatian hiatus, limnic pelites with lignites are found in the Upper Sarmatian, whilst the Pannonian is characterized by fluvial deposits. A Pliocene magmatic phase, represented by the Kollnitz basalt, would appear to be related to a synchronous volcanic phase in the Styrian Basin (Tollmann 1985; Strauss *et al.* 2003).

Carpathians, Carpathian Foredeep and Pannonian Basins System: overview (M.K.)

Despite being a part of the Alpine-Carpathian Orogen (Figs 17.2 & 17.16), the Carpathians are very different from the Alps, mainly due to the presence of broad Neogene basins and extensive acidic to calc-alkaline volcanic activity. The difference is caused by the tectonic evolution of the Carpathians, with a transition from 'A-type' subduction during frontal collision with compressional regime during the earliest Miocene, to oblique collision with the European platform and 'B-type' subduction during the remainder of the Neogene (Tomek & Hall 1993; see also Froitzheim *et al.* 2008).

The subduction in front of the orogen caused folding and nappe thrusting, which resulted in the development of an accretionary wedge of the Outer Carpathians. In the foreland, the Carpathian Foredeep developed on the slopes of the European Platform, due to the deep subsurface load of the downgoing plate and the loading of the accretionary wedge. The foredeep shows a significant pattern of depocentre migration from NW to SE (Meulenkamp *et al.* 1996).

Subducting slab pull and the subsequent stretching of the overriding plates were followed by basin opening in the Pannonian backarc area under an extensional tectonic regime (Royden 1988). In addition to the pull of the sinking slab and the subsequent stretching of the overriding plate, the evolution of the Pannonian backarc basins system was influenced by asthenospheric mantle diapirism and by related deep structural unroofing of the basement units (Tari *et al.* 1992; Horváth 1993). During the initial rifting and synrift stage of the backarc system, two types of basins developed: pull-apart basins (Vienna Basin, East Slovakian Basin, Derecke Basin) and grabens or half-grabens (Danube Basin, North Hungarian–South Slovakian Basin, Styrian Basin). The youngest late Miocene to Pliocene period is characterized by thermal post-rift subsidence, but in many places fault-controlled basin subsidence also occurred (Horváth & Cloetingh 1996). During Pliocene times, tectonic inversion of the backarc basin began.

For a fuller understanding of the Neogene geodynamic evolution of the Carpathian Orogen it is necessary to understand that the Outer Carpathian belt is an external tectonic unit, while the internal units are parts of two consolidated, palaeo-Alpine lithospheric fragments or microplates (Alcapa and Tisza-Dacia microplates). These microplates have a very complex movement trajectory relative to their present-day positions (e.g. counter-clockwise rotation and movement of *c.* 150–300 km of the Alcapa microplate and clockwise rotation of the southern Tisza-Dacia microplates (Csontos *et al.* 1992; M. Kováč *et al.* 1994; Kováč & Márton 1998).

Five orogenic and basin evolution stages have been recognized in the Western Carpathians, each of which is characterized by its own tectonic regime (Konečný *et al.* 2002). (1) The compressional regime is related to 'A-type' subduction at the Alpine-Western Carpathian orogenic front during the Early Miocene (Eggenburgian to Ottnangian). (2) The transpressive–transtensional tectonic regime is related to the escape of the Western Carpathian lithospheric fragment (microplate) from the Eastern Alpine area, accompanied by its oblique collision with the North European platform edge in the latest Early Miocene and initial Middle Miocene (Karpatian to Badenian); initial rifting of the Pannonian backarc basin occurred in a transtensional to extensional regime. (3) The extensional regime is related to 'B-type' subduction at the Carpathian orogenic front, accompanied by the synrift stage of the Pannonian backarc basin system during the Middle Miocene (Badenian to Sarmatian). (4) Post-rift extension of the Pannonian backarc basin system occurred due to thermal subsidence and the onset of isostatic uplift of the orogen during the Late Miocene (Pannonian to Pontian). (5) A Pliocene transtensional tectonic regime controlled isostatic uplift and tectonic inversion of the Western Carpathian basins; this was accompanied by the onset of a transpressional tectonic regime in the Pannonian backarc basin system.

Danube Basin (M.K.)

The Danube Basin represents the NW part of the Pannonian backarc basin system (Fig. 17.16). In Slovakia it is geographically termed the Danube Lowland while in Hungary it is referred to as the Little Hungarian Plain. It is situated between the Eastern Alps, the Western Carpathians and the Transdanubian Range in Hungary.

Tectonic setting

The Danube Basin is *c.* 240 km long and 100 km wide and strikes roughly NE-SW. The western border comprises units of the Central Eastern Alps, the Leitha, Hundsheim and Malé Karpaty mountains. The northern margin is represented by the Považský Inovec and Tribeč mountains belonging to the Central Western Carpathians. The Burda Mountains form the margin in the NE while the Hungarian Transdanubian Range Mountains represent the SE border of the basin. The pre-Cenozoic basement is built up by the Austro-Alpine and Slovako-Carpathian units in the western, northern and central part of the basin; the basement of the SE margin comprises units of the Transdanubicum (Fusán *et al.*1987; Fülöp *et al.* 1987).

The basin is divided into several depocentres. Along the northern margin these are from west to east: the Blatná, Rišňovce and Komjatice depressions (separated by the Považský Inovec and the Tribeč mountains). In the NE part, the Želiezovce Sub-basin is located between the Levice Horst and the Transdanubian Range Mountains (Lankreijer *et al.* 1995). The southern, Hungarian part of the basin is separated by the NNE–SSW trending Mihályi High into major sub-basins which parallel the Repce Fault in the west and the Raba Fault in the east. The deepest part of the Danube Basin is the Gabčíkovo Sub-basin (Vass *et al.* 1990) with a maximum thickness of Neogene sediments of >8500 m (Kilényi & Šefara 1989; Hrušecký *et al.* 1993, 1996).

The northern Vienna Basin and the northern Danube Basin shared a common tectonic evolution which is characterized by an Early Miocene palaeostress field with a NW–SE orientated principle compression. (P. Kováč & Hók 1993; P. Kováč *et al.* 1994). During this time, a wrench fault elongate basin developed in the northern part of the present Danube Basin.

An initial phase of rifting commenced during the late Early Miocene (Karpatian) following the extrusion of the Western Carpathians from the East Alpine domain (Csontos *et al.* 1992). In the northern Danube Basin pull-apart depocentres opened in the Blatná and the central Gabčíkovo depressions/sub-basins (Hrušecký *et al.* 1996; Kováč *et al.* 1999). In the southern part of the basin, north–south orientated normal faults were activated (Nemčok *et al.* 1989; Csontos *et al.* 1991; Tari *et al.* 1992; Royden 1993*a*, *b*; Hrušecký *et al.* 1996).

During the Middle Miocene (Badenian) the palaeostress field changed and NW–SE orientated extension prevailed (Csontos *et al.* 1991; Nemčok *et al.* 1989; Nemčok 1993; Vass *et al.* 1993*a*). The Badenian synrift stage of basin development was characterized by structural unroofing of the deeply buried basement units, especially along the western margin (Rechnitz-Sopron area). During the Late Badenian and Sarmatian, north–south, NNE–SSW and NE–SW normal faults were active in the northern Danube Basin (Peničková & Dvořáková 1985; Vass *et al.* 1993*a*). During the early late Miocene (early Pannonian), tectonic subsidence is documented only in the central and southern parts of the Danube Basin (Lankreijer *et al.* 1995). Predominantly low-angle normal faults were activated (Pogácsás *et al.* 1996). During the late Pannonian and Pontian, thermal post-rift subsidence commenced (Becker 1993; Horváth 1993) and was followed by Pliocene basin inversion, which was associated with minor compression from the SW during the Late Miocene and Pliocene (Horváth & Cloetingh 1996). Despite the Pliocene tectonic inversion, in some depocentres subsidence still occurred. Subsidence in the Danube Basin centre, in the Gabčíkovo Sub-basin, is interpreted as being related to secondary 'sag basins', which are not superimposed on the older Miocene basin depocentres (Wernicke 1985).

Sedimentary and stratigraphic development

The Neogene sedimentation in the area of the present Danube Basin began during the Eggenburgian transgression. The sea flooded the northern margin of the Western Carpathians and penetrated from the Alpine and Carpathian foredeep into the Vienna Basin, the Dobrá Voda, Vad'ovce and Blatná sub-basins, the Váh river valley, the Bánovce Basin and the upper Nitra Basin. A connection with the North Hungary–South Slovakia and East Slovakia basins can be documented (Kováč *et al.* 1998). Coarse clastic littoral to shallow-marine deposits of the Dobrá Voda Conglomerate (Kovác *et al.* 1991) in the Blatná Sub-basin and the Klačno Conglomerate in the Bánovce Basin represent the basal part of the Eggenburgian Čausa Formation. Up-section, the 500 m thick Čausa Formation consists of calcareous and sandy clays and silts with some tuff layers deposited in a deeper neritic marine environment (Čechovič 1959; Brestenská 1980; Kováč *et al.* 1999; Hók *et al.* 1995).

Ottnangian and Karpatian terrestrial, fluvial and lacustrine sediments were deposited along the western margin of the Danube Basin and document the initial phase of rifting. These sediments are known from the Sopron area in Hungary where, on the slopes of the Eastern Alps, the coal-bearing Brennberg Formation. (Csaszár 1997) was deposited during the Ottnangian and was overlain by the fluvial to limnic gravels of the Ligeterdö Formation (Csaszár 1997). Ottnangian and Karpatian marine deposits are known from the Dobrá Voda and the Blatná sub-basins in the north and from the Transdanubian Range Mountains. along the southern margin of the Danube Basin. Calcareous siltstones of the Ottnangian to Lower Karpatian Planinka Formation were deposited in the Dobrá Voda Sub-basin (Kováč *et al.* 1992), and the Bánovce Formation represents this facies in the Bánovce Basin (Vass in Keith *et al.* 1994). In this latter basin, the Ottnangian part of the sedimentary succession is *c.* 300 m thick while the Karpatian part is *c.* 250 m thick. The lower part was deposited in a brackish environment, with lowered oxygen content, and the upper part in a marine, shallow to deep neritic environment. The marine pelites ('Schlier') are characterized by the appearance of foraminifer associations containing *Uvigerina graciliformis*.

At the Early–Middle Miocene boundary, rifting in the Danube Basin was accompanied by calc-alkaline volcanic activity. The stratovolcanoes are buried below the Middle–Upper Badenian sedimentary fill in the northern and central parts of the present-day basin and document crustal extension during basin formation (Hrušecký *et al.* 1996). From the Hungarian part of the Danube Basin younger Badenian, Sarmatian to early Pannonian trachyte volcanism is recorded in the Pástori Formation (Császar 1997).

Karpatian deposition commenced with terrestrial sediments in the central part of the basin (*c.* 500 m thick near Györ). The Bajtava Formation, at the eastern basin margin, contains Lower Badenian marginal transgressive conglomerates, sandstones and volcaniclastics, overlain by calcareous clays, siltstones and rare sandstones deposited in a neritic environment (Kováč *et al.* 1999). In the NW part of the basin calcareous clays and siltstones of the 3000 m thick Middle Badenian Špačince Formation were deposited. In the Blatné Sub-basin the delta-front sands of the Madunice Formation indicate shallowing in the latest Middle and Late Badenian (Adam & Dlabač 1969). The Upper Badenian part of Pozba Formation overlies the Špačince Formation. It is up to 2000 m thick. This consists of calcareous clays, siltstones and sandstones with volcaniclastics; in marginal areas algal limestones developed. In the southern (Hungarian) part of the basin, the Baden Clay Formation was deposited (Csaszár 1997). This consists of clays and marls of open-marine facies, with a rich thin-shelled mollusc fauna as well as foraminifers. Its maximum thickness is *c.* 1000 m. In the southern part of the basin coralline algal limestones of the Rákos Formation and marls of the Szilágy Formation were deposited; these have a combined thickness of up to 100 m. At the NW basin margin freshwater deposits with coal-bearing clays and lignite seams of Late Badenian age are also known (Vass *et al.* 1990).

The Badenian deposits are discordantly overlain by the Sarmatian Vráble Formation (Harčár *et al.* 1988). This offshore facies comprises calcareous clays, siltstones and sandstones up to 600 m thick (Adam & Dlabač 1969). In the nearshore areas conglomerates, sandstones, limestones, local lignites and tuffs were deposited. The maximum thickness (1300 m) of the Sarmatian strata is documented from the Rišňovce Subbasin (Biela 1978). In the southern, Hungarian part of the basin, the equivalent of the Sarmatian offshore facies is the Kozárd Formation with a maximum thickness of 150 m. Marginal development is here represented by the Tinnye Formation (100 m) with frequent occurrences of mollusc-bearing calcareous sands and sandstones (Császár 1997).

Pannonian and Pontian-age sediments of the Danube Basin were deposited along the northern margin of Lake Pannon. The northernmost part of the basin was shallow, while the central and southern parts were hundreds of metres deep (Kovač *et al.* 1999). The basin was gradually filled by deltaic deposits entering the basin from the NNW, transporting clastic material from the uplifting Alpine-Carpathian orogenic belt. The Upper Miocene and Pliocene deposits, containing clays, siltstones and sandstones, are up to 3500 m thick in the Gabčíkovo Depression/Sub-basin (Adam & Dlabač 1969; Vass *et al.* 1990).

In the Hungarian part of the basin, Lower Pannonian deposition began with the open-water marls of the Endröd Formation (Vass 2002) indicating a water depth of up to 800 m. The overlying Middle to Upper Pannonian Újfalu Formation was deposited in delta-front to delta-plain settings. The Pontian to Lower Pliocene Zagyva Formation represents alluvial plain, fluvial to lacustrine environments, and comprises *c.* 1000 m of sands, silts and coal-bearing clays (Csaszár 1997). In the Slovak part of the Danube Basin the Lower and Middle Pannonian sediments are represented by the Ivánka Formation (Harčár *et al.* 1988), comprising calcareous clays, siltstones and sandstones. The Upper Pannonian/Pontian, *c.* 100 m thick Beladice Formation (Harčár *et al.* 1988) consists of calcareous clays and siltstones, with coal-bearing clays and lignite seams. The Lower Pliocene Volkovce Formation contains deltaic deposits of the palaeo-Hron river in Komjatice and the Gabčíkovo sub-basins. The Upper Pliocene Kolárovo Formation (Dlabač 1960) represents the palaeo-Váh river deposits in the Blatná Sub-basin. Quaternary deposits of the Danube Basin are represented by loess deposition (up to 150 m) and fluvial and alluvial deposits.

East Slovak Basin (M.K.)

The East Slovak Basin (ESB) is situated between the Western and Eastern Carpathians (Fig. 17.16). The western border of the basin is formed by the Tatric and Veporic units of the Central Western Carpathians. The NE and eastern margins consist of units of the Pieniny Klippen Belt, and the Humenné Mountains. Towards the south the Neogene successions pass into the Hungarian Hernád Basin. The SW and SE margins are formed by the Slovak-Hungarian Zemplín area and the Ukrainian Seredné area (Rudinec 1978, 1989).

Tectonic setting

The ESB represents the NW part of the Transcarpathian Basin which covers parts of Slovakia, Ukraine and Romania, reaching 220 km in a NW–SE direction. The ESB is generally up to 9000 m deep, although the Ukrainian depocentres were only 2000–3000 m deep (Rudinec 1989). Various structural and geological units belonging to the Western and Eastern Carpathians form the basement of the ESB (Sviridenko 1976; Rudinec *et al.* 1981; Rudinec 1984; Vass *et al.* 1988; Soták *et al.* 1990 1993, 1994). The NE margin is represented by the Mesozoic and Palaeogene units of the Pienniny Klippen Belt which separates the units of the Outer Carpathians Flysch Belt from the Mesozoic complexes of the Humenné Mountains (Mahel 1986). The NW part of the basin basement comprises the Central Western Carpathian Mesozoic and Palaeozoic units of the Čierna Hora Complex, while the SW part consists of the Zemplín and Ptrukša complexes (Rudinec 1984; Vass *et al.* 1988; Soták *et al.* 1993).

The tectonic development of the ESB reflects changes both in terms of its geotectonic position, as well as due to changes in the palaeostress field orientation (Kováč *et al.* 1995). The Lower Miocene (Eggenburgian) sediments were deposited in a forearc basin position on the margin of the moving Central Western Carpathians. The palaeostress field reflects NE–SW to NNE–

SSW orientated main compression (Nemčok 1993). Compressive tectonics led to the disintegration of the Early Miocene basin and to the development of backthrusts in the Pieniny Klippen Belt (Seneš 1956; Roth 1980; Plašienka *et al.* 1998).

During the late Early Miocene (Karpatian), a palaeostress field with a north–south orientated main compression prevailed. NW–SE orientated normal and later right-lateral strike-slip faults were formed. During Karpatian and Early Badenian initial rifting, pull-apart depocentres opened in the central part of the ESB. Tectonically controlled subsidence shows a cyclic character, with periods of high sedimentation followed by periods of basin isolation and shallow-water evaporite deposition. Crustal extension was associated with updoming of a lithospheric mantle diapir leading to structural unroofing of the Iňačovo-Kritschevo Peninic Unit (Soták *et al.* 1993).

The Upper Badenian and Sarmatian tectonic regime of the ESB can be characterized as the development of an interarc/backarc type basin. The change from transtension, during initial rifting, to pure extension, during the synrift phase, resulted in tectonically controlled subsidence and a high sedimentation rate during the Early Sarmatian (Pereszlényi *et al.* 1991). The main depocentres shifted to the SE. The basin development coincided with increased volcanic activity at this time (Lexa *et al.* 1993).

During the Upper Miocene the thermal, post-rift phase began and the ESB became part of the Pannonian Basin System (Horváth *et al.* 1988; Mattick *et al.* 1988). Subsidence was controlled by north–south to NW–SE extension. The Pliocene was characterized by tectonic inversion and a palaeostress field with NE–SW orientated compression which led to folding of the Pannonian deposits (Kováč *et al.* 1995) and uplift of the youngest sediments in the East Slovak Lowlands (Mořkovský & Lukášová 1986, 1991).

Sedimentary and stratigraphic development

The Lower Miocene deposits are situated mainly in the NW part of the basin, Middle Miocene deposits fill its central part, and the late Middle Miocene and the Upper Miocene depocentres are in the SE part of the ESB (Janáček 1969; Rudinec 1978, 1989). The following summary is based mainly on Vass & Čvercko (1985), Zlinská (1992) and Rudinec (1978, 1989) for the Slovakian part and on Vialov (1986) and Andreyeva-Grigorovich *et al.* (1997) for the Ukraine.

The Eggenburgian open-marine sediments belong to the Prešov Formation and were deposited in a forearc basin environment. The succession, which is up to 1000 m thick, consists of clays, siltstones, sandstones and conglomerates. The end of Eggenburgian deposition is marked by the Čelovce Member where marine siltstones and clays grade into lagoonal and freshwater deltaic deposits consisting of conglomerates, sandstones, siltstones and lignites. Ottnangian deposits have not been recorded in the ESB. In the Ukrainian part of the Transcarpathian Basin the Burkalo Formation represents similar early Miocene deposits. In the ESB, 1600 m thick Karpatian-age sediments comprise the Teriakovce, Sol'ná baňa and Kladzany formations. The Teriakovce Formation is up to 500 m thick and grades from conglomerates and sandstones into siltstones and clays deposited in a deep-marine, neritic to shallow bathyal environment. The overlying Sol'ná baňa Formation (400 m) is represented by evaporite deposition which took place in a shallow-water environment as a result of basin isolation. Renewed latest Early Miocene tectonic subsidence is reflected by the deposition of clastic material via turbidity currents of the Kladzany Formation. The succession contains clays and siltstones with sandstone intercalations and attains a maximum thickness of 1000 m. In the Ukrainian part of the Transcarpathian Basin, the Tereshul Formation is the equivalent of the Karpatian deposits in the East Slovak Basin.

The Lower Badenian sediments in the eastern and central part of the ESB are represented by volcaniclastic deposits of the Nižný Hrabovec Formation. The formation is up to 600 m thick and consists of marine clays, siltstones with sandstone intercalation and rhyolite tuffs. In the western part of the basin the Lower Badenian is represented by the Mirkovce Formation with marked redeposition of Karpatian-age microfaunas in its basal parts. In the Ukrainian part of the Transcarpathian Basin the Lower Badenian is represented by the Novoselytsa Formation.

During the Middle Badenian, clays, siltstones, sandstones and rare tuffs of the 600 m thick Vranov Formation were deposited in the central part of the basin. The sands were derived from the uplifting Outer Carpathians in the NE. The sedimentary environment gradually changed from a deep-marine to a shallow-marine one and the end of deposition is represented by the lagoonal evaporites of the Zbudza Formation. In the Ukrainian part of the Transcarpathian Basin these evaporites are united in the Tereblya Formation and the lower part of the Solotvino Formation.

The Upper Badenian transgression reached the ESB from the south, from the Pannonian region. The deep-water pelitic facies in the basin centre pass into coarse clastic deltaic deposits prograding from the NW margin of the basin. The lower part of the Upper Badenian succession is represented by the Lastomír Formation, attaining a thickness of up to 2000 m in the SE part of the basin. The formation consists of calcareous clays with siltstone intercalations deposited in delta-slope and prodelta environments. The upper part of the succession is represented by the Kolčovo Formation (up to 1700 m thick), whose deposition extended up into early Sarmatian. In the Ukrainian part of the Transcarpathian Basin, the Upper Badenian is represented by the upper part of the Solotvino, the Teresva and the Baskhev formations.

In the Sarmatian, the depocentre of the ESB widened towards the SE in the region of the Vihorlat Mountains and towards the SW in the Košice Sub-basin. Fan deltas and braided river system deltas prevailed (Janočko 1993). The Lower to Middle Sarmatian Stretava Formation (1600 m thick) is a monotonous unit of calcareous clays with rhyolite tuff intercalations which was deposited in a deltaic environment. The delta prograded from the NW towards the SE. Its marginal facies are represented by the Košice Member (Kaličiak 1991). The overlying Upper Sarmatian Ptrukša Formation, consisting of calcareous sandstones and tuffs, is up to 300 m thick. In the western part of the basin freshwater systems prevailed, as indicated by the lignite-bearing Middle to Upper Sarmatian Kochanovce Formation. In the Ukrainian part of the Transcarpathian Basin the Sarmatian is represented by the Dorobrotiv, Lukiv and Almash formations.

The Upper Miocene of the ESB comprises up to 600 m of brackish to freshwater deposits. The Pannonian Sečovce Formation consists of calcareous clays, coal-bearing clays, coal seams and tuffs (Albínov tuff; Janáček 1969) whilst the Pontian Senné Formation was deposited initially in fluvial (Pozdišovce Member) and later in lacustrine, environments. Up to 200 m of clays, sands, gravels (with andesite pebbles) and tuffs comprise the Pliocene Čečehov Formation.

In the Ukrainian part of the Transcarpathian Basin the Pannonian Iza and the Pontian Koshelevo formations attain a joint thickness of up to 400 m and represent the Upper Miocene succession. The Pliocene is represented by the Dacian Ilnitsa Formation, up to 500 m thick, and the Pleistocene by the clays, sandstones and gravels of the Chop Formation (600 m).

Volcanism
The development of the East Slovak Basin was associated with voluminous volcanic activity related to backarc extension and subduction beneath the front of the Carpathian arc. Extension-related acid rhyolite volcanism is known from the Lower and Middle Miocene successions. The Middle to Upper Miocene is predominantly calc-alkaline andesite volcanism related to subduction processes. The Late Badenian and Early Sarmatian volcanic activity is of arc type (Vass *et al.* 1988; Kaličiak & Pospíšil 1990; Szabó *et al.* 1992; Lexa *et al.* 1993). During the Sarmatian, the basin evolved into an interarc basin and andesite volcanic activity culminated (Vass *et al.* 1988). The volcanic chains can be traced along the eastern margin of the basin towards the Gutin Mountains and along its southern border along a now-buried area between Zemplín and Beregovo (Slávik 1968).

North Hungarian–South Slovak basins (M.K., D.V., L.S., A.N.)

The North Hungarian–South Slovak basin is bordered to the north by the Western Carpathians (Fig. 17.16). The western margin is represented by the Transdanubian Range Mountains, while the eastern margin consists of units of the Bükk Mountains. In the south, it is bordered by the Mid-Hungarian region which forms the boundary between the Alcapa and the Tisza-Dacia microplates.

Palaeogeography and tectonic setting
The pre-Cenozoic basement is formed in the north by units of the Central Western Carpathians (Veporicum, Gemericum, Silicicum, Meliaticum), in the west by units of the Transdanubicum and in the east by units of the Igal–Bükk Zone. The Slovak part is geologically not an individual basin but represents the northern margin of three basin complexes which overlie one another: the Buda Basin (= North Hungary Palaeogene Basin), the Fil'akovo-Pétervására Basin and the Novohrad-Nógrad Basin. A detailed overview including all relevant literature used in the present text was published by Kováč *et al.* (2002). The entire region rotated 50° counterclockwise during the Early to Middle Miocene (Márton *et al.* 1996).

The Buda Basin (North Hungary Palaeogene Basin) began to form following a long period of emersion of the Transdanubian Range Mountains (Transdanubicum units), as indicated by lateritic weathering and the creation of karst bauxite deposits (Báldi & Báldi-Béke 1985). Its evolution was terminated by the tectonic extrusion, post-late Oligocene, of the Alcapa Microplate from the Alpine-Dinaride domain towards the Carpathian-Pannonian realm. This led to the disintegration of the Palaeogene basin by right-lateral displacement along the Mid-Hungarian region. Two segments were created, Buda and Slovenia (Ljubljana), which are at present 300 km distant from one another (Nagymarosy 1990; Csontos *et al.* 1992).

In the Early Miocene (Eggenburgian), the Fil'akovo-Pétervására Basin was formed. Its extent was less then that of the Buda Basin and it lacked the southern connection to the open sea but opened towards the basins of the Outer and Inner Carpathians (Sztanó 1994; Halásová *et al.* 1996). At the end of the Eggenburgian, extensive felsic volcanism, related to asthenospheric mantle uplift associated with the uplift of the area (active rifting), occurred, together with a coeval marine regression. Initial backarc rifting resulted in the deposition of the Bukovina Formation which consists of continental deposits with rhyodacite tuffs (Zagyvápálfalva Formation (Császár 1997), rhyodacite tuffs in Gyulakeszi, Hungary).

During the late Early Miocene, the development of the backarc basin continued (Vass *et al.* 1993*a*) and the Novohrad-Nógrád Basin formed. Gradual marine transgressions from the south in the Ottnangian are documented by paralic sedimentation in Boršod (northern Hungary) and by marine ingressions in southern Slovakia. Basin subsidence reached a maximum during the Karpatian, followed by rapid regression. The area was uplifted and erosion was marked by latest Karpatian times. The final transgression extended into the area of south Slovakia, in the early Middle Miocene (Badenian). From Middle Badenian times onward the region was emergent and subjected to intense weathering.

Sedimentary and stratigraphic development
This section is largely based on the work of Báldi & Báldi-Beke (1985), Báldi (1986), Báldi-Beke & Báldi (1991), Vass (2002), Vass *et al.* (1979, 1983, 1987), Vass & Elečko (1982, 1992), and Hámor (1988).

Zala Basin. The southwesternmost Palaeogene deposits of the Transdanubian Range area occur in the Zala Basin (Fig. 17.17). The oldest member of the succession is the littoral to neritic Upper Lutetian Szőc Limestone (180 m). This formation is overlain up to 600 m of calcareous and sandy marls of the pelagic, epibathyal Padrag Marl, reflecting the period of maximum deepening during the Late Lutetian and Early Priabonian. In the upper part of the formation, andesitic tuffs of the Zalatárnok-Zalaszentmihály volcanic centre are interbedded (Szentmihály Andesite). Near the basin centre, a stratovolcanic complex more than 1000 m thick, has been penetrated. The Eocene deposits are covered by up to 2000 m of Neogene sediments (Kőrössy 1988).

South Bakony Mountains. In the southern Bakony the Palaeogene sequences begin with local bauxite deposits (Gánt Formation). Bauxite occurrences can be traced along the northern and southern margins of the Bakony Mountains. The bauxite is overlain by the Lower Lutetian Darvastó Formation (*c.* 40 m) comprising terrestrial clays and quartzitic conglomerates in its lower part. Its upper part consists of neritic limestones and marls with a rich *Alveolina* and *Nummulites* fauna (Kecskeméti 1998). The Darvastó Formation is conformably overlain by the Upper Lutetian to Lower Bartonian Szőc Limestone (*c.* 100 m). This biogenic limestone contains various shallow-marine organisms such as corallinaceans, bryozoans, echinoids and larger foraminifera. The Szőc Formation passes upwards into the pelagic, bathyal Padrag Marl (*c.* 250 m; Bartonian to Priabonian). Pebbly mudstones, sandy turbidites and tuffitic intercalations occur in its upper part. The top of the eroded Eocene is unconformably covered by the terrestrial and fluviatile gravels and clays of the Upper Oligocene Csatka Formation (*c.* 120 m).

North Bakony Mountains. In this region Eocene deposition in an archipelago landscape commenced during the latest Lutetian somewhat later than in the South Bakony Mountains. Basal clastics and redeposited bauxite are covered by the Dorog Coal Formation (*c.* 50 m) comprising paralic marls with diverse brackish and marine mollusc fauna. The coal is conformably overlain by neritic marls and by the epibathyal Padrag Marl (*c.* 200 m). Above a hiatus, the Upper Oligocene non-marine Csatka Formation follows (*c.* 800 m); this part of the succession is similar to that of the South Bakony Mountains. Terrestrial and freshwater deposits alternate; thin coal seams occur in the basal parts and fluviatile gravels and claystones are characteristic in the upper part.

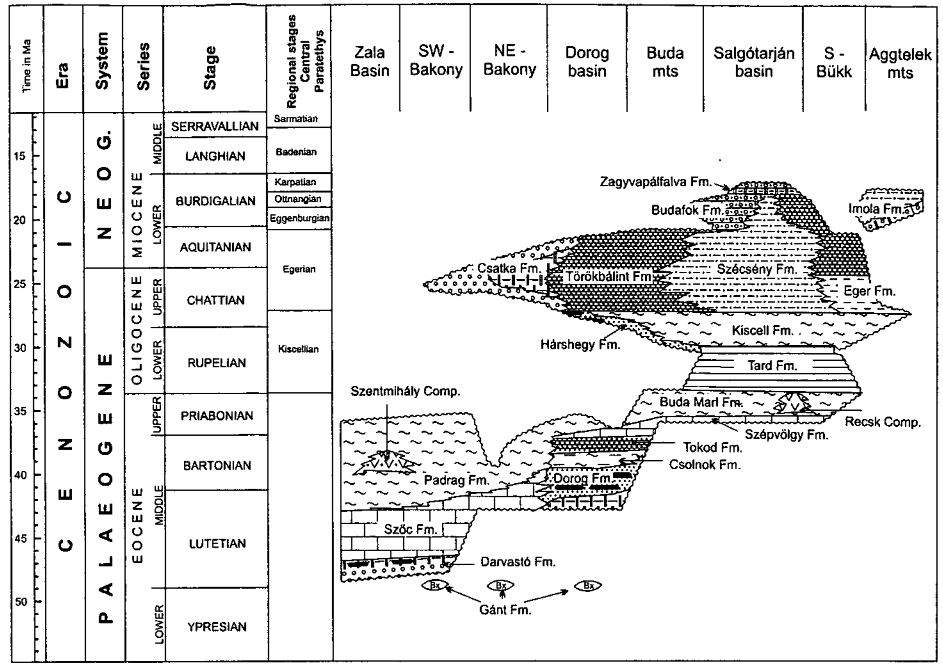


Fig. 17.17. Palaeogene and lower Neogene lithostratigraphy of the Hungarian basins. See Figure 17.18 for legend.

Buda Basin. The Lutetian transgression gradually flooded the Buda Basin (Fig. 17.17; North Hungary Palaeogene Basin) forming lagoons in which lignite deposition (Obid Member, Dorog Formation) occurred. Shallow-water carbonate sedimentation and deposition on the outer shelf is represented by the Priabonian limestones of the Szépvölgy Formation (Kázmér 1985) and the overlying marls of the Buda Formation. At the end of the Eocene, a regression in the Transdanubian region led to a shift of the depocentre eastwards into the region of the Buda Hills.

During the Oligocene, the Buda Basin rapidly deepened as indicated by the deposits of the Číž Formation in Slovakia and the equivalent Hárshegy and Kiscell formations in Hungary. The oldest Oligocene sediments are freshwater and fluvial deposits of the Skálnica Member (Kiscellian age) deposited in the Rimava and Lučenec sub-basins. These are overlain by the littoral deposits of the Hostišovce Member consisting of clays, silts and sandstones with thin coal seams. Ongoing transgression led to the formation of the sponge/bioclastic and intraclast limestones of the Batka Limestone Member deposited along the slopes of barrier islands. The subsequent isolation of the Central Paratethys is manifested by the euxinic facies of the Tard Clay Formation cropping out in the Budapest area. The development of new depositional centres in the Oligocene was accompanied by a transgression during the Kiscellian as marked by the clays of the Kiscell Formation (800 m) in Hungary and the upper Číž Formation in Slovakia. In the Late Oligocene (Egerian) the evolution of the Buda Basin continued with the deposition of the Lučenec Formation. Its lower part consists of up to 150 m of transgressive breccias, conglomerates and sandstones with a shallow-water marine fauna (i.e. Panica Member). Stratigraphic equivalents include the Hungarian Törökbálint Formation and the Slovak Budikovany Formation. The Egerian offshore facies is represented by shaly calcareous siltstones up to 1000 m thick (Szécsény Member = Lower part of Szécsény Schlier Formation). Limestones and conglomerates of the Bretka Member, containing *Miogypsina gunteri*, represent the nearshore facies in the upper part of the Lučenec Formation. The Egerian succession terminates with the regressive delta sediments of the 180 m thick Slovak Opatovce Member and the lignite-bearing, brackish, marshy-fluvial Becske Formation.

Fil'akovo-Péterváśára Basin. During the Eggenburgian, the short-lived Fil'akovo-Péterváśára Basin evolved. The Fil'akovo Formation concordantly overlies the Lučenc Formation in the Cerová vrchovina Mountains. Its Hungarian equivalent is the Pétérvásara Formation. The 500 m thick formation was deposited in shelf environments, often showing a strong tidal influence (Sztanó 1994). The Slovak Čakanovce Member and the upper part of the Hungarian Szécsény Formation (Császár 1997) represent the offshore facies. During the late Eggenburgian the 200 m thick, mainly fluvial deposits of the Bukovinka (Slovakia) and Zagyvapálfa formations (Hungary) were deposited. These comprise sandstones, clays and rhyodacite tuffs (in Hungary Gyulakészi tuff) with abundant remnants of warm subtropical to tropical flora. The Hungarian Zagyvapálfa formation contains the famous footprint sandstone at Ipolytarnóc with very well preserved traces of mammals and birds, and an enormous fossil trunk of *Pinus* sp. (Kordos 1985; Hably 1985).

Novohrad-Nógrád Basin. During the middle Early Miocene (Ottnangian), the Novohrad-Nógrád Basin began to form. In Slovakia the lacustrine Salgótarján Formation is up to 250 m thick. In its lower part it consists of 30–80 m of locally cross-bedded sands and clays with three characteristic coal seams (Pôtor Member). The overlying Plachtince Member comprises clays with rare coals and tuffs. Carbonate clays include foraminifera and nannoflora of zones NN3 and NN4 indicating marine ingressions during the Ottnangian. The Salgótarján Formation gradually passes into the Karpatian Modrý Kameň Formation, of which the Medokýš Member represents the basal part of the succession. It consists of *c.* 60 m of fine-grained laminated sandstones and siltstones, locally with convolute bedding (Vass & Beláček 1997) and tempestites with mixed marine and endemic-brackish fauna. The overlying Krtíš Member (40 m) represents littoral sediments. Deposition culminated with the 300 m thick Sečiance Member containing a rich bathyal marine fauna. Its equivalents in north Hungary include the Egyházasgerge and the Fót formations and the Garáb Schlier Formation (Csaszár 1997). At the end of the Early Miocene a major regression occurred in the entire Novohrad-Nógrád Basin leading to erosion.

A subsequent transgression in the early Middle Miocene (Badenian) led to the deposition of the Príbelce Member (Vass *et al.* 1979), which is an equivalent of the Hungarian Baden Clay Formation (Csaszár 1997). This transgression was accompanied by intense volcanic activity, represented by andesite volcaniclastics in the Krupinská planina Mountains. The Hrušovo Member comprises tuffaceous deposits with marine fauna and nannoflora of the Early Badenian. During the Middle Miocene the Novohrad-Nógrád Basin became continental. After a phase of emersion and denudation during the Middle and Late Miocene, some subsidence is recorded in the latest Miocene or Pliocene (Pontian). Fluvial and lacustrine deposits of the Poltár Formation (gravels, sands and kaolin clays) were deposited in the Lučenec, Rimava, Rožňava and Turňa sub-basins. Coeval basaltic volcanism (Podrečiany Formation) led to the formation of maars with laminated lacustrine deposits rich in organic matter (e.g. alginates at Pinciná; diatomic clays at Jelšovec) (Vass *et al.* 1998). Dacian to lowermost Quaternary volcanism is indicated by the basaltic Cerová Formation which includes various volcanic structures (cones, lava flows, lava plateaus, maars). As a consequence of the uplift of the Cerová vrchovina Mountains and due to the selective erosion of the softer deposits of the Fil'akovo Formation, the basaltic flows, originally filling palaeovalleys, now form the relief (relief inversion).

Continental development. South of the Mecsek Mountains more than 500 m of clastics of the Eocene Szentlőrinc Formation represent an isolated Palaeogene continental depositional environment in the Szigetvár area (Császár *et al.* 1990).

Pannonian Basins System (A.N.)

The Pannonian Basins System (PBS) (Fig. 17.16) is a complex of several (sub-)basins that formed during the Neogene. Geographically, they form the Little Hungarian Plain and the Great Hungarian Plain. The PBS is surrounded by several Neogene basins (e.g. the Vienna and Styrian basins in the west, the Bánovce, Nitra and Transcarpathian basins in the north, the Transylvanian Basin in the east, and the Morava, Sava and Tuzla basins in the south). All of these basins formed in the Early–Middle Miocene and contain several thousand metres of sediment. Generally, a gradual shift in basin formation can be observed from the NW to the SE (see Csontos *et al.* 1992). Geologically, the major basins described below are the most important.

The Little Hungarian Plain Basin

The Little Hungarian Plain Basin (LHPB) (Figs. 17.16 & 17.18) extends to the north of the Transdanubian Range and south of the Kőszeg, Sopron, Little Carpathian and Inovec mountains. In its initial, early Middle Miocene, stage of formation, the LHPB was not connected to the Vienna and Zala basins. During the Middle Miocene, it became connected with the Zala Basin but remained separated from the Vienna Basin by the Mihályi Ridge (Tanács & Rálisch 1990). The Tatricum in Slovakia (Fusán *et al.* 1987), the Bakony Unit in Hungary and the Alpine nappe systems form the basement of the LHPB. This basement is structured by tectonic highs (e.g. the Mosonszentjános and Mihályi highs) and deep sub-basins (e.g. the Csapod, Pásztori and Szigetköz sub-basins).

Subsidence of the LHPB began during the late Early Miocene. Initial deposits include terrestrial and freshwater clays, breccias and conglomerates (up to 400 m) deposited along the central axis of the basin, in a 25 to 30 km wide belt. The first Middle Miocene marine sediments (Badenian) are the sandstones and limestones deposited at the basin margins. Pelites with sandstone intercalations appear in the more basinal parts. These deposits are up to 300 m thick and include the Baden Clay Formation and the Rákos Limestone. Volcanic tuffs are frequent in the marine sequences and thick stratovolcanic successions occur to the NE in Slovakia. By the Badenian the entire LHPB had been flooded by the sea. Only a few NE–SW trending ridges formed islands. The thickest Badenian units are located in the NE of the basin. The Badenian–Sarmatian transition appears to be continuous in the deepest part of the LHPB. The Sarmatian successions, up to 450 m thick, consist of clayey marls, claystones, sandstones and tuffitic sandstones (Tinnye and Kozár Formations). Upper Miocene (Pannonian) successions of the Peremarton supergroup (up to 1200 m thick) overlie the Sarmatian. The lower part of this supergroup is mainly clayey (e.g. Endrőd Marl = Belezna Calcareous Marl, Lenti Marl, Nagylengyel Marl) while the Újfalu (= Tőfej) Sandstone and Algyő (= Dráva) Formation occur in the upper part. The associated Late Pannonian transgression of Lake Pannon led to flooding of the entire area (Tanács & Rálisch 1990). The thickest units (*c.* 2100 m) of the LHPB were deposited during the latest Miocene and Pliocene. These include the Dunántúl Supergroup which is composed of the Újfalu Sandstone, the Rábaköz and the Hanság Red-bed formations (Jámbor 1989).

The **Hungarian Central Range** (HCR) represents a complex of small sub-basins extending in age from the Ottnangian to the Middle Badenian. During the Late Badenian it formed a slowly, but continuously, rising SW–NE striking ridge. The uplift is still ongoing. This ridge includes the Bakony, Vértes, Gerecse, Buda, Cserhát, Bükk and Aggtelek-Rudabánya mountains. Between the uplifting blocks, small sub-basins developed, while the later Neogene sedimentation was confined to the margins of the

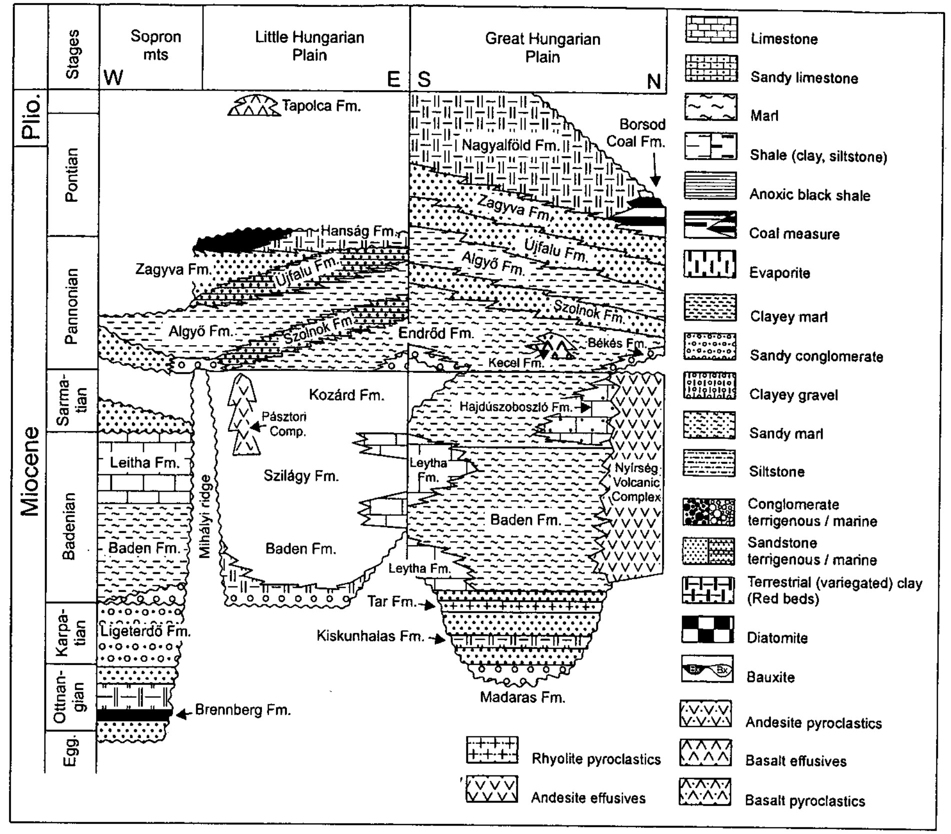


Fig. 17.18. Lithostratigraphy of the Sopron Mountains in Hungary, the Little Hungarian Plain and the Great Hungarian Plain.

uplifting mountains. The HCR comprises the Nagygörbő and the Varpalota basins, and the Budapest area.

The Nagygörbő Basin developed during the Early Miocene along the NW margin of the Bakony Mountains (Jámbor 1980; Hámor & Jámbor in Steininger *et al.* 1985). The Ottnangian deposition commenced with terrestrial clays and conglomerates, including the Lower Rhyolite Tuff (130 m). This member is overlain by Karpatian clayey marls, siltstones and sandstones (150 m) yielding marine fossils in the upper part. The Badenian sandy marls, limestones, sandstones and conglomerates are up to 420 m thick, unconformably followed by 40 m of Sarmatian conglomerates and limestones. The Pannonian is represented by 120 m of siltstones and gravels, while the Pontian and Pliocene comprise 200 m of sands, clays and lignites.

In the Várpalota Basin (North Bakony) 50 m of terrestrial clays (possibly Eggenburgian?) are overlain by the Ottnangian–Karpatian Bántapuszta Formation (250 m), a marine complex of conglomerates, calcareous gravels, sandy limestones and tuffaceous clays (Kókay 1973; Jámbor 1980). An Upper Badenian unit up to 350 m thick consists of brackish to freshwater clays, coal seams, diatomites and alginites and unconformably overlies this succession. The Sarmatian is characterized by tuffaceous sandstones and claystones. The upper part of the depositional succession of the Várpalota Basin is composed of 60 m of Pannonian clayey marls, 120 m of Pontian sandstones, siltstones and freshwater limestones.

In the Budapest area the Neogene succession begins with 100 m of Eggenburgian marine conglomeratic sands (Budafok Sand). This is the littoral equivalent of the Szécsény and Pétervására formations in northern Hungary. The conglomeratic sands are overlain by Ottnangian and Karpatian gravels, calcareous conglomerates and tuffitic siltstones of the Fót Formation (90 m). In the Middle Badenian non-marine beds and tuffs were deposited, and were subsequently overlain by corallinacean limestones and clays in the Late Badenian. The Sarmatian Sóskút Limestone lies conformably on the Badenian. The total thickness of the Middle Miocene is 200 m in this region. Pannonian clays and Pontian sands (up to 300 m) unconformably overlie the Middle Miocene deposits (compiled from Jámbor *et al.* 1966).

The **Somogy-Mecsek-Kiskunhalas Basin** (SMKB) consists of a series of sub-basins extending from south Transdanubia to the Danube-Tisza interfluve. The sub-basins of the SMKB have a SE–NW strike, and are bounded by two large fault systems: the Mecsekalja Lineament in the south and the Kapos Lineament in the north. The SMKB extends from Transdanubia to the southern part of the Great Hungarian Plain (Tanács & Rálisch 1990). An important feature of the sedimentation in this region is the early onset of subsidence (latest Eggenburgian to earliest Ottnangian) in contrast to the areas of Great Hungarian Plain and the Little Hungarian Plain.

The evolution of the SMKB can be related to three transgressions. Subsidence began in the Early Ottnangian but affected only the western part of the zone (i.e. the Somogy and the West Mecsek region). Terrestrial and fluvial sedimentation was mainly to the west and north of the Mecsek Mountains. No subsidence took place in the Kiskunhalas area during the Ottnangian and Early Karpatian. The next phase began in the Late Karpatian. The character of the sedimentation changed from terrestrial through brackish to marine. This transgression also invaded the Kiskunhalas area. Maximum sediment accumulation was to the north (Mecsek) and in the Kiskunhalas Sub-basin. During the third phase, the transgression covered the area from the SW. During the Pannonian and Pontian the Mecsek area was uplifted and formed an island in Lake Pannon. The area north and south of the Mecsek area subsided markedly during the Late Miocene (1500 to 2000 m thick units). In the southern part of the Kiskunhalas Sub-basin (Felgyő and Kömpöc areas) intense subsidence continued into the Pliocene. The sedimentary development in the SMKB is herein subdivided into four areas as described below.

1. In the West Somogy area of the SMKB the succession begins with the 1100 m thick Szászvár Formation, a series of terrestrial/fluviatile conglomerates, sandstones, clays, clayey marls and clays. It is overlain by the marine sandstones and clayey marls of the Budafa Formation of Karpatian to Early Badenian age (970 m). Up-section follows the 490 m thick Tekeres Schlier, a series of clayey marls and siltstones. A hiatus separates the lower part of the Lower Badenian and the overlying 690 m thick Middle and Late Badenian rocks, represented by the Rákos Limestone and *Turritella-Corbula*-bearing marls (Szilágy Formation). The next hiatus spans the Sarmatian and large parts of the Pannonian. The transgressive Újfalu Sandstone represents only the upper part of the Pannonian. These beds are followed by the Zagyva Formation. The total thickness of the Pannonian is only 330 m. The overlying Pontian comprises 250 m of sands and clays (Somló Formation) and 650 m of the Tihany Formation.

2. In the West Mecsek Mountains (compiled from Forgó *et al.* 1966 and Hámor 1970) the Neogene succession begins with a 500 m thick series of Ottnangian–Karpatian terrestrial/fluvial conglomerates, sandstones and siltstones (Szászvár Formation). The regressional upper part contains coal seams and clays. The Gyulakeszi ('Lower') Rhyolite Tuff is intercalated into the lowermost part of this formation. The second, Karpatian–Lower Badenian succession comprises limnic shales (Komló Member), the brackish and marine Budafa Formation, and the shallow marine Tekeres Schlier. The Middle–Late Badenian is represented by the 130 m thick Rákos Limestone. This littoral/shallow neritic lithofacies may be substituted by the Szilágy Marl (*Turritella-Corbula* beds) of more basinal character. The Sarmatian comprises 20–80 m of littoral limestones. Because of uplift during the Late Miocene and Pliocene, Neogene sedimentation ceased during the Sarmatian in the West Mecsek Mountains.

3. The development in the East Mecsek Mountains differs considerably in terms of the distribution and thicknesses of the formations (Forgó *et al.* 1966; Hámor 1970). The terrestrial/fluviatile Szászvár Formation (380 m) represents the initial period of sedimentation. It consists of clays, conglomerates, sandstones and coal seams in its upper part. The overlying Karpatian fish-shale (370 m) includes conglomerates, thin marine siltstones and an intercalation of the Tar ('Middle') Rhyolite Tuff in its upper part. The Badenian is represented by the corallinacean limestones of the Pécsszabolcs Formation. This is overlain by the paralic coal seams of the Hidas Formation. The Late Badenian transgression deposited the Szilágy Marl in the basinal parts and the Rákos Limestone in nearshore environments. The total thickness of the Badenian is 280 m. The conformably overlying Sarmatian limestones are 150 m thick.

4. The Kiskunhalas Basin is located in the southern part of the Danube-Tisza interfluve. Its depth may be >5000 m. The oldest Neogene sediments belong to the Kiskunhalas Formation, and comprise coarse-grained clastics, which correspond to the Szászvár Formation in the Mecsek Mountains. These beds, up to 800 m thick, are of Early Karpatian age in the deeper sub-basins or younger (Early Badenian?) in the shallower ones. The marine Middle Badenian conformably overlies these non-marine beds. It is represented by conglomerates, breccias and the offshore Szilágy Marl (*c.* 600 m). No Sarmatian deposits are known from this area. The transgressive pelitic and sandy Pannonian Peremarton Supergroup follows (850 m). Effusive lava beds of the Kiskörös-Kecel basalt are intercalated into the Lower Pannonian pelitic beds. The Pontian and Pliocene regressive sucession (*c.* 1800 m thick) is represented by the mainly sandy Dunántúl Supergroup. Clays occur in the upper part of the Pliocene.

The **Mid-Transdanubian Zone** is a deep Neogene basin located between the Balaton and Kapos lineaments in Hungary. The deepest part of the basin may be up to 5000 m. The oldest member of the Neogene comprises a few hundred metres of conglomerates with frequent red-bed intercalations. Its age may be pre-Badenian or Early Badenian. It is overlain by thin marine Badenian deposits and a thick Badenian volcanic complex. The younger part of the Neogene was cored by the Mezőcsokonya-4 borehole which recorded 2000 m of Neogene rocks, but did not drill the entire thickness of the Badenian volcanic complex. This was unconformably overlain by 450 m of the Pannonian Peremarton Supergroup and by 1200 m of the Pontian and Pliocene Dunántúl Supergroup.

Zala and Drava basins

The Zala Basin is located in SW Hungary (Fig. 17.16) and is subdivided into a northern and a southern sub-basin (Körössy 1988; Szentgyörgyi & Juhász 1988). The adjoining Drava Basin is situated in the Drava lowlands, which partly cover SW Hungary and northern Croatia. These two basins are connected both geographically and also in terms of their depositional history. The main difference is the strike of the basins. While the axis of the Zala Basin is SW–NE, the Drava Basin is NW–SE (Körössy 1988, 1989; Tanács & Rálisch 1990). There are also differences in the sediment thicknesses between the basins, with the Neogene succession being somewhat thicker in the Dráva Basin. The Zala Basin formed parallel to the Rába Fault, which

can be considered as a major low-angle fault (Rumpler & Horváth 1988). Seismic evidence indicates that the Budafa area, where the Dráva and Zala basins join, underwent very late compression resulting in folding of the uppermost Miocene/Pliocene strata (Körössy 1988), although the Quaternary units are undeformed. These folded structures can be considered as the west–east axial prolongation of the flat Sava folds which extend from Slovenia into Hungary. The maximum subsidence of the area took place in the Pannonian and Pontian. The thickness of the Pontian (probably including Pliocene parts) is up to 2500 m in the Dráva Basin.

The Zala Basin is subdivided into a northern and a southern sub-basin. The northern Zala sub-basin is characterized by the Transdanubian Range. Here, the Karpatian deposits are often missing, and the Badenian, Sarmatian, Pannonian and Pontian deposits are relatively thin and taper out towards the north and east. Coarse-grained conglomerates and breccias form the Lower Miocene (presumably Karpatian) unit. The upper part of this unit consists of poorly sorted conglomeratic sandstones and non-marine pebbly siltstones (215 m total thickness). Badenian deposits comprising 619 m of clays, silts and corallinacean-bearing sandstones of the Tekeres Schlier and the Szilágy Formation follow. The Sarmatian Kozárd Formation (344 m) conformably overlies the Badenian. It consists of clayey marls and sandstones. The 1049 m thick Late Miocene (Pannonian) Peremarton Supergroup comprises clayey marls in the lower part and the Újfalu Sandstone and Algyő Formation in the upper part. The Pontian (1471 m) consists of alternations of sandy-clayey marls and sandstones, with clays, conglomerates and coal seams in its upper part. The deeper southern Zala sub-basin formed above basement, which consists of older crystalline complexes and Mesozoic units, but is generally poorly known. Basin development was generally comparable with that of the northern sub-basin. In the deepest part (Budafa area), sedimentation began during the Karpatian with the deposition of conglomerates. The overlying marine Badenian is very thick (>2900 m) comprising clayey marls, rare sandstone intercalations and a 150 m thick stratovolcanic complex in its lower part. The Sarmatian, if preserved, comprises sandstones and bituminous, clayey marls with fish scales. As in the northern sub-basin, the Upper Miocene and Lower Pliocene are represented by the Peremarton Supergroup (1155 m) and the Dunántúl Supergroup (825 m).

In the Drava Basin the thickness of the Neogene in the Croatian segment is up to 7000 m. The basement consists of Palaeozoic and Mesozoic crystalline and sedimentary units. The Neogene succession, as presented here, is based on Velic & Sokac (in Steininger *et al.* 1985). Deposition commenced during the Ottnangian and Early Karpatian, represented by 2000 m of conglomerates, sandstones, limestones and clayey marls. During the latest Karpatian, the marine character of the basin became dominant, but fully marine conditions were typical only from the Badenian, when the marine transgression reached the Drava Basin. Corallinacean limestones are widespread along the basin margins, while marls and pelites were deposited in the deeper parts of the basin. The total thickness of the Badenian is 1200 m. In the 400 m thick Sarmatian units there are alternations of marls, laminated bituminous marls, rare sandstones and diatomites. Towards the top of the Sarmatian, the regressive trend resulted in an erosional unconformity between the Sarmatian and the Pannonian. Thus, the top part of the Sarmatian is missing. The Pannonian units (up to 1000 m thick) comprise marls, clayey marls and subordinately sandstones, which were deposited in freshwater to brackish water environments. The overlying Pontian units (2500 m) are lithologically very similar but have coal seams and clays in the uppermost part.

Great Hungarian Plain

The Great Hungarian Plain (GHP) which forms part of the PBS covers parts of Hungary, Romania and the former Yugoslavia. The area of maximum subsidence in the GHP is situated in the valleys of the Danube and the Tisza (Fig. 17.16). Basement relief and sedimentary fill of the GHP are very heterogeneous (Nagymarosy 1981; Fig. 17.18). The area can be subdivided into extremely deep sub-basins (up to 7000 m), with a more or less complete Middle Miocene to Pliocene sedimentary sequence, and intrabasinal ridges and highs, with an incomplete sedimentary record (up to 2000 m and usually lacking the Middle Miocene). The sub-basins of the GHP include the Jászság Sub-basin (in the NW central zone), the Nyírség Sub-basin (in the NE), the Derecske Sub-basin (in the east central zone), and the Makó and Békés grabens (in the south). A further sub-basin located in the western part of the GHP (Kiskunhalas Sub-basin) is tectonically related to the South Transdanubian Mecsek Zone.

Palaeogeography and tectonic setting

The evolution of the GHP can be summarized as a history of gradual subsidence beginning in the latest Early Miocene and continuing through the Middle Miocene into the Pliocene or even Quaternary times. The rapid subsidence may have been related to the crustal attenuation of the area (Horváth *et al.* 1986), resulting in the formation of pull-apart basins located in strike-slip fault zones (e.g. the Balaton and the Mecsekalja lineaments). The sub-basins in the north subsided rapidly during the Middle Miocene and the Early Pannonian while in the south the main phase of subsidence took place during Late Pannonian, Pontian and Pliocene times (Tanács & Rálisch 1990). The oldest Neogene subsidence occurred in a zone extending from the Mecsek Mountains (Transdanubia) to the Kiskörös area where subsidence began in the middle Early Miocene (Ottnangian). Subsidence in the rest of the GHP commenced in the latest Early Miocene or Middle Miocene.

Marine transgression reached the GHP in Badenian times. During the Badenian and Sarmatian the GHP formed an archipelago of large islands, shallow and deep-marine basins. During the Sarmatian and Early Pannonian even the topographic highs (e.g. the Battonya-Pusztaföldvár and Algyő highs) were submerged. During the Early Pannonian the GHP attained its maximum depth. This event is documented first by the presence of the pelagic Endrőd-Tótkomlós Marl, an important potential source rock, and later by the turbidite-rich Szolnok Formation. At the end of the Pannonian, basin subsidence slowed down and gradual infilling of the basin began. Deltas prograded from the NNW to the SSW. While the northern part of the basin had already filled up and the paludal Bükkalja coal seams were deposited, in the Derecske, Makó and Békés sub-basins the deposition of delta-slope and delta-plain sediments continued. The youngest sub-basins, existing in the Pliocene, were in the Banat and Morava sub-basins located to the south in former Yugoslavia and in Romania.

Sedimentary and stratigraphic development

The pre-Badenian formations of the GHP are tentatively assigned to the Lower Miocene (Ottnangian and Karpatian). The oldest of the pre-Badenian formations is the terrestrial-limnic Madaras Formation (up to 100 m thick), which consists of conglomerates and thin tuff beds. The overlying Kiskunhalas Formation (80–

900 m) comprises sandy siltstones with conglomerates in limnic-terrestrial facies. The first Middle Miocene lithostratigraphic unit is the Tar ('Middle') Rhyolite Tuff and related volcanic complexes (e.g. Nyírség Volcanic Complex). The first marine event occurred in the Badenian and comprises the littoral/shallow neritic Abony Limestone with conglomerate intercalations, the Makó Formation (marls and siltstones) and the Ebes Limestone. In the Sarmatian both the brackish-water Hajduszoboszló Sandstone and the Dombegyháza Limestone were deposited in littoral to neritic settings.

The Pannonian and Pontian facies belts shifted in time and space due to the gradual infilling of the GHP (Pogácsás *et al.* 1990). Thus, separation of the Pannonian and Pontian units is difficult and the formations outlined below cannot be precisely dated (Jámbor 1989; Juhász 1991; Fig. 17.19). The oldest series of formations belongs to the transgressive Maros Group and include evidence of basaltic volcanism in some places (e.g. Kecel Formation). The basal Békés Conglomerate occurs mainly in the Makó and Békés grabens. The conglomeratic beds are overlain by the Endrőd Marl and the Tótkomlós Calcareous Marl Member, which were deposited in a deep offshore environment. Submarine fans (Dorozsma Marl Member) also occur. These were derived from steep slopes and occur only in the Makó and Derecske sub-basins. The Vásárhely Marl Member is a more pelitic variant of the Dorozsma Marl. The lava flows and tuffs of the Kecel Formation are interbedded with the sediments of the Dorozsma Marl.

The deposition of the overlying Jászkunság Group coincided with maximum deepening and the beginning of basin infill. The lower part of the group is represented by the offshore marls of the Nagykörü Member (Endrőd Formation) which is overlain by or interfingered with the turbidite-rich Szolnok Formation. Pontian sedimentation commenced with the deposition of the Csongrád Group comprising the Algyő and Törtel formations. The Algyő Formation (alternating clayey marls and sandstones) represents the delta slope of rivers filling the GHP. Delta-front and delta-plain environments are indicated by the sandy Újfalu (= Törtel) Formation. The overlying Heves Supergroup was deposited in the north during the Late Pontian and in the south during Pliocene times. The older unit of the supergroup is the sandy-clayey Zagyva Formation which represents an alluvial-plain environment and interfingers with the Bükkalja Lignite. The Pliocene succession ends with the clays of the lacustrine-paludal Nagyalföld Formation.

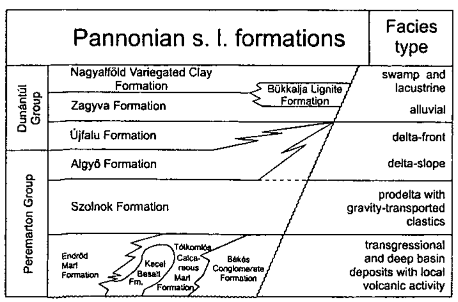


Fig. 17.19. Lithostratigraphy and facies types of the Pannonian in Hungary.

Szolnok 'Flysch' Basin (A.N.)

The Szolnok Basin (= Szolnok Flysch Belt) is a 130 km long and 20 km (rarely 40 km) wide, SW–NE striking basin that extends beneath the Great Hungarian Plain from the town of Szolnok (Hungary) into the Maramures area of Romania (Fig. 17.20) (Dudich & Bombita 1983). The basin strikes parallel to an elevated crystalline ridge between Túrkeve and Körösszegapáti. It can be traced toward the east to Carei and Satu Mare in Romania and disappears beneath the Gutii Mountains. It is known from several boreholes, especially from Hungary, but also from eastern Romania. North of the ridge the basement drops and is overlain by the deep-marine deposits (flysch) of the Szolnok Basin (Körössy 1959) and this contact may be tectonic.

Palaeogeography and tectonic setting

The Szolonok Basin is located on the NE margin of the Tisza microcontinent. Together with the Mesozoic Tisza Nappes they form a tectonic unit and represent the basement of the Neogene Pannonian Basin. Körössy (1959, 1977) and Szepesházy (1973) emphasized the strongly tectonized character of the deposits of the Szolnok Basin, with dips of between 70° and 90°. Palaeontological investigations indicate that deposition was not continuous, but occurred over discrete time intervals with long hiatuses between them. The preservation and recent distribution of the Szolnok Basin deposits is strongly controlled by Early Miocene compressional tectonics and subsequent Miocene denudation. Tectonic imbrication and erosion is also suspected, which would explain the lack, or scarcity, of Palaeocene and Cretaceous units. Thus, the Szolnok Basin has an exotic position within the Carpathian arc, having no direct connection with any of the Carpathian or Dinaric units. In terms of its palaeogeographic position, the Szolnok Basin may represent either (1) the tectonically displaced continuation of one of the Outer Carpathian units, or (2) the continuation of the Inner Carpathian units (i.e. the Transcarpathian 'flysch' in the Marmures area).

Stratigraphic development

The Senonian to Oligocene turbidite-dominated succession of deposits of the Szolnok Basin is often referred to as 'Szolnok Flysch' in the literature (see Szepesházy 1973). The >1000 m thick unit is covered by up to 3000 m of Neogene and Quaternary sediments. The precise thickness of the turbidite dominated deposits is unknown, since boreholes did not reach basement. The lithological composition of the Szolnok Basin succession is incompletely known, but it is non-continuous and can be subdivided into discrete time slices. Studies of core materials (Báldi-Beke *et al.* 1981; Nagymarosy & Báldi-Beke 1993) have shown that only a few Cretaceous and Palaeogene nannoplankton zones can be recognized.

The Szolnok Basin deposits overlie clastic deposits with alkaline basaltic complexes of Early Cretaceous age. Deposition commenced during the Late Cretaceous and is represented by calcareous shales, marls, clayey and calcareous marls of the Izsák Marl. In the West Carpathians, the Upper Campanian to Lower Maastrichtian Puchov Formation represents an equivalent for these beds. Turbiditic sandstones with graded conglomerates, and rarely with breccias, are interbedded into the generally shaly unit. Based on the palaeoecology of foraminifers, a pelagic and bathyal depositional environment is suggested for the Puchov Formation (Majzon 1966; Szepesházy 1973). In some units, non-calcareous and non-fossiliferous shales may indicate deposition below or close to the CCD. The top part of the Cretaceous and almost the entire Palaeocene are missing from the core material.

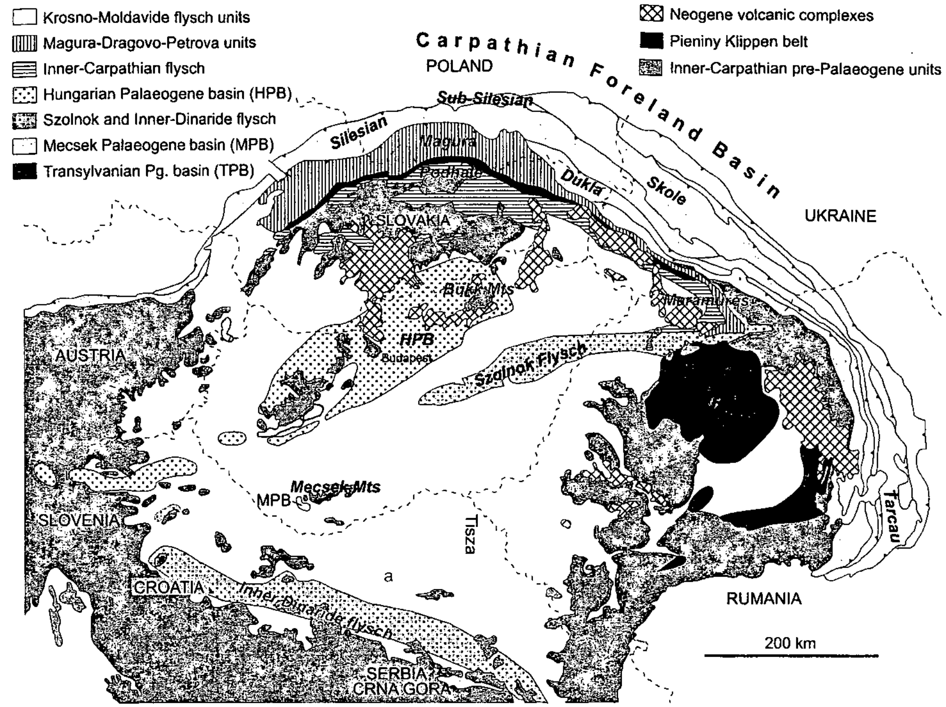


Fig. 17.20. Palaeogene basins in the Carpathian–Pannonian–Inner Dinarides region.

Some cores may cover the transition of the uppermost Palaeocene and Lower Eocene (NP9 nannoplankton zone: Nagymarosy & Báldi-Beke 1993; Szepesházy 1973; Majzon 1966).

The Middle and Upper Eocene part of the succession comprises shales, thin sandstones, polymictic conglomeratic sandstones, conglomerates and breccias. Rare nummulitid- and coralline algae-bearing limestones (with chaotic structure, probably redeposited) were observed. The Eocene calcareous nannoplankton assemblages indicate pelagic conditions and are fundamentally different from the coeval nearshore assemblages of the neighbouring epicontinental Transdanubian Palaeogene basin (Báldi-Beke 1984; Nagymarosy & Báldi-Beke 1993), proving the absence of a close palaeogeographic relation.

The presence of Early Oligocene sediments is uncertain. The Oligocene part of the sequence comprises clayey marls (similar to the Kiscell Clay) with sandstone intercalations. *Lepidocyclina*-bearing conglomerates are also recorded. The Oligocene nannoplankton assemblages are less pelagic and show more nearshore features than the Eocene ones, suggesting that the depositional environment gradually changed from pelagic deep-water to a shallower, nearshore one.

Transylvanian Basin (S.F.)

The Transylvanian Basin (TSB) is surrounded by the Eastern and Southern Carpathians and the Apuseni Mountains, and represents a post-Cenomanian–Neogene intra-Carpathian sedimentary basin, locally preserving up to 5000 m of sediments (Fig. 17.16). The basin developed on basement consisting of metamorphic, ophiolitic and sedimentary units assembled during the Middle Cretaceous (Ciupagea *et al.* 1970; Săndulescu 1984).

Palaeogeography and tectonic setting

The history of the TSB basinfill was strongly influenced by the evolution of the Carpathians. Its sedimentary record is subdivided into four tectonostratigraphic megasequences: Upper Cretaceous, Palaeogene, Lower Miocene, and Middle–Upper Miocene (Krézsek & Bally 2006).

The Upper Cretaceous clastic sequence was deposited in a basin generated by the collapse of the Mid-Cretaceous Orogen (Sanders 1999; Willingshofer *et al.* 2001). The succession is up to 1000 m thick and consists of coarse and fine-grained siliciclastics and limestones deposited in continental and marine (both shallow and deep) settings. Deposition of the Palaeogene sediments was coeval with a compressional post-rifting phase (Krézsek 2005). The sedimentation rate was low but constant and the alternating marine and continental cycles are separated by two third-order sequence boundaries related to minor inversion of the Palaeogene basin. During the Early Miocene, the flexural sub-basin generated by the thrusting of the Pienides was developed almost exclusively in the northern part of the TSB (up to 1000 m of sediments). Several unconformities related to synsedimentary tectonic activity have been recorded in the shallow- to deep-marine sediments.

Major tectonic changes from the beginning of the Middle Miocene (Badenian) transformed the TSB into a backarc basin (Săndulescu 1984), characterized by a dominantly compressional regime and a high rate of subsidence between the Late Badenian and the Early Pannonian. Early Badenian hemipelagic to shallow-marine sedimentation (less than 100 m) was followed by an important phase of Middle Badenian evaporite deposition (300 m). Subsequent siliciclastic deposition (up to 3000 m) occurred in deep-marine (Late Badenian), outer-ramp (Early Sarmatian), shelf and delta (Late Sarmatian to Pannonian) settings (Krézsek & Filipescu 2005). Most of the Upper Miocene basin fill was eroded due to the post-Pannonian uplift and erosion.

Sedimentary and stratigraphic development

The **Palaeogene megasequence** (Fig. 17.21 MS1 & MS2) is divided by two unconformities (Hosu 1999) into three sedimentary phases, corresponding to the Danian(?)–Early Priabonian, the Late Priabonian–Early Rupelian and the Late Rupelian–Chattian (see also Filipescu 2001*a*). Close to the K/Pg boundary, tectonic compression led to a major change in the depositional environments, initiating erosion and a change to alluvial fan and fluvial sedimentation (Jibou Formation; Hosu 1999). Fossil vertebrates have been described from the lacustrine deposits in the vicinity of the Thanetian–Ypresian boundary (Gheerbrant *et al.* 1999), while deep-marine deposits with Danian foraminifera were identified in the northern part of the basin (Filipescu & Kaminski in press).

Beginning in Bartonian times, a marine transgression initiated the deposition of the Călata Group (= 'lower marine series'). Sea-level rise led to the onset of the supratidal to intertidal evaporitic facies of the Foidaş Formation, overlain by the shallow-marine inner-platform deposits of the Căpuş Formation, and finally the offshore deposits of the Mortănuşa Formation. In marginal areas, the low rate of basin subsidence led to the development of carbonate platforms (Rusu 1995; Proust & Hosu 1996), as represented by the Inucu Formation (mudstones with intercalations of bioclastic limestones, and *Crassostrea bersonensis* bioherms), the Văleni Limestone (bioclastic shoals with very diverse micro- and macrofaunas), the Ciuleni Formation (siliciclastics and bioclastic calcarenites dominated by bivalves such as *Crassostrea orientalis* and foraminifera), and the Viştea Limestone (bioclastic bars with Priabonian large foraminifera, molluscs, echinoids and crustaceans).

Compressional tectonics interrupted the shallow-marine sedimentation and changed the depositional environment to a coastal plain and fluvial system (e.g. Valea Nadăşului Formation; Hosu 1999). Its fossil record consists of freshwater gastropods, plants and mammals.

The Upper Priabonian–Lower Rupelian fill represents a new marine cycle (Turea Group = upper marine series) consisting mainly of shallow-marine sediments. Deposition commenced with a Priabonian transitional freshwater to evaporitic unit (Jebuc Formation), followed by a carbonate platform (Cluj Limestone) with *Crassostrea transilvanica* and *Nummulites fabianii*. Maximum flooding is indicated by offshore bryozoan marls (Brebi Formation) which also include the Priabonian–Rupelian boundary (*Pycnodonte gigantica* biohorizon). Progressive shallowing is recorded in the overlying Hoia Limestone (skeletal packstones) and the Mera Formation (siliciclastic and calcareous).

Tectonic events at the end of the Early Rupelian led to a progressive shift to fluvial and siliciclastic ramp environments (e.g. Moigrad Formation). The Upper Rupelian is strongly transgressive, including dysoxic environments (bituminous shales of the Ileanda Formation).

The Chattian is mainly siliciclastic with an overall progradational pattern and widespread continental deposition, probably generated by a eustatic sea-level fall and climate change. While most of the southern part of the basin was exposed, and continental deposits were dominant in the Gilău and Meseş area, marine inner shelf (Buzaş Formation) and offshore to slope (Vima Formation) deposits were dominant in the Preluca area.

The base of the **Lower Miocene megasequence** (MS3) is coincident with deposition of the marine Vima Formation which was continuous across the Oligocene–Miocene boundary. Subsequent retrogradation was coeval with the transgressive event which was generated by the onset of compressional tectonics and which created a flexural basin (Györfi *et al.* 1999). Microfaunas with Mediterranean affinities (Popescu *et al.* 1995) invaded the basin, and the overlying Eggenburgian littoral sands of the Coruş Formation preserve a typical assemblage of bivalves with *Oopecten gigas* (Moisescu & Popescu 1980). Early Miocene maximum flooding is associated with the deposition of the Eggenburgian Chechiş Formation, an offshore unit preserving diverse microfaunal assemblages with *Globigerinoides trilobus* (Popescu 1975). Following the deposition of the Chechiş Formation, the basin was filled during the remaining Early Miocene with deep-marine turbidites and coarse-grained fan deltas belonging to the Hida Formation.

The base of the **Middle to Upper Miocene megasequence** (MS4) coincides with the Middle Miocene (Badenian) transgression which was synchronous in all of the Paratethyan basins. Carbonate and clastic sedimentation dominated shallow ramp environments (Filipescu 1996), while hemipelagites were deposited in the deeper areas (less than 100 m thick). The Lower Badenian Dej Tuff testifies to the local intense volcanic activity (Seghedi & Szakács 1991) and represents a useful regional stratigraphic marker.

Initial transgressive conditions are documented by two important planktonic foraminfera invasions (*Praeorbulina glomerosa* and *Orbulina suturalis* biozones; Popescu 1975). Benthic assemblages show affinities to offshore and shoreface siliciclastic and carbonate environments. The upper part of the Lower Badenian succession indicates a progressively shallower facies, terminating in lowstand conditions (Krézsek & Filipescu 2005). At the beginning of the Middle Badenian, an important transgressive event renewed the planktonic and benthic assemblages, suggesting a deeper depositional environment when compared to the Early Badenian (Filipescu 2001*b*).

A major relative sea-level fall from the late Mid-Badenian (near the end of the Langhian) induced progressive restriction in basin circulation and created conditions for the deposition of sabkha to shallow-ramp gypsum in the west (Cheia Formation) and deep-ramp salt (Ocna Dejului Formation). The salt crops out along two major lineaments situated at the western and eastern margins of the basin as a result of its post-depositional tectonics (Krézsek 2004).

The marine flooding event at the beginning of the Late Badenian (Popescu 1972; Dumitrică *et al.* 1975) was probably both globally and regionally (increased subsidence rates due tectonic shortening in the Eastern Carpathians: Săndulescu 1988) controlled. Transgression resulted in the deposition of hemipelagic sediments, subsequently overlain by the highstand prograding submarine fans of the Pietroasa Formation (Filipescu 2004). By the end of the Late Badenian, increased regional compression resulted in a relative sea-level fall and lowstand deposition. Due

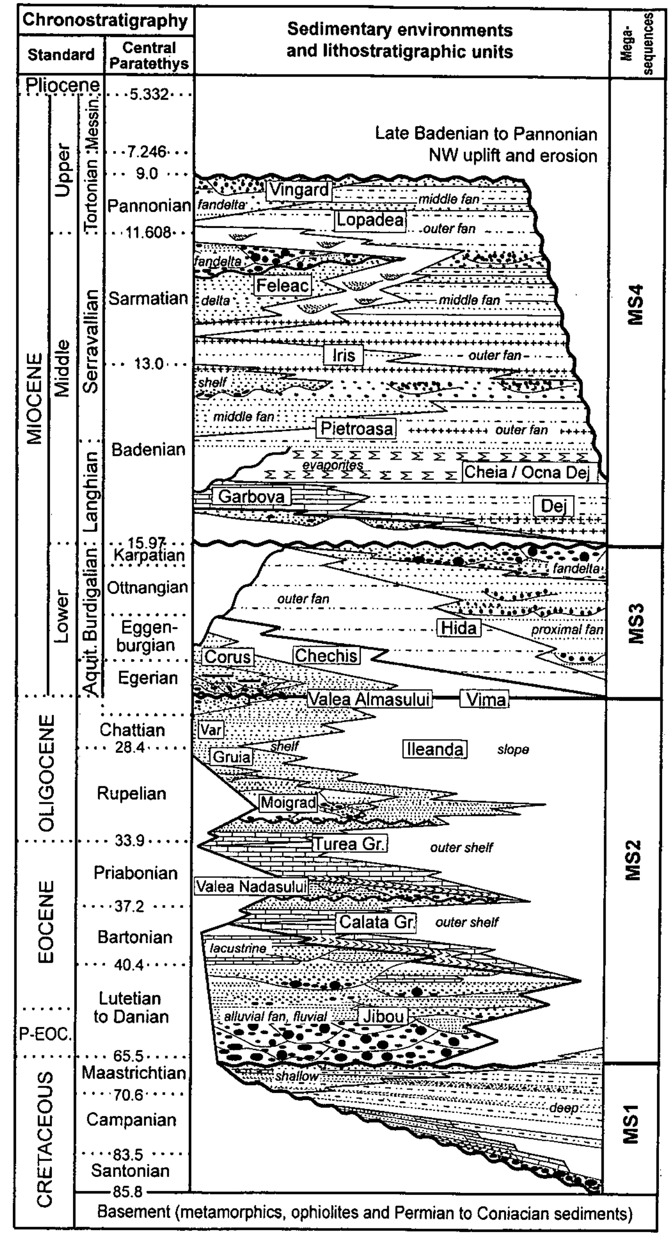


Fig. 17.21. Lithostratigraphy of the Transylvanian Basin (based on Krézsek & Bally 2006).

to the particular tectonic setting, the total thickness of the Upper Badenian ranges from a few metres (in the west) to >1500 m (in the SE).

The subsequent transgressive deposits near the Badenian–Sarmatian boundary (planktonic foraminifera and *Anomalinoides* biozones) were followed by prograding submarine fans (Sarmatian Iris Formation). The subsequent sea-level fall resulted in increased salinity in the basin (*Varidentella reussi* Biozone: Popescu 1995), and extensive erosion in the northern part of the basin. Another transgressive event (*Elphidium reginum* Biozone) was associated with deltaic progradation (*Dogielina sarmatica* Biozone) (Krézsek & Filipescu 2005). By Late Sarmatian times, the onset of uplift along the basin margins (Apuseni Mountains) resulted in the deposition of deltas, fan deltas, coarse submarine fans (Feleac Formation), and a major unconformity. Sarmatian deposits range in thickness between 300 m and >1000 m (in the basin centre).

The overlying shallow-ramp and deep-marine siliciclastic deposits are associated with a new transgression at the end of the Sarmatian (*Porosononion aragviensis* Biozone). This extended into the earliest part of the Pannonian (Lopadea Formation); outer-ramp condensed sedimentation predominated. Fan deltas, ramp and deep lacustrine sediments are typical of the subsequent prograding systems (Krézsek 2005). The brief re-establishment of connections with the outer Carpathian region can be inferred from a brackish marine ingression (*Ammonia* acme), generated by an extrabasinal relative sea-level rise. Fan deltas, proximal lacustrine fans, incised channels and alluvial fan systems were the most characteristic features of the middle part of the Lower Pannonian. The subsequent transgressive and highstand conditions are documented only in the eastern part of the TSB, and comprise an overall coarsening-upward succession (Vingard Formation). The thickness of the Pannonian deposits ranges from 300 m (SW) to 500 m (SE). The Pliocene to Holocene evolution of the TSB was characterized by uplift, erosion (Sanders 1999) and structural complications produced by salt tectonics (Krézsek & Bally 2006).

Volcanism

Mid- to Late Miocene Carpathian deformation was accompanied by intense magmatic activity in the volcanic arc covering most of the eastern margin of the TSB. Volcanic products sealed parts of the Pannonian sedimentary record. There are several well-known volcanic tuff lithohorizons in the Badenian–Pannonian sedimentary record of the TSB (Mârza & Mészáros 1991), which are related to Neogene calc-alkaline, subduction-related magmatic activity of the Carpathians and Apuseni Mountains (Pécskay *et al.* 1995).

Outer Carpathians in Poland (N.O.)

The Polish Outer Carpathians (POC) are part of a mountainous arc that stretches for >1300 km from Vienna to the Iron Gate of the Danube River at the Serbian–Romania boundary (near Orsova) (Figs 17.16 & 17.22). In the west, the Carpathians are linked to the Eastern Alps and in the east they pass into the Balkan chain. The POC are *c.* 315 km long and 60–90 km wide, and are located in southern Poland.

Tectonic setting

Traditionally, the Western Carpathians are subdivided into two distinct ranges (Książkiewicz 1977; see also Froitzheim *et al.* 2008). The Inner Carpathians are considered to be the older range, while the Outer Carpathians represent the younger range. Between them, the Pieniny Klippen Belt represents a Cenozoic strike-slip boundary; this is a strongly tectonized terrain *c.* 600 km long and 1–20 km wide. The Outer Carpathians are formed by stacked nappes and thrust sheets, which show different lithostratigraphic and structural development. From south to north they include: the Magura, Fore-Magura-Dukla Group, and the Silesian, Sub-Silesian and Skole units. In the Outer Carpathians the main décollement surfaces are located at different stratigraphic levels. All of the Outer Carpathian nappes are flatly overthrust onto the Miocene deposits of the Carpathian Foredeep (Oszczypko 1998; Oszczypko & Tomaś 1985). However, along the frontal Carpathian thrust a narrow zone of folded Miocene deposits developed. In Poland these are represented mainly by the Zgłobice, and partially by the Stebnik units. The detachment levels of the folded Miocene units are usually connected with Lower and Middle Miocene evaporites.

Palaeogeography

The Outer Carpathian Basin occupied the northern margin of Neo-Tethys (Oszczypko 2004, 2006). Geomagnetics revealed that the boundary between the North European Plate and the Central West Carpathian Block was probably located 60–100 km south of the present-day position of the Carpathian margin (Oszczypko & Ślączka 1985; Żytko 1997; Oszczypko 2006). The Outer Carpathian sedimentary area can be regarded as the remnant of an oceanic basin transformed into a foreland basin (Oszczypko 1999). This basin developed between the colliding European continent and various intra-oceanic arcs. The important driving forces behind tectonic subsidence in the basin were syn- and post-rift thermal processes as well as the emplacement of the nappe loads related to subduction processes (Poprawa *et al.* 2002). Similar to other orogenic belts, the Outer Carpathians were progressively folded towards the continental margin. According to the reconstructions of Roure *et al.* (1993), Roca *et al.* (1995) and Behrman *et al.* (2000), the Early Oligocene Outer Carpathian Basin was at least 380 km wide.

Palaeomagnetism

According to Marton *et al.* (1995*b*) the Western Carpathians and its hinterland are characterized by a counterclockwise rotation of 80°. The first rotation of 50° took place during the Late Ottnangian to Early Karpatian and the second one (of 30°) occurred during the Badenian.

Sedimentary and stratigraphic development

The Outer Carpathians are composed of mainly Upper Jurassic to Lower Miocene deep-marine deposits ('flysch') (Fig. 17.23). The sedimentary successions of the main tectonic units differ in terms of facies development and thickness. The thickest sedimentary cover belongs to the Silesian unit, which varies from 3000 m in the west to >5000 m in the east. The thickness of the other tectonic units is markedly thinner and varies between 3000–3800 m in the Skole Unit, *c.* 1000 m in the Sub-Silesian Unit, 2300–2500 m in the Dukla Unit, and 2500–3500 m in the Magura Nappe (Poprawa *et al.* 2002). In terms of facies distribution, thickness of deposits, and palaeocurrent directions (only in the Magura area), the Silesian and Skole basins can be considered as independent sedimentary areas. The Magura and Silesian basins were separated by the Silesian Ridge, while the Sub-Silesian area formed a submarine high dividing the Skole and Silesian basins during the Late Cretaceous to Eocene (Książkiewicz 1962).

The sedimentary succession of the POC comprises four depositional megasequences, which reflect the main stages of basin development (Poprawa *et al.* 2002; Oszczypko 2004): Middle

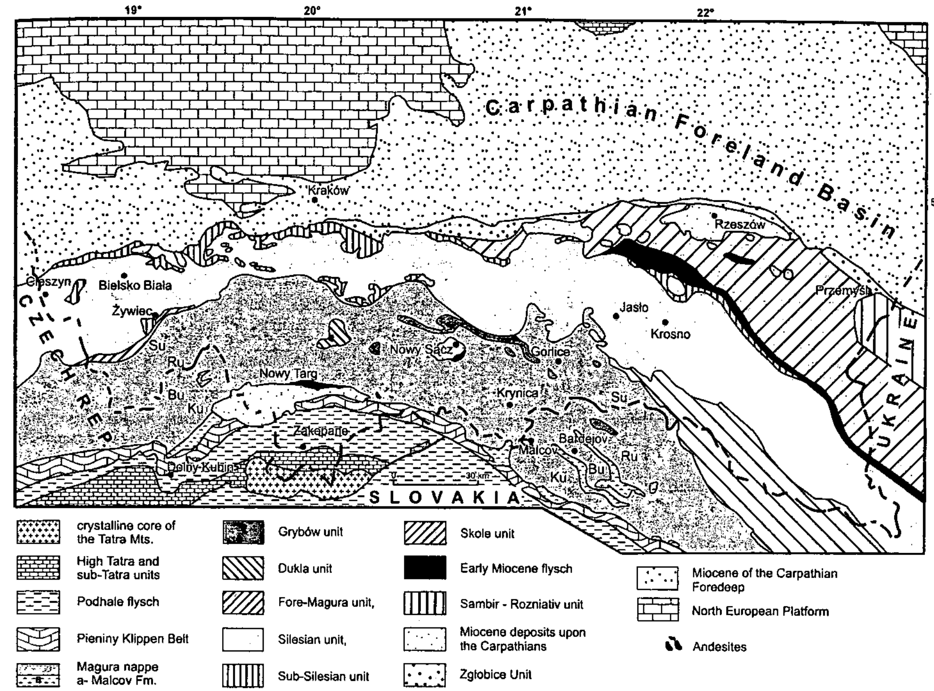


Fig. 17.22. Geological sketch map of the Polish Carpathians and their foreland basin. Abbreviations of facies-tectonic subunits of the Magura Nappe: Bu, Bystrica Subunit; Ku, Krunica Subunit; Ru, Rača Subunit; Su, Siary Subunit.

Jurassic–Early Cretaceous basin opening; Late Cretaceous–Palaeocene inversion; subsidence during Palaeocene to Middle Eocene times; and Late Eocene–Early Miocene synorogenic closing of the basins. In the Outer Carpathian area the K/Pg boundary is located within a continuous turbidite sequence.

The Palaeocene represents a time of major changes in sedimentary conditions in the Outer Carpathian basins, ranging from uplift of the intrabasinal source areas to general subsidence and a sea-level rise. These changes are marked by the deposition of thin- to medium-bedded turbidites, traditionally referred to as the 'Inoceramian Beds' (Bieda *et al.* 1963), and recorded from the Magura, Dukla and Skole successions. The 300–500 m thick units are characterized by calcareous, micaceous sandstones intercalated with clay shales. During this time the evolution of the Silesian Basin was still influenced by the Silesian Ridge, from which non-calcareous, thick-bedded turbidites and conglomerates (Istebna Beds) were shed into the basin. The upper part of the Istebna Beds comprises up to 200 m of shales. The Lower Eocene is characterized by the occurrence of thick-bedded turbidites and conglomerates (up to 500 m), which form elongated lenses (Ciężkowice Sandstones). The northern periphery of the Silesian Basin was dominated by marls with exotic pebbles (sub-Silesian succession).

During the Eocene, a broad connection between the Outer Carpathian basins and the world oceans was established (Golonka *et al.* 2000; Golonka 2004), resulting in the unification of facies, including the position of the CCD level, as well as low sedimentation rates. This general trend dominated the Early and Middle Eocene in the Skole, Sub-Silesian, Silesian and Dukla basins, as well as in the northern part of the Magura Basin. The deepest parts of these basins were occupied by hemipelagic, non-calcareous shales. During the Middle Eocene these shales passed up into shales with intercalations of thin-bedded sandstones (Hieroglyphic Beds).

In the Magura Basin, Palaeocene deposits are overlain by shales (up to 150 m), which pass upwards into the thin-bedded turbidites of the 200–600 m thick Beloveza Formation (Oszczypko 1991). The upper part of the Lower/Middle Eocene succession comprises the thick-bedded turbidites of the Magura Formation, which is 1000–2000 m thick (Birkenmajer & Oszczypko 1989). Eocene deposition in the Magura Basin was controlled by the activity of an accretionary wedge which developed on the southern margin of the basin during the Late Palaeocene collision between the Inner Western Carpathian block and the Pieniny Klippen Belt. During the Eocene the migrating load of the accretionary wedge led to further subsidence and a shift in the depocentres to the north. As a result, a long and narrow submarine fan developed. The northern, deepest part of the basin, often below the CCD, was dominated by basinal turbidites and hemipelagites. The sedimentation rate varied from 6–18 m/Ma on the abyssal plain to 103–160 m/Ma on the outer

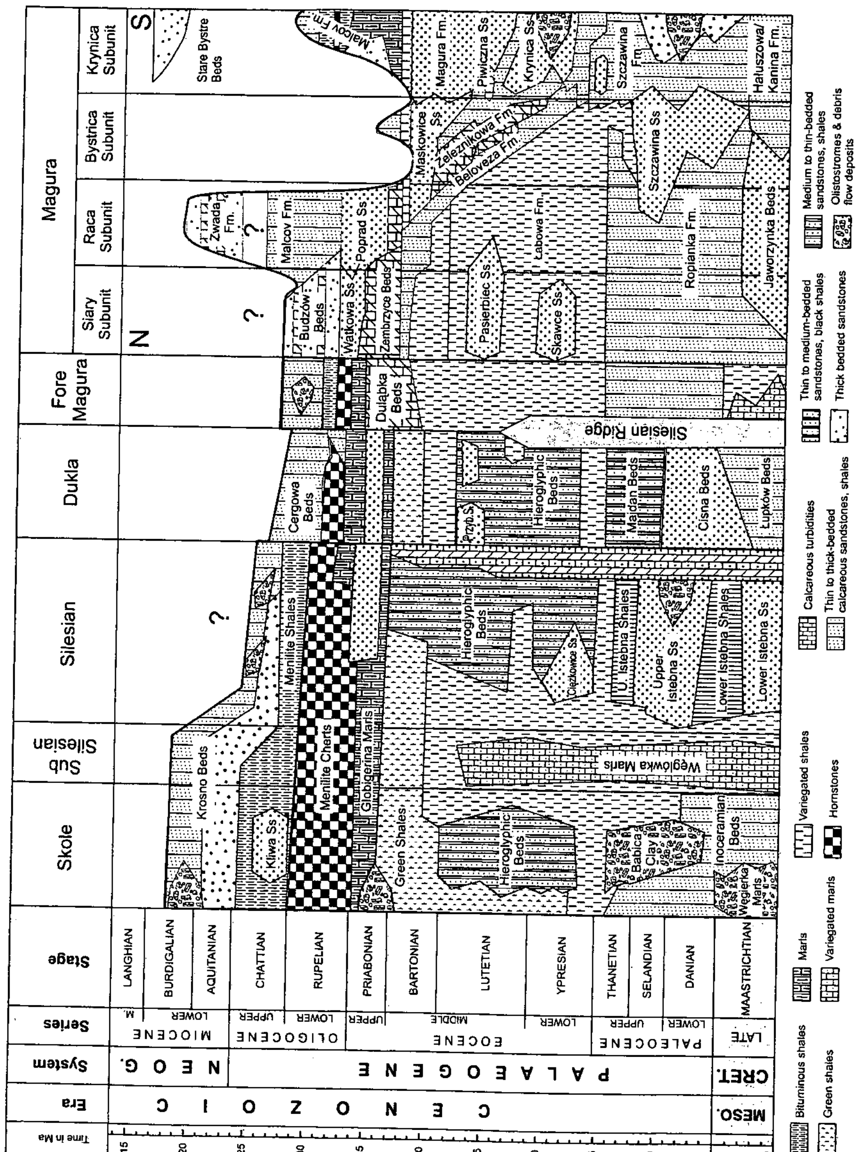


Fig. 17.23. Regional stratigraphic scheme of the Palaeogene and Lower Miocene deposits of the Polish Outer Carpathians (after Oszczypko 2004).

fan, and 180–350 m/Ma in the area affected by the middle fan-lobe systems (Oszczypko 2004). The fan was supplied from the SE, probably from the Inner Carpathian/Inner Dacide area.

During the Priabonian and Rupelian, uplift in the Outer Carpathian Basin (Poprawa *et al.* 2002; Oszczypko 2004) was accompanied by the transformation of the Outer Carpathian remnant oceanic basin into a foreland basin. This led to the replacement of deep-water shales and basinal turbidites by pelagic *Globigerina* Marls, followed by Oligocene bituminous Menilite shales deposited in the newly restricted basin. This type of deposition dominated all of the Outer Carpathian basins, except for the main part of the Magura Basin (Van Couvering *et al.* 1981). During that time, the pelagic *Globigerina* marls and Menilite shales of the Magura Formation, as well as the turbidites of the Malcov Formation (Middle/Upper Oligocene), were deposited in the SE part of the Magura Basin. The northernmost part of the Magura Basin, which was supplied from the NW (Silesian Ridge), is characterized by Middle Eocene shales that pass upwards into marls and thin-bedded turbidites (Zembrzyce Beds, Upper Eocene), thick-bedded glauconitic sandstones (Wątkowa Sandstone, Lower Oligocene), and finally into glauconitic sandstones and marls (Budzów Beds, Oligocene) (Oszczypko-Clowes 2001). In the marginal parts of the Outer Carpathian basins, *Globigerina* Marls pass into Menilite Shales (up to 250 m thick) with chert horizons at the base. The Menilite Shales, which contain lenses of thick-bedded sandstones (Cergowa and Kliwa Sandstones), gradually pass upwards into a complex of thick-bedded, medium-grained, calcareous sandstones (Lower Krosno Beds, up to 800–1000 m). Higher up in the succession, thin-bedded, fine-grained sandstones pass into marly shales (Upper Krosno Beds, Early Miocene, up to 3000 m). These lithofacies are diachronous across the Silesian, Sub-Silesian and Skole basins (Ślączka & Kamiński 1998).

Following the phase of Late Oligocene folding, the Magura Nappe was thrust northwards onto the terminal Krosno Basin. This resulted in the final phase of subsidence in the Outer Carpathian Basin (Late Oligocene–Early Miocene). Subsidence was accompanied by a progressive migration of the depocentre axes towards the north and increased depositional rates ranging from 350 m/Ma in the Rupelian (northern part of Magura basin) to 600 m/Ma at the end of the Oligocene (SE part of Silesian Basin). During the Eggenburgian a piggyback basin developed on the Magura Nappe (Zawada and Kremna formations, at least 550 m thick) and a marine connection with the Vienna Basin via Orava was established (Oszczypko *et al.* 1999, 2005; Oszczypko & Oszczypko-Clowes 2002, 2003).

Volcanism

The final stages of tectonic evolution of the Polish Outer Carpathians were associated with volcanic activity related to subduction beneath the front of the Carpathian arc. In the Krościenko-Szczawnica area small andesite intrusions are located along the northern margin of the Pieniny Klippen Belt. The time of the volcanic activity in this area has been dated as 11.35 ± 0.45 Ma (Birkenmajer & Pecskay 2000).

Carpathian Foredeep in the Czech Republic (S.N., N.D., Š.H., P.P.)

The Carpathian Foredeep in the area of the Czech Republic (CCF) is limited to the NW by the passive margin of the Variscan Bohemian Massif and its Mesozoic platform cover and by the overriding nappes of the Western Carpathians in the SE. The CCF continues southwards into the Austrian part of the North Alpine Foreland Basin and northwards into the Polish Carpathian Basin. The basin is *c.* 210 km long and *c.* 6–40 km wide.

Tectonic setting

The CCF is a peripheral foreland basin that developed due to subsurface loading of the Alpine-Carpathian orogenic belt on the passive margin of the Bohemian Massif. The CCF exhibits striking lateral variations in terms of basin width, depth, and the stratigraphy of the sedimentary infill. There are also variations in pre-Neogene basement composition and tectonic subsidence. The main area of subsidence shifted from the south, in the Late Oligocene/Lower Miocene to the north in Middle Miocene times. This reflects the change in orientation of the main convergence direction and a generally eastward-directed lateral extrusion of the Carpathian blocks. The reactivation of NW–SE and NE–SW trending basement faults was also important for the basin shape and for its extent. The southern part of the CCF had a complicated evolution due to its position at the junction between the Eastern Alps and the Western Carpathians.

Sedimentary and stratigraphic development

Deposition in the CCF began in Egerian–Early Eggenburgian times (Fig. 17.24). Terrestrial (alluvial and fluvial, Žerotice Member) and deltaic sediments occur locally at the base of the succession in the SW of the basin. During the subsequent transgression from the south, SE and east, these deposits progressively passed into marine ones (clastic coast and shallow-marine environments). The shallow-marine facies (about 40 m palaeowater depth) was interdigitated with lagoonal and deltaic deposits (Nehyba *et al.* 1997). The upper part of this Lower Miocene succession comprises deltaic and clastic coast deposits (Čejkovice Sands). The source area was the deeply weathered margin of the Bohemian Massif. A distal rhyolite airfall tephra horizon occurs in the Upper Eggenburgian (Nehyba 1997). Dunajovice and Dyjákovice sandstones and the Dobré Pole Claystone (glauconitic sands and calcareous clays with rich foraminiferal fauna) were deposited under marine conditions in the SE part of the CCF during Early Miocene times (Adámek 2003; Chlupáč *et al.* 2002). Eggenburgian deposits also occur in the northern part of CCF in the area of Ostrava. Basal fluvial sands and gravels pass upwards into marine claystones, bryozoan limestones, sandstones and conglomerates (Jaklovec Conglomerate) (Chlupáč *et al.* 2002). Lower Miocene basaltic volcanism is known from the Opava region.

A second depositional cycle is confined to the middle Early Miocene (Ottnangian) and was related to the onset of thrusting, which led to markedly different depositional histories in the proximal and distal parts of the CCF. In the distal (western) part, the deposition was directed by the interplay of basement relief, sediment supply and eustasy, while Late Eggenburgian thrusting affected the deposition in the proximal parts of the basin. The deposition of a synorogenic clastic wedge, related directly to flexural subsidence due to the thrust loading, reflects this change (Nehyba 2004). At the base of the succession a typical basal forebulge unconformity occurs (Crampton & Allen 1995). Towards the passive margin, the unconformity is overlain by progressively onlapping sediments above an increasing stratigraphic gap. Changes in the heavy mineral spectra (Nehyba & Buriánek 2004) reflect the new situation. Brackish to fluvial sands and clays (Vítonice Clays), and sands and gravels of the Rzehakia Beds, comprise the Ottnangian deposits in the distal parts of the basin. These deposits (with the endemic bivalve *Rzehakia socialis*) were recognized in the Miroslav and Frýdek-

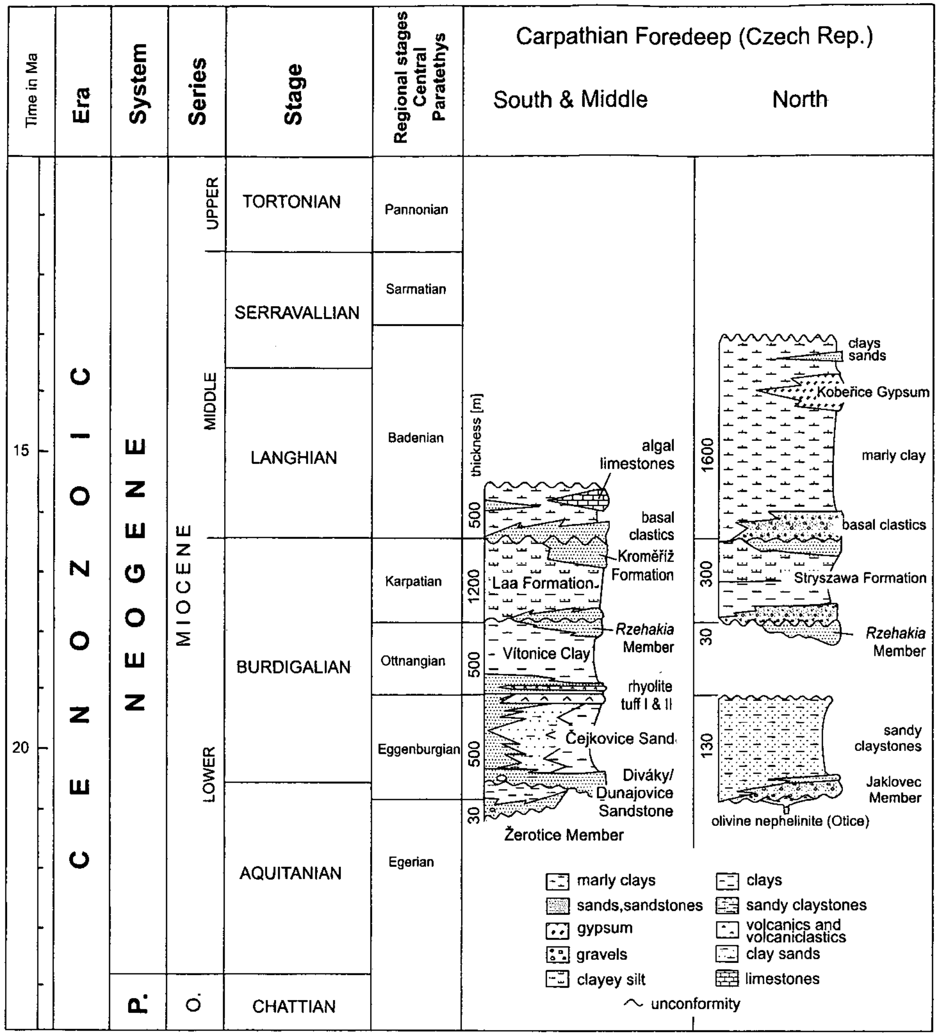


Fig. 17.24. Regional stratigraphic scheme of the Neogene of the Carpathian Foredeep in Moravia (Czech Republic) (after Brzobohatý in Chlupac *et al.* 2002; Adamek *et al.* 2003; Adamek 2003).

Místek areas (Čtyroký 1991). The flora indicates a humid climate (Doláková *et al.* 1999). Layers of acidic volcaniclastics (Ottnangian) were also deposited (Nehyba & Roetzel 1999). Fine-grained siltstones, claystones and fine sandstones (Late Eggenburgian and Ottnangian in age) were deposited in the proximal parts of the basin.

Continued compression along the Western Carpathian Front led to the progression of a westward-prograding and eastward-thickening clastic wedge during the late Early Miocene (e.g. Laa Formation; Adámek *et al.* 2003). Coastal, shallow-marine and deep-marine (bathyal) conditions alternate. Laminated silty clays (= Schlier) with abundant microfauna are typical (e.g. foraminifers; Petrová 2004). These were deposited on the deeper shelf under fluctuating oxygen contents. In the middle and southern parts of the CCF, the marine transgression covered an area of relatively flat relief and extended far across the Bohemian

Massif; marshes were established on the continental margin (Doláková & Slamková 2003). In the northern part of the CCF, the upper Lower Miocene (Karpatian) deposits are represented by the Stryszawa Formation comprising terrestrial basal conglomerates, sandstones and claystones, and lacustrine/lagoonal siltstones that pass into marine deposits (claystones and siltstones, and subordinate gravels and sandstones). Clays, silts, sandstones and conglomerates with gypsum form the uppermost part of the succession (Adámek *et al.* 2003).

Tectonic activity along the active margin affected the deposits of the CCF. Sediments in the foreland of the Carpathian nappes were shingled and stacked in the form of duplexes. Sandstones, conglomerates and clays (gravity current deposits) in front of the Ždánice Nappe in the middle part of the basin belong to the Kroměříž Member.

During the early Middle Miocene the CCF geometry was reorganized, because the NNW- and NW-directed compression of the Carpathian orogenic wedge changed to a NNE- and NE-directed compression during the Late Karpatian and Early Badenian. In the south, deposits are located in the central parts of the basin at some distance from the passive and active basin margins. The areas with the maximum sediment thickness are situated along the basin axis forming an almost symmetric basin with a SSW–NNE orientation. Deposits of coarse-grained Gilbert deltas, coastal environments, lagoonal deposits and deeper-marine deposits have been recognized. These are partly correlated with the Grund Formation in the North Alpine Foreland Basin. The maximum thickness of the deltaic gravels and gravelly sands is about 175 m. Deltas are located along both basin margins. Abundant intraclasts (sometimes several metres in diameter) reflect cannibalization of the older basin infill. In the northern part of the basin, the basal conglomerates and sandstones may have been deposited in marine environments (Chlupáč *et al.* 2002). The deeper-marine deposits (*c.* 600 m) comprise mudstones with silts, clays and shell debris. These are interpreted as outer-shelf deposits or hemipelagites. The presence of otoliths confirm a palaeowater depth of about 400 m in the central part of the basin (Brzobohatý 1997). Coastal and lagoonal deposits, as well as occurrences of algal and bryozoan limestones and calcareous sandstones, are restricted both areally and in thickness. Horizons of distal airfall tephra (Nehyba 1997) are related to the acidic calc-alcaline volcanism of a volcanic arc in the Carpatho-Pannonian area (Middle Rhyolite Tuff). Isolated relicts of Lower Badenian deposits can be found far to the west (e.g. Hostim, Kralice, Borač, Ústí nad Orlicí). Early Badenian terrestrial deposits (sands and clays with lignite seams) are developed in the Opava area in the northern part of the basin.

The subsequent cessation of marine sedimentation was largely due to the final stages of thrusting of the orogenic wedge along the NE part of the basin and to flexural uplift of the foreland in the SW. Middle and Upper Badenian deposits are known only from the north of the CCF in the Ostrava and Opava areas where the basin continued into Poland. Evaporites (Kobeřice Gypsum) and clays (*c.* 60 m), including a distal rhyolitic tephra horizon, are of Middle Badenian age (Nehyba 2000). Clays and claystones with abundant plant remnants and rare limestones represent the Upper Badenian (Cicha *et al.* 1985).

Fluvial and lacustrine deposition occurred during the Middle and Late Miocene. Moldavite-bearing sands and gravels are found in SW Moravia (Trnka & Houzar 1991). The reactivation of NW–SE trending faults led to the formation of brackish basins in the Mohelnice Graben and in the Upper Morava Valley. The Upper Miocene and Pliocene fluvial and lacustrine deposits have a thickness of about 200 m (Růžička 1989) in this area (= Lower and Upper Formation).

Carpathian Foredeep in Poland (N.O.)

The *c.* 350 km long Polish Carpathian Foredeep Basin (PCFB) is part of a large sedimentary basin that stretches from the Danube in Vienna (Austria) to the Southern Carpathians in Romania (Fig. 17.22). Towards the west it is replaced by the North Alpine Foreland Basin. The width of the PCFB varies from 30–40 km in the west to 90 km in the Rzeszów area and 50 km at the Polish–Ukrainian border. It can be subdivided into an 'inner' and the 'outer' part. The boundary between them runs along the Carpathian Thrust.

Palaeogeography

The Early to Middle Miocene PCFB developed as a peripheral foreland basin and is related to the advancing Carpathian Front (Oszczypko 1998; Oszczypko *et al.* 2006, and references therein). During late Early Miocene times the Western Outer Carpathians overrode the European platform. The load of the Carpathian nappes resulted in the development of a 25–50 km wide Early Miocene flexural depression (inner foredeep). This basin developed both on the top of the advancing Carpathian Front and on the platform basement. The relatively narrow brackish or freshwater, moat-like basin was filled with coarse clastic sediments derived from the emergent front of the Outer Carpathians and from the platform. The Lower Miocene deposits of the inner foredeep formed a clastic wedge along the Northern Carpathians, comparable with the 'Lower Freshwater Molasse' of the western North Alpine Foreland Basin.

Latest Early Miocene times were characterized by the development of SW–NE and NW–SE trending grabens. Subsidence, which affected both the inner and outer foredeep, was accompanied by the Early Badenian transgression (Oszczypko 1998; Oszczypko *et al.* 2006) which flooded both the foredeep and the marginal part of the Carpathians. Simultaneously, the shoreline shifted northwards, from 30–40 km to 100 km in the western and eastern parts of the basin, respectively. The palaeobathymetry of the Early Badenian basin varied between upper bathyal depths in the axis of the basin, to neritic and littoral conditions in the marginal areas. The deepest part of the basin (inner foredeep), which reached mid-bathyal depth, was located about 20 km south of the present-day position of the Carpathian Front. In the Carpathian offshore area, Lower Badenian deposits onlapped, deformed and eroded deposits of the Skole and Silesian/Sub-Silesian nappes and this resulted in an angular unconformity. Middle Badenian shallowing in the PCFB caused partial isolation of the basin, and initiated a Badenian salinity crisis in shallow areas (inner–middle neritic? depths; see Kasprzyk 1993; Peryt 2001; Bąbel 2004; Oszczypko *et al.* 2006). A new period of intense subsidence was initiated during the Late Badenian and was confined to the inner and outer foredeep areas (Oszczypko 1998). Late Badenian subsidence continued into the Sarmatian. Depocentres shifted 40–50 km towards the NE and rotated clockwise by up to 20°.

Tectonic setting

The evolution of the PCFB was controlled by synsedimentary fault tectonics. The majority of the NW–SE and WNW–ESS trending faults were reactivated during the Late Cretaceous–Palaeogene inversion. These faults controlled Late Badenian and Sarmatian subsidence in the PCFB. The highest rate of Late Badenian subsidence was up to 2000 m/Ma in the SE part of the area. In the NW Rzeszów area, the rate of subsidence reached about 1200–1300 m/Ma, and was accompanied by a sedimentation rate of only 1000 m/Ma. Towards the NE margin of the PCFB, the subsidence and sedimentation rate decreased to 100–200 m/Ma. The area of maximum Sarmatian subsidence was the Wielkie Oczy Graben. The total subsidence in this zone varied from 1500 m in its NE part to 3000 m in the SE.

A relatively narrow belt of deformed Upper Badenian–Sarmatian foredeep deposits known as the Zgłobice Unit crops out at the front of the Polish Outer Carpathians (Kotlarczyk 1985; Książkiewicz 1977). This unit consists of several fault-and-fold structures with a maximum width of up to 10 km. To the west of Kraków, there is a narrow zone of folded Miocene deposits (Książkiewicz 1977). In the western part of the Carpathians (Andrychów-Kęty area) several thrust sheets of the Sub-Silesian Unit, covered by lower and upper Badenian deposits, are recorded. These sheets are thrust onto the Badenian deposits in the outer PCFB.

Sedimentary and stratigraphic development

The PCFB is asymmetric and filled with predominantly clastic Miocene sediments (Oszczypko *et al.* 2006) (Fig. 17.25). The outer foredeep is filled with Middle Miocene (Badenian and Sarmatian) marine deposits, which range from a few hundred metres thick in the northern part to 3500 m at the Carpathian Front. The inner foredeep, located beneath the Carpathian nappes, is composed of Lower to Middle Miocene autochthonous deposits. The Lower Miocene strata are mainly of terrestrial origin, whereas the Badenian and Sarmatian sediments are marine. The deposits of the Carpathian Foredeep are underlain by Permo-Mesozoic terrestrial and shelf sediments and locally by Palaeogene deposits. The platform basement, with a Neogene sediment cover, dips southward beneath the Outer Carpathian nappes for at least 50 km (Oszczypko & Ślączka 1985; Oszczypko 1998).

The oldest Miocene deposits were found in a sub-thrust position in a few boreholes in the western inner foredeep, and comprise marine mudstones of late Egerian–Eggenburgian age (Garecka *et al.* 1996; Oszczypko & Oszczypko-Clowes 2003). Higher up in the sequence, slide deposits attain a thickness of up to 370 m. The overlying Karpatian deposits comprise coarse- to medium-grained, polyimictic conglomerates and are followed by a conglomerate–sand–mudstone unit up to 650 m thick. The upper part of the succession may have been deposited in an alluvial fan setting. In the Cieszyn area, the Lower Miocene foreland deposits are overlapped by a 40–90 m thick unit of Late Karpatian transgressive conglomerates.

The Badenian sediments rest directly on the platform basement, except for the SE part of the inner foredeep, where they overlie Lower Miocene deposits. The thickness of the Lower Badenian deposits is variable, ranging from 1000 m in the western inner foredeep (Skawina Formation) to 30–40 m in other parts (Baranów Formation) (Ney *et al.* 1974). Sedimentation of the Skawina Formation (clays and sands) began during the *Praeorbulina glomerosa* Zone in the inner foredeep and during the *Orbulina suturalis* Zone in the outer foredeep (Garecka *et al.* 1996). The mid-Badenian evaporitic horizon (lower part of the NN6 zone; see Peryt 1997; Andreyeva-Grigorovich *et al.* 2003) either overlies these deposits or rests directly upon the platform basement. This evaporitic horizon consists of rock salt, claystones, anhydrite, gypsum and marls. Between Wieliczka and Tarnów, the thickness of the salts reaches 70–110 m (Garlicki 1968) and decreases towards the east to a few dozen metres. The mid-Badenian evaporites are overlain by thick (up to 1.5 km in the central part, and up to 2.5 km in the eastern part) Upper Badenian–Sarmatian sandstones and shales (Gaździcka 1994), which display large-scale clinoforms (Krzywiec 1997). During the early Sarmatian, littoral carbonates and clastics were deposited. After the Sarmatian the final northward thrust moved the Outer Carpathians onto the PCFB.

Carpathians in Western Ukraine (O.Y.A., V.V.A.)

Four main structural units are recognized in the Western Ukraine from west to east (Figs 17.4 & 17.16): the Transcarpathian Basin, the Carpathians, the Carpathian Foredeep, and the Volhyno-Podolian Plate (i.e. the SW margin of the East European Platform). The Mukatchevo and Solotvino sub-basins are part of the Transcarpathian Basin which is separated from the Romanian Transylvanian Basin by the Maramures, Lepus, Rodna and Preluka mountains. To the SW the basin is bordered by the Great Hungarian Plain. The Mukatchevo and Solotvino sub-basins are divided by the volcanic Vyhorlat-Hutin Zone (Voznesensky 1988).

Palaeogeography and tectonic setting

The pre-Neogene basement of the Transcarpathian Basin consists of Palaeozoic to Palaeogene units. In Neogene times, the Solotvino Sub-basin underwent maximum subsidence during the Badenian and up to 2000 m of sediments were deposited, while the deposits in the Mukatchevo Sub-basin are >2500 m thick. Generally, the successions are comparable with those of the Great Hungarian Plain and the East Slovak Basin. The deposits are usually only slightly deformed. Close to the Chop-Vyschkovo Anticline, however, beds dip at up to 90° and displacements of *c.* 20 km are recorded in the northern part of the Solotvino depression (Voznesensky 1988).

The Carpathian Foredeep is situated to the NE of the Carpathians Front and is divided into an inner and an outer zone. The inner zone consists of the Boryslav-Pokuttya and Sambir nappes whose development was caused by the uplift of the Carpathians. The Boryslav-Pokuttya Nappe overthrusts the Sambir Nappe (Kulchytsky 1989). The flexural downbending of the outer zone commenced in the Badenian. The subsequent post-Sarmatian folding resulted in complex deformations in the inner zone, but did not affect the outer zone (Vialov 1986). The Sambir and Boryslav-Pokuttya nappes were detached from their basement and moved *c.* 25–30 km. Sedimentary cover attains a thickness of up to 6000 m.

Sedimentary and stratigraphic development

The lowermost deposits of the **Transcarpathian Basin** are of Eggenburgian age and comprise up to 90 m of sandstones, siltstones and sandy clays (Burkalo Formation, zone NN3; Andreyeva-Grigorovich *et al.* 2002). Karpatian deposits are represented by the Tereshul conglomerates (up to 100 m), comprising unsorted boulder–pebble conglomerates. The conglomerate components are slightly rounded and consist mainly of Jurassic and Cretaceous limestones, sandstones and claystones derived from the Palaeocene (Andreyeva-Grigorovich *et al.* 1997). The overlying Middle Miocene (Badenian) deposits are represented by the Novoselitsa and Tereblia formations, comprising rock salt (up to 270 m) which formed as a result of a marked regression (Vialov 1986; Voznesensky 1988). The marine Novoselitsa Formation and the lower parts of the Tereblia Formation have nannoplankton assemblages of the NN5 zone, while NN6 zone assemblages occur in the upper part of the Tereblia Formation (Andreyeva-Grigorovich *et al.* 2002). The Late Badenian transgression led to the establishment of

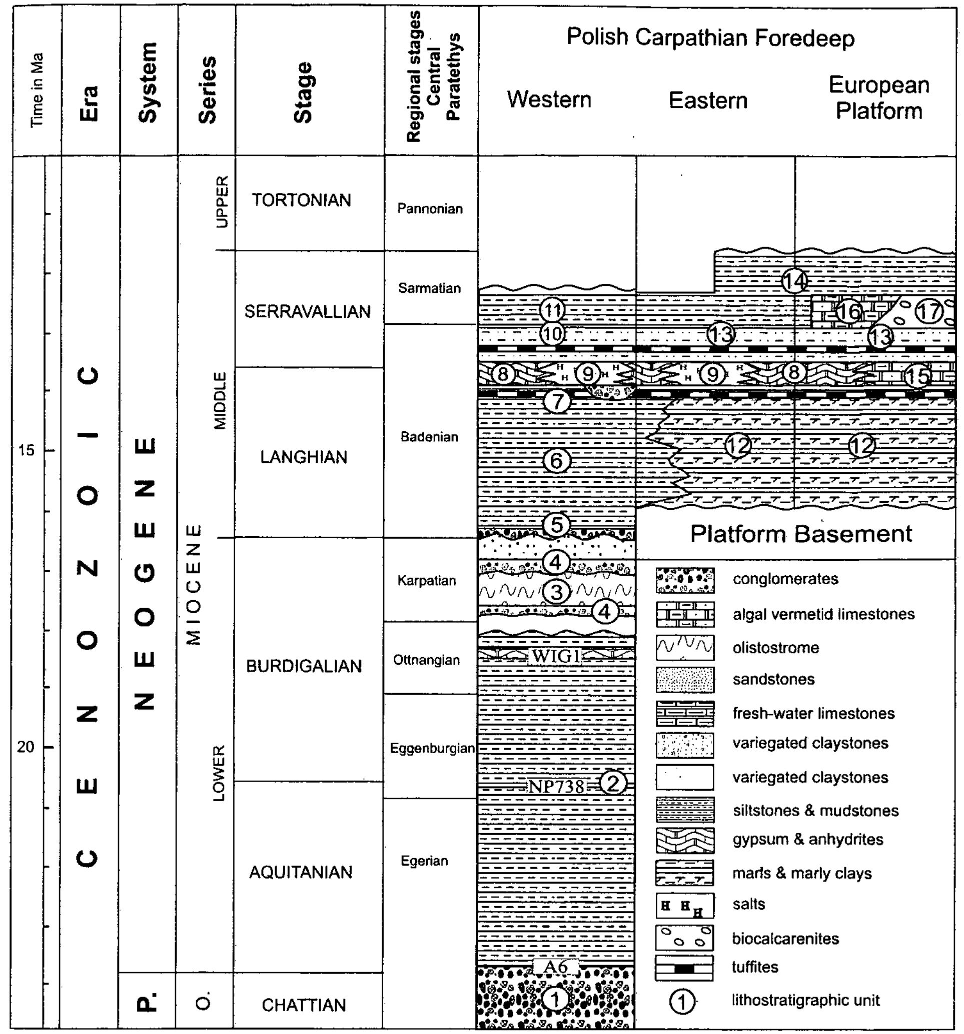


Fig. 17.25. Regional stratigraphic scheme of the Miocene deposits of the Polish Carpathian Foredeep (after Oszczypko 1998; Andreyeva-Grigorovich *et al.* 2003). Lithostratigraphic units: 1, Andrychów Formation; 2, Zebrzydowice Formation; 3, Sucha Formation-flysch olistoplaque; 4, Stryszawa Formation; 5, Debowiec Conglomerate; 6, Skawina Formation; 7, Sypka Góra conglomerates; 8, Krzyżanowice Formation; 9, Wieliczka Formation; 10, Gliwice beds; 11, Kedzierzyn beds; 12, Baranów beds; 13, Chodenice and Grabowiec beds; 14, Krakowiec beds; 15, sulphur-bearing limestones; 16, algal-vermetid reef limestones; 17, Biocalcarenite of the Chmielnik Formation.

marine environments as reflected by the deposits of the Solotvino, Teresva and Baskhev formations. During the Badenian, separation into the Mukatchevo and the Solotvino sub-basins occurred. Sarmatian units are represented by the clays, sands and tuffs of the Dorobratiy, Lukiv and Almash formations, which were mainly deposited in the Mukatchevo Sub-basin in lagoonal and terrestrial environments. Upper Miocene and Pliocene sediments are represented by the Iza, Koshelevo, Ilnitsa and Chop formations, which are restricted to the Mukatchevo Sub-basin. These are mainly volcanic and terrigenous deposits, including important lignite seams (Voznesensky 1988).

In the **Inner Carpathian Foredeep** the Boryslav-Pokuttya and Sambir nappes comprise the Polyanytsa, Vorotyshcha, Stebnyk and Balych formations (Vashchenko & Hnylko 2003) (Fig. 17.8). Calcareous clays, siltstones, sandstones, conglomerates and olistostromes are characteristic of the Polyanytsa and Vorotyshcha formations. The Polyanytsa Formation is absent in the Sambir Nappe and the Vorotyshcha Formation has only a restricted distribution. The Polyanytsa Formation is usually assumed to be of Eggenburgian age (Andreyeva-Grigorovich *et al.* 1997; Veklich *et al.* 1993), although there is some evidence that sedimentation commenced during the Oligocene (Vashchenko & Hnylko 2003). The boundary between the Polyanytsa and Vorotyshcha formations is diachronous and ranges from the Oligocene–Miocene boundary up to the Eggenburgian. Sedimentation of the Vorotyshcha Formation lasted until the Badenian, but its upper boundary is also diachronous (Vashchenko & Hnylko 2003). Sandy-argillaceous sediments characterize the Ottnangian to Lower Badenian Stebnyk Formation in the Boryslav-Pokuttya Nappe. In the Sambir Nappe the basal part of the Stebnyk Formation is cut by thrusting, and passes upwards into the sandy-argillaceous Karpatian to Lower Sarmatian Balych Formation. In the NW of the Sambir Nappe, the Balych Formation is overlain by the Radych Conglomerates (Vashchenko & Hnylko 2003).

The **Outer Carpathian Foredeep** was generated on the margin of the East European Platform, where thick marine, brackish and continental Neogene sediments overlie Proterozoic, Palaeozoic and Mesozoic deposits (Kulchytsky 1989). The Middle Miocene (Badenian) succession begins with the Bogorodchany Formation (*c.* 100 m thick) which is characterized by marly-tuffacous sediments (up to 40 m). Its upper part consists of sandy clays with sandstones rich in pectinids. The overlying Tirass Formation (up to 40 m) is characterized by evaporites which are overlain by the laminated clays and siltstones of the Kosiv Formation (up to 700 m), including an important bed with radiolarians (20 m) (Vialov 1986). The Lower Sarmatian is represented by the Dashava Formation, up to 5000 m thick, comprising clays with sandstones and tuffs.

Volcanism

The Neogene magmatism of the Transcarpathian is characterized by intensive volcanic activity. It was associated with the flexural subsidence of the Transcarpathian region along deep faults, which acted as conduits for the upwelling magmas. The thickest accumulations of tuffaceous and effusive rocks were deposited (1) during the Early Badenian in the central part of Solotvino and the eastern part of the Mukatchevo basins, (2) during the Badenian and Sarmatian in the Beregovo and Vyshkovo areas, and (3) during the Pliocene in the Vyhorlat-Hutin Zone. Badenian volcanism was characterized by acid liparite-dacitic and dacitic lavas and their associated tuffs. From the Sarmatian to the Pliocene, the acid volcanism was replaced by intermediate, and later by alkaline, volcanism, i.e. andesites, andesite-basalts and basalts (Voznesensky 1988).

Southern Alps and Dinarides: overview (D.B.)

The Southern Alps are separated from the Eastern Alps (i.e. Austro-Alpine units) by the Periadriatic Lineament (PAL) (Fig. 17.26). In terms of palaeogeography, both units are considered to represent the northern extension of the Adriatic Microplate which also includes the Po Plain and the adjacent Adriatic Sea. During the Palaeogene, when the Eastern Alps formed an archipelago, the Southern Alps formed their southern continuation. Southern Alpine units in NE Italy include: (a) Palaeozoic basement; (b) Permian–Pliocene sedimentary cover, including continental, platform and basinal deposits; (c) volcanic and subvolcanic deposits, mostly occurring in the Dolomites–Recoaro (Mid-Triassic) and Colli Euganei–Monti Lessini areas (Palaeogene), and (d) a thick succession of Quaternary alluvial deposits filling the Po Plain. The Variscan and Alpine orogenic cycles are recorded in this area. The former involved pre-Permian thrusting, polyphased folding and metamorphic imprinting of the basement whereas the latter is recorded by the Dinaric (Palaeogene) SW-vergent thrusts and by the widely distributed décollement cover, basement nappes and transfer faults of the Neo-Alpine (Neogene) south-vergent episodes, prograding toward the Po Plain foreland.

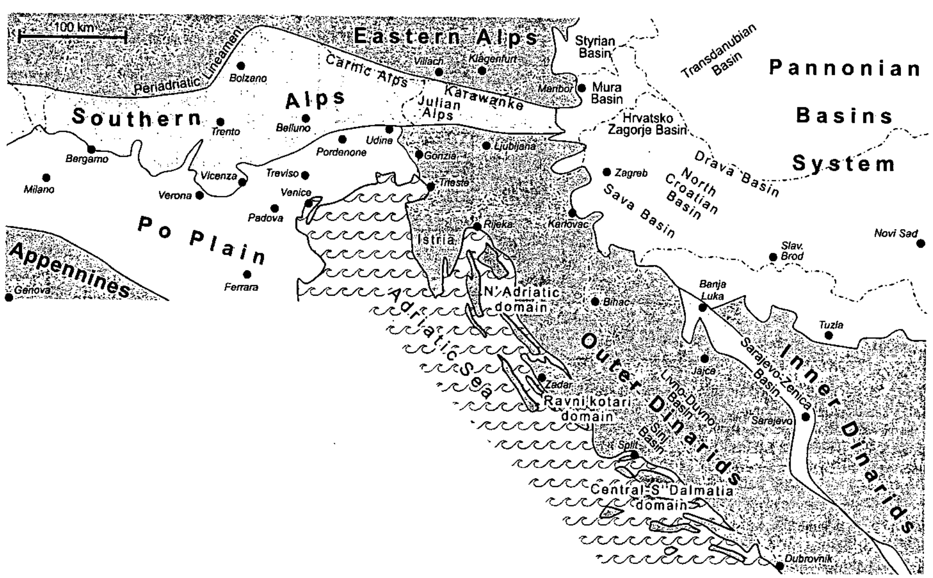


Fig. 17.26. Geology and main sedimentary basins of the Southern Alpine and Dinarides region.

The Friuli Platform and the Julian Basin continue southwards to the Adriatic and Dinaric platforms, and the Bosnian Basin, respectively (Cati *et al.* 1987; Herak 1987). Cati *et al.* (1987) have proposed a scenario with an additional Friuli Basin that subdivided the Friuli Platform into SW (SW of Pordenone) and NE (Udine area and south of Udine) sectors. This basin supposedly continued to the Outer (= External) Dinarides. As a consequence, the SW and NE Friuli platforms should be connected with the Adriatic (present-day Dalmatian zone) and Dinaric (present-day High Karst zone) platforms, respectively.

Southern Alps in Italy: Venetian Pre-Alps (D.B., G.B., P.M., J.H.N.)

The Palaeogene of the Southern Alpine area of NW Italy encompasses parts of the Trentino-Alto Adige and Veneto regions. At the beginning of the Cenozoic, this area was subdivided into two basins which are roughly separated by the present-day north–south alignment of the Brenta River.

Palaeogeography and tectonic setting

The western basin, encompassing the Monte Baldo and Monti Lessini, the Monti Berici and Colli Euganei as well as the Vincentin Pre-Alps, is characterized by widespread Palaeocene, Lower–Middle Eocene and Oligocene volcanic activity and by the deposition of shallow-water carbonates. The Eocene succession comprises series of shallow-marine platform systems and related basinal units. Some areas became emergent in the Early Oligocene, while others were sites of shallow-marine siliciclastic sedimentation, with an increasing carbonate sedimentation rate in Late Oligocene times. At the Oligocene–Miocene boundary, most of this western area was transgressed leading to the development of widespread marine sedimentary systems.

The Cenozoic of the eastern basin (Valsugana, Belluno, Feltre, Vittorio Veneto, Alpago areas) is characterized by the deposition of thick, mostly siliciclastic sediments including Scaglia- (Late Cretaceous to Early Eocene), deep-marine (Palaeogene) and continental (Neogene) successions, as well as a lack of volcanic activity. Basin development during the Chattian to Langhian was influenced by the rapidly rising axial zone of the Alps, which represented an important source of clastic sediments. This is indicated by the presence of clasts derived from the Penninic and Austro-Alpine successions (e.g. Massari *et al.* 1986*a*).

The present-day structure of the Venetian Southern Alps is the result of the superposition of two main Cenozoic compressive phases, designated as the Eo-Alpine and Neo-Alpine tectonic phases. Evidence for an Eo-Alpine (Late Cretaceous–Early Palaeogene) deformation phase, as present in the central-western part of the Southern Alps, is lacking in this area. From the Late Palaeocene onward, the Intraplacca rift became active, resulting in the development of the Cenozoic volcanism of the Venetian area (e.g. Doglioni & Bosellini 1987). In the NE Veneto, the Eo-Alpine phase produced overthrusts and WSW-verging folds which completely deformed the Permo-Cenozoic sedimentary cover.

The Neo-Alpine tectonic phase acted during the Neogene and reached its acme during Late Miocene–Pliocene times. Much of the uplift of the Venetian mountains, and the formation of a series of south-verging overthrusts which progressively migrated towards the Venetian plain, are due to this cycle. Although the Venetian and Trento area constitutes one of the lesser deformed regions of the Southern Alps, the geometry of the Neogene deformation is rather complex (Doglioni & Bosellini 1987).

In the Veneto region, Neo-Alpine tectonic activity is indicated mainly by south-verging overthrusts. Just south of the Dolomites, the overthrusts also involve the crystalline basement (Valsugana Lineament, Pieve di Cadore Lineament; e.g. Doglioni 1984). To the south, in the Venetian Pre-Alps, the sedimentary cover is deformed by folds and overthrusts. The presence of a basal décollement surface, consisting of Upper Permian evaporites, is still under debate. Neogene deformation in the SW part of the Venetian area (i.e. Monti Lessini, Monti Berici, Colli Euganei) is rare. Folds and overthrusts occur in the westernmost part of the Monti Lessini and in the Monte Baldo; these are SE-verging and are linked to the Late Miocene Giudicarie compression (Castellarin 1981, 1984; Grandesso *et al.* 2000). Diachronism in the deformation of the Southern Alps was noted by Castellarin & Vai (1981), who showed that the main shortening event in the western area of the Southern Alps occurred during the Late Oligocene to Early Miocene, as marked by the coarse-grained clastics of the Lower Gonfolite Group, whereby the eastern area experienced the climax of deformation during the Late Miocene and Pliocene.

From the Chattian to Langhian, the Lessini Block can be considered as the foreland basin of the Dinaric Range (Massari *et al.* 1986*a*). An important palaeogeographic and geodynamic change occurred in the Mid-Miocene with the onset of a major phase of extension in the Pannonian Basin and the Inner Dinarides (Horváth 1984). The ongoing northward convergence of Adria and Europe culminated in a major Upper Miocene compressional phase, whose effects were widespread across the Alps and Apennines. Major south-verging thrusting, uplift and southward tectonic progradation occurred in the Venetian area during this phase, concurrent with episodic subsidence and very high sedimentation rates in the Venetian Basin. Strike-slip faults outlined the basin and controlled sedimentation at this stage (Massari *et al.* 1986*b*). A further outward shifting of the depocentres occurred during the Pliocene to Quaternary in the foredeep belt of the Peri-Adriatic ranges (Po, Venetian and Adriatic basins), where subsidence, coupled with compressional deformation, continued. This was associated with the accumulation of a succession, several kilometres thick, in the Po Plain and in the Adriatic (Massari *et al.* 1986*a*).

The final phase of Neogene deformation began in the Pliocene and is still active, as demonstrated by the high seismic activity of some areas. The Aviano overthrust, buried beneath the alluvial deposits of the Upper Venetian Plain, delimits the southern hills of Marostica, Asolo, Montello and Conegliano (Costa *et al.* 1996).

Sedimentary and stratigraphic development

Western Veneto and southern Trentino–Alto Adige. The Jurassic Trento Platform of the Southern Alps is a major structural and palaeogeographic domain of the Adria Plate continental margin (Winterer & Bosellini 1981). It was a region of shallow-marine carbonate accumulation until the uppermost Early Jurassic. The platform was drowned during the Bajocian to Kimmeridgian and formed a deeper shelf characterized by the famous 'Rosso Ammonitico Veronese'. Lower Cretaceous–Lower Eocene sediments ('Maiolica', 'Scaglia rossa') (Fig. 17.27) are of deeper-marine origin (Bosellini 1989). During the complex collision between Europe and the Adria Plate, the former Trento Plateau reacted rigidly and was block-faulted; the submerged Trento Platform was segmented into separate blocks, showing variable uplift, and was punctuated by several volcanic piles. Shallow-marine carbonate sedimentation was initiated on these topographic highs and eventually coalesced to form the Lessini Shelf (Luciani 1988; Bosellini 1989; Zampieri 2000; Beccaro *et al.* 2001).

The distribution of Palaeogene sediments in the Southern Alpine area of NE Italy was strongly influenced by the Alpone-Chiampo Graben (Barbieri *et al.* 1991; Barbieri & Zampieri 1992; Zampieri 1995). This graben was active during the Palaeogene and extends along a NNW–SSE axis in the central-eastern Lessini Shelf. The graben thus divides the Lessini Shelf into two areas which evolved separately, as shown by the different stratigraphic successions from earliest Late Palaeocene onwards (Antonelli *et al.* 1990). The eastern margin of the graben encompasses the eastern Lessini Shelf and the western Monti Berici and the succession comprises planktonic marls and resedimented calcarenites (P. Mietto, unpubl. data). Further to the west, in the western Lessini Shelf, shallow-marine carbonates predominate, including many of the famous fossiliferous localities found in the Monti Lessini (De Zanche & Conterno 1972; Mietto 1975; Beschin *et al.* 1998). The poorly studied western margin encompasses the central-eastern Lessini Shelf up to the San Bonifacio hills.

In the SE Monti Berici an important tectonic threshold, probably coincident with the Riviera Berica area, has been identified; commencing in middle Eocene times, this boundary separated the Berici-Lessini area, comprising mainly shallow-marine deposits, from the Euganean Basin characterized by deep open-water sedimentation (Piccoli *et al.* 1976). This threshold played a fundamental role in the Oligocene palaeogeography. During the Oligocene a new graben formed in the Marostica area (Barbieri *et al.* 1991; Zampieri 1995).

Thin and discontinuous Upper Palaeocene to Lower Eocene marly limestones are widely distributed in NE Italy, e.g. Monte Baldo (Luciani 1989), Lessini Shelf (Garavello & Ungaro 1996), Monti Berici (Massari & Medizza 1973) and Colli Euganei (Massari *et al.* 1976), as well as the southern Altopiano di Asiago areas (Ungaro & Garavello 1989). During the early to middle Eocene, sedimentation was differentiated into pelagic and neritic facies. The latter are widespread across the entire area from middle Eocene (Lutetian) times onwards. A representative area for Ypresian sedimentation is found near the village of

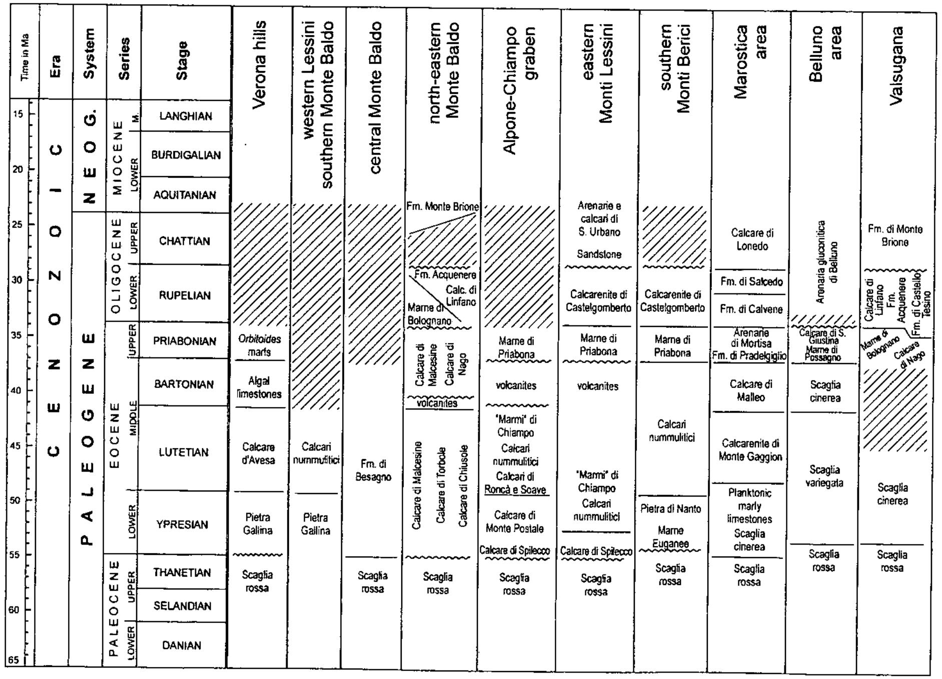


Fig. 17.27. Lithostratigraphy of the western Veneto, southern Trentino–Alto Adige and eastern Venetian areas.

Avesa in the Verona hills. Here, a 70 m thick succession of bedded marls and marly limestones (= Pietra Gallina; Hottinger 1960) is deposited above hardgrounds on top of the Scaglia rossa.

In the western Lessini Mountains (Fumane area), micritic and marly limestones with planktonic foraminifera range in age from Thanetian to Lutetian (Ungaro 1990; Garavello & Ungaro 1996). These limestones are overlain by *c.* 50 m of Middle Eocene nummulitic calcarenites (Formazione di Avesa; Sarti 1980; Ungaro 2001). In the western Monti Lessini area, near Verona and in the southern Monte Baldo area, the Lower Eocene succession consists of a few metres of muddy calcarenites and marls belonging to the Pietra Gallina; these are overlain by nummulitic calcarenites, which were deposited from Ypresian to Bartonian times (De Zanche *et al.* 1977).

In the central Monte Baldo area, Ypresian to Lutetian nummulitic calcarenites with coralline algae and corals (Formazione di Besagno, 50–70 m thick) overlie the Upper Cretaceous–Upper Palaeocene Scaglia rossa, or the lower Ypresian deposits. In the northern Monte Baldo area, Early to Middle Eocene sediments comprise shallow- and deep-water carbonates. These overlie the Scaglia rossa (Bosellini & Luciani 1985), the basinal deposits (Scaglia cinerea, Calcare di Chiusole) as well as the slope and platform carbonates (Calcare di Malcesine, Calcare di Torbole), which represent the initial platform sediments of the area (Bosellini *et al.* 1988; Luciani 1989).

Within the Alpone-Chiampo Graben, micritic and clayey limestones, apparently identical to the underlying Scaglia rossa, have been ascribed to the middle Thanetian (Barbieri & Medizza 1969). Malaroda (1967) grouped all of the sediments from the Upper Cretaceous Scaglia rossa (below) to the Lower to Middle Eocene Calcari nummulitici (above) into the Calcare di Spilecco. This unit consists of basal massive limestones overlying volcaniclastics, followed by larger foraminiferal limestones with brachiopods, crinoids and calcareous algae. These are considered to be Early–Middle Ilerdian (= Ypresian) in age by Papazzoni & Less (2000). Ungaro (2001) ascribed the Calcare di Spilecco cropping out in the Cavalo-Poiano area to the Palaeocene–Early Eocene. Volcanic layers (including basaltic flows, hyaloclastites and volcaniclastics) are often intercalated with these carbonates (Barbieri & Medizza 1969). The Lower–Middle Eocene succession differs markedly from other areas in having two main carbonate bodies which, towards the NE, coalesce into a *c.* 100 m thick succession (Valle dell'Agno; Beccaro *et al.* 2001).

In the Bolca area, the middle Eocene nummulite limestones (Marmi del Chiampo) are interbedded with 100–200 m thick hyaloclastites. Both the calcarenites and the biogenic carbonates of the Monte Postale area, as well as the laminated carbonates of the Monte Bolca area, are laterally equivalent with the lower member of the Marmi del Chiampo. According to Papazzoni (2000*a*), coralline and coral limestones of the Monte Postale area are Early Eocene in age, while the Calcari di Roncà e Soave, Marmi di Chiampo and the Pietra Gallina are middle Eocene in age. Locally, these data are confirmed by calcareous nannoplankton (Beccaro *et al.* 2001). Within the thick volcaniclastic deposits, some lignitic deposits are present, including the famous fossil Lagerstätten of the Purga di Bolca area (Sorbini 1989). At the top of the Calcari di Chiampo, a rich fossiliferous volcaniclastic horizon occurs (San Giovanni Ilarione horizon; Fabiani 1915; Mietto 1975). The thick overlying basalt flows show a rapid environmental change from marine to continental settings, linked with the accumulation of volcanic products which progressively infilled the Alpone-Chiampo Graben. This event occurred in the lowermost Bartonian and is marked by the deposition of paralic facies including the famous Roncà horizon (De Zanche & Conterno 1972). During the Bartonian, subaerial volcanic activity led to the formation of the Monte Calvarina and Monte Faldo, and the original Alpone-Chiampo Graben was replaced by an emergent volcanic ridge on which muddy and lignitic sediments were deposited (Valle dell'Agno; Barbieri *et al.* 1991).

The Bolca area has been studied since the end of the nineteenth century (Oppenheim 1894; Fabiani 1912, 1914, 1915; Schweighauser 1953; Malaroda 1954; Hottinger 1960; Cita & Bolli 1961). Barbieri & Medizza (1969) summarized the geological and stratigraphical setting. The famous Pesciara fish-beds have been known for more than four centuries (Sorbini 1989; Papazzoni 2000*b*).

In the Monti Berici, the Scaglia rossa is followed by a few metres of turbiditic tuffs, tuffaceous marls and marls, which pass upward into marly calcarenites (Pietra di Nanto; Fabiani 1915). This unit is Lutetian in age and grades upward into larger *Nummulites* calcarenites, up to 50–90 m thick, traditionally named the Calcari nummultici (e.g. Ungaro & Bosellini 1965; Mietto 1988; Bassi *et al.* 2000). In the SE part of the Monti Berici, Early Eocene pelagic limestones have been ascribed to the Scaglia rossa (Piccoli *et al.* 1976). In the eastern Monti Berici (Mossano area), outside the Alpone-Chiampo Graben, the Calcari nummulitici extend up the Bartonian–Priabonian boundary (Ungaro 1969; Papazzoni & Sirotti 1995; Bassi *et al.* 2000). In the western Colli Berici, this formation comprises only Lutetian rocks (Fabiani 1915); shallow-marine carbonate sedimentation was interrupted by the extrusion of basalt flows which led to the infilling and emersion of the Alpone-Chiampo Graben at the end of the middle Eocene (Ungaro & Bosellini 1965). At the Bartonian–Priabonian boundary, a general regressive phase has been recognized in the Monti Berici and Monti Lessini areas (e.g. Ungaro 1969; Luciani *et al.* 2002; Bassi *et al.* 2000). In the Colli Euganei, basinal conditions continued through the entire Early Eocene; Palaeocene sediments are absent in the Berici area (Massari *et al.* 1976). In the Colli Euganei, pelagic marls are deposited above the Ypresian Scaglia rossa (Massari *et al.* 1976; Lucchi Garavello 1980) and extend up to the basal Oligocene. Locally, calcarenites and basaltic pillow lavas occur (De Vecchi & Sedea 1995).

In the Marostica area (southern Altopiano di Asiago), the Palaeocene is represented by the Scaglia rossa facies, which is overlain by marly limestones and marls with planktonic foraminifera (Garavello & Ungaro 1982). Lower–Middle Eocene micritic, marly limestones with planktonic foraminifera (Scaglia cinerea) pass upwards into Lutetian–Bartonian larger foraminiferal calcarenites, glauconitic marls and massive calcarenites (Calcareniti di Monte Gaggion; Frascari Ritondale Spano 1969; Garavello & Ungaro 1982; Ungaro & Garavello 1989). The Lower–Middle Eocene deposits of the southern Altopiano di Asiago (Calvene-Pradipaldo area) were described by Garavello & Ungaro (1982), Trevisani (1994) and Papazzoni & Trevisani (2000).

An important transgression is recorded across the entire western Venetian area at the beginning of the Priabonian. In Calvene, a biogenic conglomerate (Conglomerato di Grumale) marks this transgression. This is overlain by massive calcarenites, with local bioherm calcarenites, *c.* 100 m thick (Formazione di Pradelgiglio; Garavello & Ungaro 1982; Ungaro & Garavello 1989; Trevisani 1994, 1997). Marly, arenaceous siltstones, sandstones and conglomerates (10 to 40 m thick) gradually overlie these calcarenites and represent the Priabonian Arenarie di Mortisa (Papazzoni & Trevisani 2000). This formation is overlain by the Lower Oligocene Formazione di Calvene. The Priabonian transgression is also evident on the elevated ridge of the former Alpone-Chiampo Graben, where the historical holostratotype of the Priabonian is located (Passo di Priabona, near Monte di Malo) (see Mietto 2000 for overview). The Priabonian succession transgresses altered Middle Eocene basalts and is characterized by the basal Conglomerato del Boro (Fabiani 1915; Antonelli *et al.* 1990).

In the western Monti Berici this conglomerate is not so evident and is known as the Conglomerato a *Cerithium diaboli* (Fabiani 1915; Ungaro & Bosellini 1965; Mietto 1988). The Marne di Priabona has a thickness of *c.* 90 m in the type area and can attain 170 m in the Mossano area (eastern Monti Berici; Bassi *et al.* 2000). The top of the succession is locally characterized by a marly horizon rich in bryozoans (e.g. Ungaro 1978; Marne di Brendola in Broglio Loriga 1968). This formation appears to decrease in thickness towards the centre of the elevated ridge of the former Alpone-Chiampo Graben; it is absent in the Valle dell'Agno, where the overlying Oligocene Calcareniti di Castelgomberto is transgressive over Middle Eocene volcanic units (Barbieri *et al.* 1980; Mietto 1992).

The basaltic units of the Monte Baldo area were subaerially exposed in the uppermost Bartonian. The overlying Middle to Upper Eocene succession comprises platform (Calcare di Nago), slope (Calcare di Malcesine) and basin (Scaglia cinerea) deposits (e.g. Castellarin & Cita 1969; Luciani 1989; Bassi 1998). The Calcare di Nago was subaerially exposed at the Eocene–Oligocene boundary (Luciani 1989).

In the southern Altopiano di Asiago, the Formazione di Pradelgiglio was deposited during the Priabonian between the Astico and Brenta rivers (Frascari Ritondale Spano 1970; Trevisani 1994). This formation consists of siliciclastic sediments which pass upwards into larger foraminiferal calcarenites with coralline algae and benthic macrofossils (Papazzoni & Trevisani 2000). The upper Priabonian Arenarie di Mortisa, overlying the Formazione di Pradelgiglio, consists of bioturbated marls with larger foraminifera, siltstones and sandstones, and foraminiferal calcarenites (Trevisani 1993).

The Oligocene is missing in the Monti Lessini area of the Verona Province (Antonelli *et al.* 1990). The presence of fossilized palaeokarstic cavities, filled by yellow soils (terra gialla di Verona), suggests intense erosion subsequent to emersion (Corrà 1977). Near Cavalo (North Verona), Oligocene beach sandstones transgress over Lower Eocene deposits (Castellarin & Farabegoli 1974). The Oligocene is well represented along the eastern Monti Lessini margin and in the Colli Berici (from Lumignano to Villaga), where an important shallow-water carbonate platform, characterized partly by hermatypic corals and patch-reefs developed (Geister & Ungaro 1977; Ungaro 1978; Frost 1981). These shallow-marine carbonates represent the Calcarenite di Castelgomberto (Bosellini 1967; Bosellini & Trevisani 1992) in which volcanodetrital lenses, rich in fossils, are locally present (e.g. Accorsi Benini 1974). On the opposite side of the Monti Berici, towards the Colli Euganei, the Marne Euganee were deposited in a pelagic basin (Piccoli *et al.* 1976). The temporary emersion of volcanic islands is documented by the formation of lignitic deposits, the most famous of which in Monteviale and Gazzo di Zovencedo (Monti Berici) contain an extraordinary land mammal fauna (e.g. Dal Piaz 1937; Bagnoli *et al.* 1997).

At the beginning of the Late Oligocene, the Calcareniti di Castelgomberto carbonate platform became emergent (Frost 1981; Gianolla *et al.* 1992); this event is documented by extreme palaeokarst formation (Bartolomei 1958; Mietto & Zampieri 1989; Dal Molin *et al.* 2001). In the Late Oligocene, following weak, local volcanic activity, this area emerged, as documented by the clays produced by the subaerial alteration of the volcanic sediments present at the top of the Calcareniti di Castelgomberto (Sovizzo; Bosellini 1964). In the northern Monti Berici (Valmarana, Col del Bosco), the Chattian succession is represented by sandstones and calcareous sandstones rich in larger foraminifera, which are overlain by coralline algal rudstones (Ungaro 1978; Bassi *et al.* 2007). Chattian sandstones and coralline algal rudstones are also present in the Monti Lessini area and are known as the Arenarie e calcari di S. Urbano (Bosellini 1967; Bassi *et al.* 2007); these are overlain by Lower Miocene marine marls (Marne argillose del M. Costi).

In the Monte Baldo area, the Early Oligocene is represented by shallow-water carbonates (Formazione Acquenere), siliciclastic carbonates (Calcare di Linfano) and marls (Marne di Bolognano) (Bosellini *et al.* 1988; Luciani 1989). Upper Oligocene sediments crop out in the Monte Brione and Monte Moscal areas; this succession is represented by conglomerates and larger foraminiferal and coralline algal calcarenites (Calcare di Incaffi; Luciani 1989) and by deep-water mixed siliciclastic-carbonate units (Formazione di M. Brione, Oligo-Miocene in age; Luciani & Silvestrini 1996).

In the southern Altopiano di Asiago area, the Oligocene units comprise siliciclastic sediments passing into paralic facies. The Arenarie di Mortisa formation is overlain by thick lower Oligocene siltstones and sandstones, often with conglomerates (Formazione di Calvene; Papazzoni & Trevisani 2000). The succession ends with the Formazione di Salcedo, a rhythmic alternation of volcanic deposits and fossiliferous units such as the Arenarie di Sangonini and the Marne di Chiavon. This area emerged in the uppermost Early Oligocene (Principi 1926; Frascari Ritondale Spano & Bassani 1973).

During the Aquitanian a new transgression led to the re-establishment of marine conditions in the western Venetian area. The Miocene successions are represented by sandstones and calcareous sandstones directly overlying the Marne di Priabona (Monti Lessini in the Verona Province), Oligocene limestones (Monti Lessini in the Vicenza Province), and submarine volcaniclastic sediments (Monti Berici; Ungaro & Bosellini 1965). In the Marostica area, the Aquitanian Calcare di Lonedo transgressed onto the Formazione di Salcedo, which comprises coralline calcarenites (Frascari Ritondale Spano 1969). The Burdigalian Molasse of Schio overlies the Calcare di Lonedo. In the Vicenza area, the Chattian Arenaria di S. Urbano is overlain by the Miocene Marne argillose del M. Costi, which are only a few metres thick (Bosellini & Dal Cin 1966; Bassi *et al.* 2007).

Eastern Venetian part of the Brenta River. During the Palaeocene–Early Eocene, pelagic units were deposited in the Belluno (Scaglia cinerea, Marna della Vena d'Oro) and in the western Treviso (Scaglia variegata) areas (Fig. 17.27). In the latter area, as in the Feltre area, sedimentation continued until the Middle Eocene (Scaglia cinerea). During the Late Eocene, a regressive succession evolved with the deposition of the Marna di Possagno and the Calcare di Santa Giustina units. In the SE Monte Grappa area (Treviso Province), the Possagno section (parastratotype of the Priabonian) is the thickest Priabonian succession in the Southern Alps. The section is *c.* 700 m thick and consists mainly of clay marls with planktonic and small benthic foraminifera (lower-middle Priabonian Marna di Possagno formation; Cita 1975; Grünig & Herb 1980). The Marna di Possagno is overlain by the upper Priabonian Calcare di Santa Giustina unit, consisting of larger foraminiferal calcarenites

(Braga 1972; Trevisani 2000).

The Belluno Flysch is a sedimentary body >1000 m thick, located between the Alpago and the Feltre areas, as well as in the Venetian foreland basin between Vittorio Veneto and Segusino. In the Alpago and Belluno areas, this unit contains only lower Eocene sediments, in the Feltre area it extends into the middle Eocene, and in the Follina area up to the late Eocene. North of Feltre, the middle-upper Eocene succession is characterized by alternations of biogenic calcarenites, marls and sandstones. The Marna di Possagno unit and the upper part of the Belluno Flysch are interpreted as slope deposits interfingering with the distal facies of the Belluno Flysch (e.g. Costa *et al.* 1996; Stefani & Grandesso 1991).

Across the eastern Venetian area, the Eocene deposits are unconformably overlain by the Chattian Molasse. The unconformity is the result of an important hiatus which becomes older towards the east (Middle Eocene–Early Oligocene in the Alpago and Belluno areas) and younger towards the west and south (Early Oligocene in the Feltre, Possagno and Follina areas). The Southern Alpine Molasse ranges from the Chattian to Recent (Massari *et al.* 1986*a*; Costa *et al.* 1996), and represents parts of the infilling of the Venetian Basin, restricted to the east by the Dinaride Mountains, and to the west by the Schio-Vicenza lineament. The Chattian–Messinian sediments form a clastic body with a maximum thickness of >4000 m at the Venetian foreland basin–plain boundary. The molasse consists mainly of sandstones, siltstones, marls and conglomerates. These are inner-platform carbonate deposits that are occasionally overlain by more distal sediments (Marna di Bolago, Marna di Monfumo, Marna di Tarzo; Massari *et al.* 1986*a*). Sedimentation commenced with shallow-marine deposits including a tide- and wave-dominated delta system, and a mixed siliciclastic-carbonate facies (Massari *et al.* 1986*a*, *b*). The Arenaria glauconitica di Belluno, up to 50 m thick, consists of fossiliferous sandstones rich in glauconite and represents the transgressive basal layer of the molasse succession. It is heteropic with the Conglomerato di M. Parei and the Calcarenite dell'Alpago (Massari *et al.* 1986*b*). The Conglomerato di M. Parei (Keim & Stingl 2000), which crops out in the Ampezzo area, contains Chattian to Aquitanian fossils in the matrix and unconformably overlies Mesozoic units. The Chattian Calcarenite dell'Alpago (Costa *et al.* 1996) comprises up to 50 m thick glauconitic calcarenites with larger foraminifera, rhodoliths and bryozoans. This is restricted to the Alpago area (Massari *et al.* 1986*b*). The Calcarenite di Castelcucco (Scudeler Baccelle & Reato 1988; Bassi *et al.* 2007) crops out in the Castelcucco and Vittorio Veneto area and is represented by larger foraminiferal and coralline algal calcarenites alternating with thin marly beds. This unit overlies the Siltite di Bastia and is disconformably overlain by the Aquitanian Siltite di Casoni (Massari *et al.* 1986*b*).

The northern margin of the Lessini Shelf is located in the Valsugana area (Borgo Valsugana to Pieve Tesino). The early to lowermost middle Eocene is represented by deep-water sediments (Scaglia cinerea) with planktonic foraminifera (Fuganti *et al.* 1965). The Late Eocene consists of shallow-marine foraminiferal and coralline algal calcarenites of the Calcare di Nago unit and deep-water sediments (Marne di Bolognano, Formazione di Castello Tesino; Luciani & Trevisani 1992). The Oligocene is represented by the Formazione Acquenere (marls and marly siltstones with *Nummulites*), the Calcare di Linfano (larger foraminiferal calcarenites: Schiavinotto 1978), the Formazione di Castello Tesino and the Formazione di M. Brione. The Priabonian to Rupelian deposits are generally characterized by a large amount of siliciclastics and a reduced thickness (Trevisani 1997).

The Neogene sediments of the Vittorio Veneto–Belluno area are mainly siliciclastics. Carbonate deposits with thick rhodolith accumulations are represented by the upper Chattian Calcarenite di Castelcucco unit in the Vittorio Veneto area, and by the coeval Calcarenite dell'Alpago unit in the Belluno–Alpago area. The Siltite di Bastia unit is overlain by the Calcarenite di Castelcucco and overlies the Calcarenite dell'Alpago. In the Alpago area the Siltite di Bastia is up to 250 m thick and decreases in thickness towards Feltre; in the Venetian foreland basin area it is a few metres thick (Ghibaudo *et al.* 1996). Between the Feltre and Alpago areas, two sandstone units with glauconite (Arenaria di Orzes and Arenaria di Libano) overlie the Siltite di Bastia, separated by a silty unit (Siltite di Casoni). Around Belluno these three formations are >100 m thick. The Arenaria di Orzes has been interpreted as an estuary shoal deposit, while the Siltite di Casoni and the Arenaria di Libano represent prodelta and delta-front facies (Massari *et al.* 1986*a*; Costa *et al.* 1996).

The Burdigalian is represented by a transgressive–regressive succession related to a eustatic event affecting the outer carbonate platform. The Marna di Bolago consists of marls and ranges in thickness from 100 m (Feltre) to 200 m (Venetian foreland basin area). The overlying Arenaria di S. Gregorio (up to 60 m of sandstones) represents a rapid transition from outer- to inner-platform environments with a fluvial influence (Costa *et al.* 1996).

The Langhian commenced with a rapid transgression, which followed the abrupt deepening of the Venetian Basin to bathyal conditions. The succession begins with the Marna di Monfumo which comprises a basal layer of fossil-rich glauconitic sandstones overlain by marls containing small bivalves (20–45 m thick). The Arenaria di Monte Baldo overlies the Marna di Monfumo and consists of coarse glauconitic-fossiliferous sandstones, and biocalcarenites alternating with grey marls. This unit represents shelf deposits with sand ridges (Massari *et al.* 1986*a*, *b*). The Arenaria di M. Baldo crops out in the Venetian foreland basin area (Vittorio Veneto, Crespano). Its thickness ranges from 50 to 300 m, with a maximum near Vittorio Veneto.

From the Serravallian to the Messinian, several important changes occurred in the Venetian basin: the axis of the foredeep shifted and the basin was incorporated into the South Alpine deformational system. In the initial stage, accelerated subsidence led to the deposition of epibathyal marls (Marna di Tarzo) overlying the Arenaria di M. Baldo, followed by organic-rich lower Tortonian mudstones. The subsequent stage was characterized by rapid slope progradation (middle Tortonian). Localized conglomerate bodies (Conglomerato di M. Piai) within this succession represent mass-flow deposits probably funnelled along a structural depression (Massari *et al.* 1986*a*). In the late Tortonian, subsidence slowed down and a stack of vertically aggrading fan-delta sequences (Arenaria di Vittorio Veneto) were deposited on the shelf created by the previous progradational episode (Massari *et al.* 1986*a*). The architecture of the Messinian alluvial system was largely controlled by tectonics. Palaeocurrent indicators and the composition of Tortonian–Messinian sandstones and conglomerates suggest a source area within the eastern South Alpine domain. The Messinian Conglomerato del Montello represents coarse fan-deltas passing upward into alluvial deposits (up to 200 m thick). Blue sandy clays, early–middle Pliocene in age, deposited above the Conglomerato del Montello crop out near Cornuda and represent a neritic setting (Massari *et al.* 1993).

Volcanism

Palaeogene magmatism within the Eastern Alps is variable in time and space. This magmatism can be interpreted in terms of the changing geodynamic framework that is reflected in the different evolution of the related mantle sources (Wilson & Bianchini 1999). Two main tectonomagmatic associations can be recognized: orogenic (subduction-related) suites and anorogenic (intraplate) suites.

Magmatic rocks characterized by orogenic affinities, including both mafic and felsic dykes as well as granitoid intrusions, occurred mainly between 42 and 25 Ma (with a climax between 33 and 29 Ma) along the Insubric-Periadriatic Lineament. Major element composition of the basic rocks indicates a spectrum of magma compositions ranging from calc-alkaline and high-K calc-alkaline to shoshonitic types (Beccaluva *et al.* 1979, 1983; von Blanckenburg & Davies 1995; Macera *et al.* 2002). The most primitive mantle compositions appear to be mantle-derived melts relatively unaffected by shallow-level crustal contamination. Their trace element distribution displays enrichments in LILE (large ion lithophile elements, e.g. Cs, Rb, K) and depletion in HFSE (high field strength elements, e.g. Nb, Ta, Ti) as typically observed in magmas from active continental margins. These geochemical features, along with Sr–Nd isotopic evidence, indicate that magmatism was induced by partial melting of lithospheric mantle domains intensely metasomatized by subduction-related fluids/melts. The significant presence of shoshonites suggests that continental crustal components were also involved in subduction.

To the south, anorogenic volcanic rocks characterize the Veneto Volcanic Province (VVP). This is Late Palaeocene to Late Oligocene in age and is represented by a series of eruptive centres orientated NNW–SSE. Magma generation appears to have been triggered by decompressional effects related to extensional deformation which affected the South Alpine Foreland in response to the general north–south compression during the Alpine Orogeny (Beccaluva *et al.* 2003, 2005). VVP lavas are mainly basic and comprise a wide compositional spectrum of mantle-derived melts including (mela) M-nephelinites, basanites, alkali-basalts and tholeiites (De Vecchi & Sedea 1995), as typically observed in low-volcanicity rifts. The relative abundance of silica-undersaturated and silica-oversaturated products varies regionally, with more abundant nephelinites and basanites to the west (Val d'Adige and western Monti Lessini), and predominantly alkali-basalts, transitional basalts and tholeiites to the east (eastern Monti Lessini and Marostica area). The sodic character ($Na_2O/K_2O > 1$), and the incompatible element distribution (characterized by negative anomalies in Cs, Rb, K, and positive anomalies in Nb–Ta) of VVP volcanics show similarities with intraplate magmas. Recent studies on spinel-peridotites (mantle fragments) entrained as xenoliths in VVP basanites and nephelinites indicated that: (a) the local lithospheric mantle was extensively metasomatized by alkali-silicate melts; and (b) the area was characterized by an anomalously hot geotherm (Siena & Coltorti 1989, 1993; Beccaluva *et al.* 2001). These factors, coupled with the adiabatic decompression that was induced by extension, triggered magma-genesis and, thus, the observed volcanism (Beccaluva *et al.* 2005). Sr–Nd–Pb analyses of the volcanics, integrated with studies of the entrained mantle xenoliths, indicate that both lithospheric and sublithospheric (possibly plume-related) mantle sources were involved in their generation (Macera *et al.* 2003; Beccaluva *et al.* 2005).

Southern Alps in Italy: Friuli Pre-Alps and Karst (N.P., G.T.)

The Palaeogene deposits of the Friuli-Venezia-Giulia region have been extensively studied (e.g. Stache 1889; Dainelli 1915; Martinis 1962; Auboin 1963; Bignot 1972; Cousin 1981; Cucchi *et al.* 1987). A short synthesis was given by Venturini & Tunis (2002). The Palaeogene scenario, inherited from the Mesozoic, consisted of the Friuli Platform, which bordered the Belluno Basin to the NW and the Julian Basin towards the NE (Venturini & Tunis 2002). Palaeogene deposits of the Friuli-Venezia-Giulia were part of the Friuli Platform and the Julian Basin. From a geographic point of view, the Friuli Platform corresponds to the present-day southern Pre-Alps and Karst, and the Julian Basin can be related to the Julian Pre-Alps. The Friuli Platform and the Julian Basin can be traced southwards to the Adriatic and Dinaric Platforms, and the Bosnian Basin, respectively (Cati *et al.* 1987; Herak 1989). Moreover, Cati *et al.* (1987) proposed a new scenario with an additional Friuli Basin that subdivided the Friuli Platform into SW (SW of Pordenone) and NE (Udine area and south of Udine) sectors. This new basin would then continue to the Outer Dinarides. As a consequence, the SW and NE Friuli platforms may be connected with the Adriatic (present-day Dalmatian Zone) and Dinaric (present-day High Karst Zone) platforms, respectively.

Tectonic setting

The tectonic setting of the Palaeogene of the Friuli-Venezia-Giulia regions comprises a series of NW–SE overthrusts, which verge southwards (Castellarin & Vai 2002). Compression led to the establishement of a foreland basin which migrated from Slovenia towards the Veneto (Doglioni & Bosellini 1987). Drowning of parts of the Friuli Platform proceeded from east to west. This tectonic phase is mainly Early Eocene in age (Carulli *et al.* 1982; Poli & Zanferrari 1995) and can be linked to compressive activity during the Meso-Alpine collisional phase (Castellarin & Vai 2002). Following Oligocene extension, a neo-Alpine compressive phase occurred (Chattian-Burdigalian), as demonstrated by the formation of NW–SE orientated thrusts (Castellarin & Vai 2002; Ponton & Venturini 2002). Subsequently, two further compressive phases occurred during Langhian–Tortonian and ?Messinian–Pliocene–Quaternary times, as demonstrated by south-orientated thrusts.

Sedimentary and stratigraphic development

The Palaeogene carbonate platform succession commences with Palaeocene sediments, which mostly overlie Late Cretaceous units; Maastrichtian is proven for some localities of the east Trieste Karst (Pugliese *et al.* 1995) and the south and north Gorizia Karst (Tentor *et al.* 1994). Maastricthian inner lagoonal sediments comprise the last rudist genera *Bournonia* and *Apricardia* as well as foraminifers (Caffau *et al.* 1998). The latest Cretaceous to early Danian succession contains several stratigraphic hiatuses, often associated with emergence and the development of peritidal cycles, such as in the eastern Trieste Karst (Pugliese *et al.* 2000). One of the basal peritidal cycles at Padriciano records the K/Pg boundary (Pugliese & Drobne 1995; Pugliese *et al.* 1995, 2000), which is also present elsewhere in the Slovenian Karst area (Drobne *et al.* 1988*a*; Ogorelec *et al.* 1995, 2001). The boundary is documented by palaeontological data (Pugliese *et al.* 1995, 2000), palaeomagnetism (Martón *et al.* 1995*a*), an iridium anomaly (Hansen *et al.* 1995), and a negative shift in $\delta^{13}C$ (Ogorelec *et al.* 1995).

After the K/Pg boundary crisis, pioneer biota (small foraminifers, thin shelled ostracods and gastropods) appeared in peritidal environments during the early Danian (Shallow Benthic Zone SBZ1 of Serra-Kiel *et al.* 1998). Characean algae indicate freshwater influence. The peritidal cycles are overlain by inner lagoonal limestones, which are Danian (SBZ1) to Thanetian

(SBZ3) in age. Alveolinid and nummulitid foraminifers, corals, corallinacean algae and sea-urchins appeared during SBZ4 (Pugliese *et al.* 1995, 2000). Sedimentation continues until the late Cuisian (SBZ12) with shallow-marine sediments containing larger foraminifera (Gozzi 2003). These Palaeogene deposits are mainly NW–SE directed (Drobne *et al.* 2000*b*) and preliminary larger foraminifera data suggest two main belts in the Trieste Karst (Drobne 2003*b*). The first one (zone 2 of Drobne 2003*b*) includes the localities Padriciano, Opicina and Duino. It is characterized by Danian to early Ypresian (SBZ1–9) carbonate ramp deposits, partly overlain by late Ypresian deep-marine sediments (Castellarin & Zucchi 1966; Pugliese *et al.* 1995; Gozzi 2003). The area of Colle of Medea (Udine) may also be part of this belt (Barattolo 1998). The second belt (zone 3 of Drobne 2003*b*) includes the Rosandra valley and is characterized by Danian to late Cuisian (SBZ1–12) carbonate ramp deposits overlain by deep-marine sediments.

Palaeogene terrigenous sedimentation took place in the Julian Basin, a narrow, elongate basin, which today is part of the SE Alps (eastern Friuli–western Slovenia). The basin had an internal margin in the north, which was subject to compression, and an external margin in the south, represented by the Friuli Carbonate Platform. The basin was characterized by mixed carbonate/siliciclastic sedimentation from late Campanian to Early Eocene times. Sediment distribution patterns are complex due to a combination of tectonics, relative sea-level changes, and subsidence (Tunis & Venturini 1992). During the main phase of turbidite deposition (Maastrichtian–early Eocene), >4000 m of sediments were deposited.

Most of the Maastrichtian units are interpreted as slope-apron environments located close to the margin of the Friuli Platform, while the Palaeocene and Eocene units reflect a position close to the outer side of the basin (Flysch di Calla, Flysch di Masarolis and Flysch di Grivò), followed by basin-plain environments and a deltaic system (Flysch di Cormons). The Flysch di Calla (early–middle Palaeocene) consists of reddish marls interbedded with thin sandstones and is a significant unit in eastern Friuli and western Slovenia (Pirini Radrizzani *et al.* 1986; Dolenec & Pavsic 1995). The Flysch di Masarolis (middle–upper Palaeocene) is mainly represented by thin siliciclastic turbidites, while the Flysch di Grivò (Late Palaeocene–Early Eocene) contains several carbonate megabreccias derived from the resedimentation of shallow-marine carbonates (Friuli Platform) as well as intercalated siliciclastic–carbonate turbidites. These extensive carbonate debris accumulations are explained by catastrophic resedimentation events in the deep-water basin. Some megabeds have a strike of more than 70 km and individual thicknesses of up to 200–260 m (Vernasso Megabed). Olistoliths can reach a length of several hundred metres.

The major megabreccia units were generated by the fault-controlled failure of the Friuli Platform margin, which became seismically active during the Late Palaeocene–Early Eocene. The Flysch di Cormons (Early Eocene–Mid Eocene) is characterized by basin-plain turbidites. Sea-level changes controlled basinal sedimentation from the latest Ypresian to the earliest Lutetian (Venturini & Tunis 1991). During the early Lutetian, a rapid progradation of prodelta, delta-front, and deltaic-plain deposits occurred in response to tectonic uplift. The prodelta deposits contain a rich and diverse macrofauna (corals, molluscs, echinids, alveolinids, nummulitids, etc.) (Maddaleni & Tunis 1993; Hottinger 1960). The corals formed patch-reefs in front of the advancing deltaic system. Together with the Slovensko Primorje, the Trieste area belongs to the Trieste-Koper Syncline. According to Pavšič & Pekmann (1996), the Trieste Flysch is characterized by interbedded siliciclastic sandstones and marlstones. The sedimentological features (probably basin plain turbidites) and the ichnofacies suggest a deep-marine environment (Tunis *et al.* 2002). In agreement with Pavšič & Pekmann (1996) and Tunis *et al.* (2002), the turbidity current flow was parallel to the WNW–ESE striking basin axis during the Mid-Eocene.

Slovenian Tethys basins (K.D., B.O., J.P., R.P.)

Slovenia is situated at the contact of the Southern and Eastern Alps with the Pannonian Basin, the Dinarides and the Adriatic Sea (Figs 17.2, 17.26 & 17.29). Palaeogene sediments in SW Slovenia are delimited to the NE by the high plateaus of Trnovski gozd, Nanos and Snežnik, as well as Gorski Kotar in Croatia, and to the south by the Čičarija hills in the Croatian part of Istria. To the west, it passes along a narrow belt across the Trieste-Komen Plateau to the Soča valley, and further along it towards the north. In a wide arc the boundary encloses the hills of Goriška Brda. Further northward the boundary rises on the ridges between the Soča and Nadiža rivers. These deposits are heavily karstified and contain numerous caves, dolinas as well as several valleys with outcrops of deep-marine sediments. Palaeogene belts associated with the predominant Dinaric direction (NW–SE) continue to the east and south to Croatia, and to the west to Friuli. In the SE part of Slovenia, Palaeocene beds are exposed only as small erosional relics, e.g. between Ribnica and Novo mesto.

Tectonic setting and palaeogeography

Slovenia belongs to four large tectonic units: the Eastern Alps and the Southern Alps separated by the Peri-Adriatic Lineament; the Pannonian Basin in the east; and the Outer Dinarides extending to the south (Premru 1980, 1982, 2005; Placer 1981, 1999*a*). The Dinarides border the Adriatic Platform along the Dinaric Thrust. This series of overthrusts verges from the north towards the SE and includes Trnovski gozd, Nanos with Hrušica, Snežnik, and in Croatia, Gorski Kotar and Velebit (Rakovec 1956; Poljak 2000; Bigi *et al.* 1990; Blašković 2000). The Čičarija Zone is presumably a deformed intermediate zone between the Dinaric Thrust and the Istria autochthon (Placer & Vrabec 2004). The tectonic relationship between the Dinarides and the Southern Alps was described by Herak (1985, 1986, 1987, 1999). He subdivided the region into three tectogenetic units, namely Dinaricum, Adriaticum and Epiadriaticum, the first two of which represent carbonate platforms and the third one an intermediate pelagic zone. This subdivision is widely used (Blašković 2000; Biondić *et al.* 1997; Carulli *et al.* 1990; Drobne & Trutin 1997; Mioč 2003; Drobne *et al.* 2000*a*, *b*), but it is still under discussion. Other studies have suggested a separation of the Karst Dinarides, or a single Adriatic Carbonate Platform (AdCP) instead of the Dinaric Platform (Tišljar *et al.* 2002; Vlahović *et al.* 2002; Dragičević & Velić 2002). During the Cenozoic, most parts of the Adriatic and Dinaric platforms were emerged, and only small areas of the Adriatic Platform remained below sea level. This occurred during the Palaeocene in Slovenia, the Trieste region (Pugliese *et al.* 1995) and Hercegovina. Sedimentation continued, with interruptions, up to the end of the Middle Eocene. The development of the Palaeogene in SW Slovenia commenced either in the central part of the Western Tethys (after Golonka 2004) or, as proposed by Channell *et al.* (1979), as an African promontory along a submarine belt. This structural unit, known as the Apulian Plate or Adriatic (Micro)-Plate consisted of large carbonate platforms: in the west the

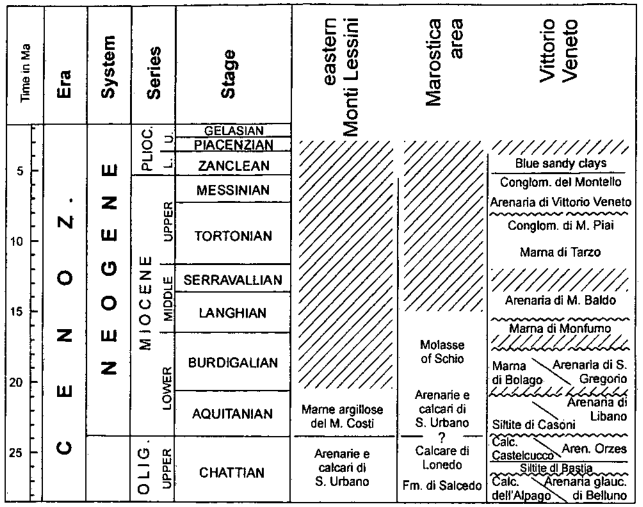


Fig. 17.28. Lithostratigraphy of selected localities in the western and eastern Venetian areas.

Apennines with Apulia, in the east the Adriatic-Dinaric Carbonate Platform (or Adriatic Carbonate Platform), alternating with deep-marine basins (e.g. Belluno, Friuli basin; Cati *et al.* 1987).

Palaeomagnetism

At the Dolenja Vas section, which crosses the K/Pg boundary, reversed polarity has been observed in the Polarity Chron C29R, with a rotation angle of 28° in a clockwise direction (Dolenec *et al.* 1995; Marton in Drobne *et al.* 1996*a*, *b*). Investigations from this zone both towards the north into the Vipava valley and south to Savudrija, indicate rotation in a counterclockwise (CCW) direction of 30°. The polarity chrons between C29R and C21 have been described by Marton *et al.* (1995*a*, 2000*a*, *b*). New interpretations indicate that the imbricated Čičarija Zone and Istria have rotated by 30° CCW in relation to Africa and Europe during the Cenozoic. The most recent data for the last rotation is late Miocene to early Pliocene (Marton *et al.* 2003*a*, *b*). These dates may be valid for the entire AdCP as well as for the Adriatic Plate (cf. Mantovani *et al.* 1990).

Sedimentary and stratigraphic development

Palaeogene rocks in SW and southern Slovenia occur in three sedimentary units that co-existed parallel to each other: the deep trench, the deep-marine (flysch) basin, and the carbonate platform (Fig. 17.30). The succession begins with the Liburnian Formation, separated into the Maastrichtian Lower Miliolid Beds, the Danian to Selandian Kozina Beds, and the Thanetian Upper Miliolid Beds. The Liburnian Formation passes into the Ypresian to mid-Lutetian Alveolinid-Nummulitid Limestone (Alveolinsko Numulitni Apnenec in Slovenian literature). The lithostratigraphic terminology is inconsistent (e.g. Delvalle & Buser 1990; Jurkovšek *et al.* 1996*a*, *b*; Košir 1997, 2004; Ćosović *et al.* 2004*a*, *b*). Locally, the Cretaceous part of the Liburnian Formation is known as the Vreme Beds, and the Upper Palaeocene part as the Trstelj Beds (Pavlovec 1963; Pleničar & Pavlovec 1981). Other terms used include Slivje Beds for the first appearance of larger foraminifers in the Upper Miliolid Beds (Delvalle & Buser 1990), or Operculinid Limestone for nummulitid accumulations at the base of the Alveolinid-Nummulitid Limestone (Pavlovec 1963).

Based on underlying sediments (deep-marine clastics or shallower- to deeper-marine carbonate platform) and the distribution of larger foraminifera, Drobne (2000, 2003*b*) defined four biosedimentary zones for the area of SW Slovenia and Istria (Fig. 17.30). They were supported by various studies (e.g. Hottinger & Drobne 1980; Hottinger 1990; Drobne & Trutin 1997; Drobne & Ćosović 1998; Drobne & Hottinger 1999; Drobne *et al.* 2000*b*, 2002; Ćosović *et al.* 2004*a*, *b*; Serra-Kiel *et al.* 1998). The following description is arranged according to these biosedimentary zones (BSZ). References for the Palaeogene stratigraphy and palaeontology are presented by Pavlovec *et al.* (1989) and Pignatti *et al.* (1998).

Biosedimentary Zone 1 (BSZ 1; includes clastic sediments from Goriška Brda, the Vipava Valley, and Kališe to the east, and Ilirska Bistrica to the south). Clastic sediments of the Scaglia and flysch beds are exposed in the Epiadriaticum between the thrust front of the Dinaricum in the north and the Cretaceous–Palaeogene platform of the Adriaticum in the SW. Deep-trench pelagic sediments are termed Scaglia or Podsabotin beds in the Slovenian terminology. They consist of red marls interbedding with limestone sheets, and were deposited before and after the K/Pg boundary, with sedimentation extending through the Palaeocene (Šribar 1965, 1967; Pavšič 1977, 1979). They can be traced along the thrust front separating the Dinaricum and Adriaticum in the line from north Gorica to Ilirska Bistrica.

Deep-marine sedimentation is continuous from the Cretaceous to the Palaeogene (Pavšič & Horvat 1988). The Palaeocene–Eocene boundary is recorded in the Nozno section (Pavšič & Dolenec 1995; Pavšič 1997; Dolenec *et al.* 2000*a*, *b*). The northernmost exposures of middle Palaeocene age (e.g. Banjšice Plateau, Soča Valley) comprise characteristic chaotic breccias (Pavšič 1995). The Upper Palaeocene to Upper Cuisian in Goriška

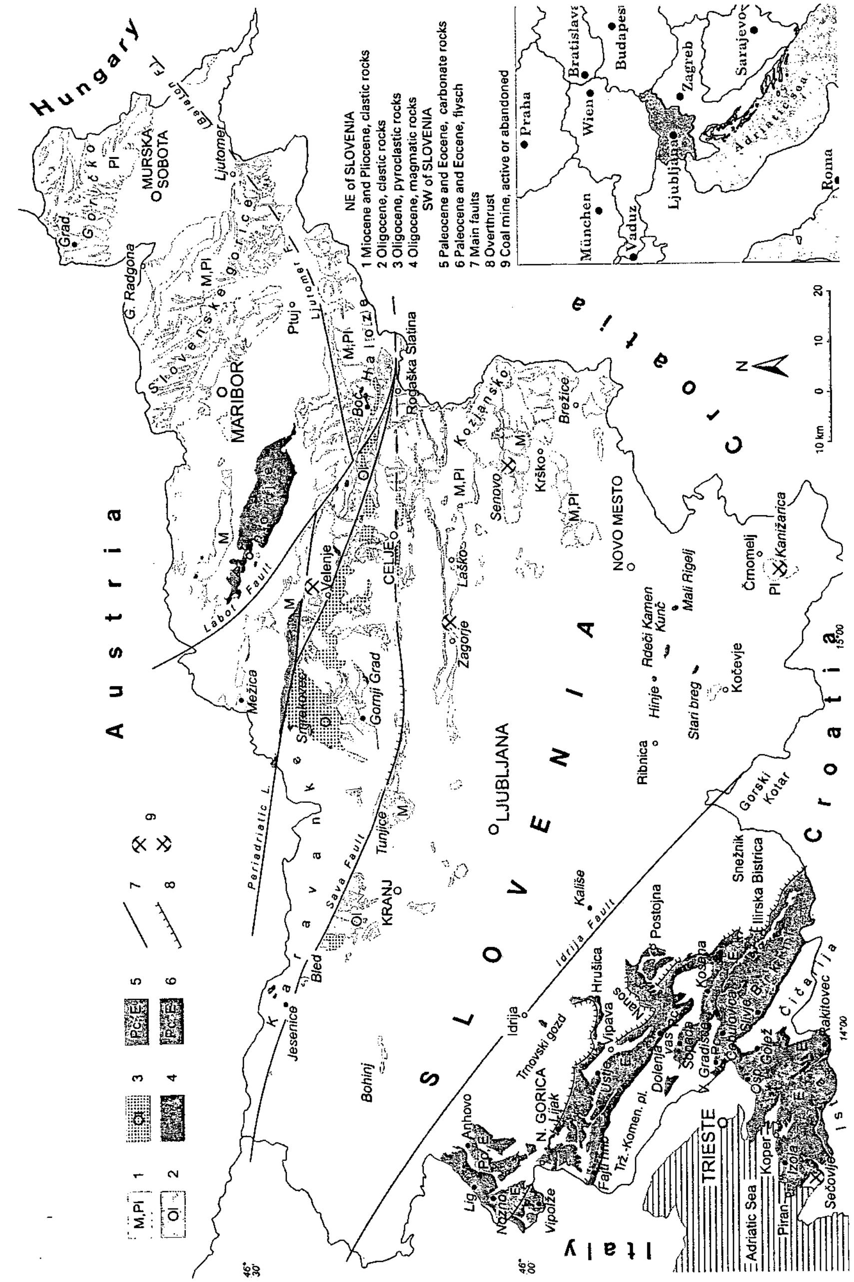


Fig. 17.29. Palaeogene and Neogene rocks in Slovenia.

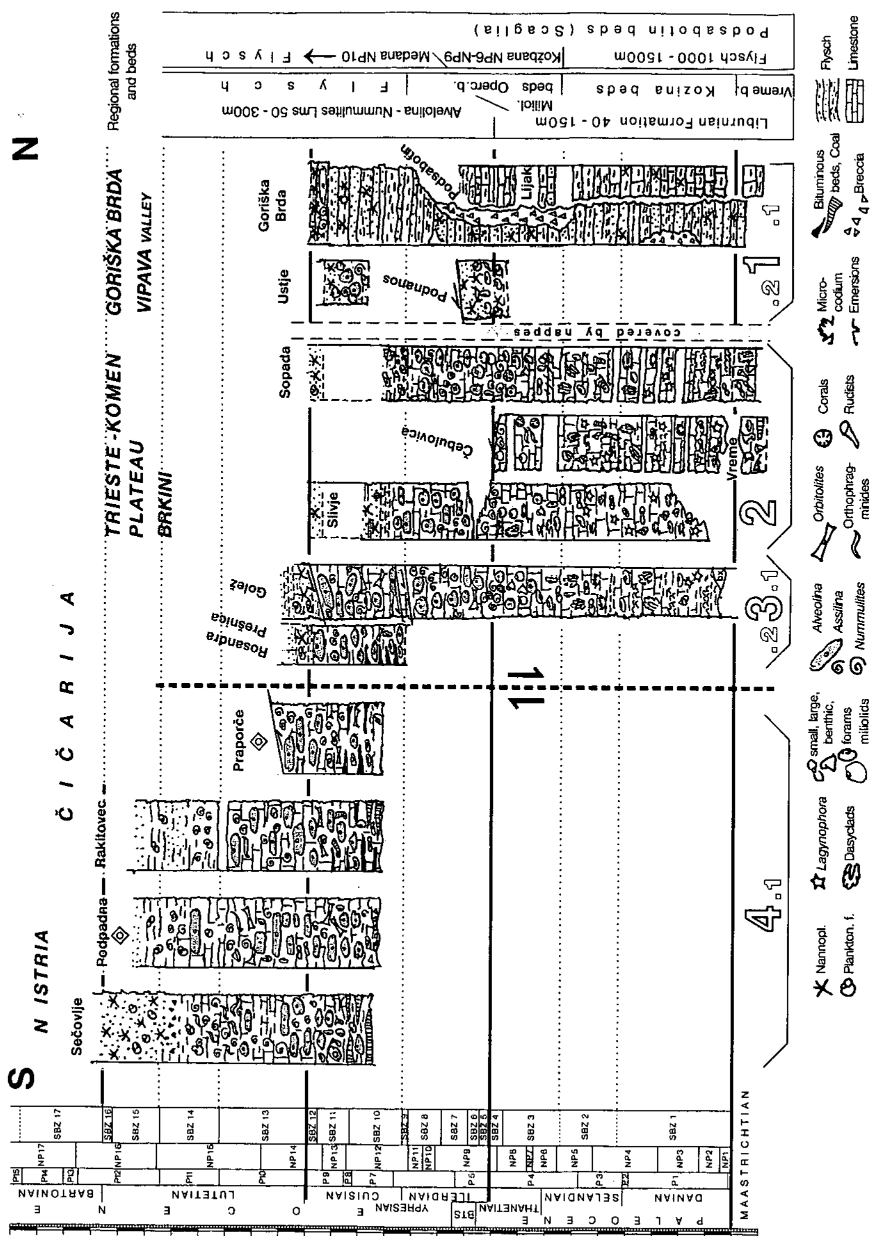


Fig. 17.30. Palaeogene sections in SW Slovenia: lithology, index fossils and biosedimentary zones 1–4 (larger foraminifera zonation SBZ after Serra-Kiel *et al.* 1998).

Brda include the Palaeocene Kožbana Beds (with megaclasts), and the more marly Eocene Medana Beds (Pavlovec 1966; Drobne & Pavšič 1991). Cuisian (SBZ12) clastic sediments in Goriška Brda are mostly developed as proximal turbidites and contain intercalations with pebbles of Cretaceous and Palaeogene limestones. They are particularly rich in nannoplankton and plankton assemblages (Cimerman *et al.* 1974) with abundant corals, gastropods and larger foraminifers (Cimerman *et al.* 1974; Drobne & Bačar 2003; Pavlovec 2004; Mikuž & Pavlovec 2002).

West of Goriška Brda, submarine slides and megaturbidites were described (Tunis & Venturini 1984; Skaberne 1989). Near Nova Gorica, a 1500 m deep borehole records a depression filled with deep-marine sediments (Marinko 1992). In the Gorica area, conglomerates with a source area to the NW and SW (i.e. Friuli platform) were deposited in several phases from Maastrichtian to Middle Eocene times (cf. Venturini & Tunis 1992).

Deep-marine sediments of the Vipava Valley syncline and its SE extension may overlie Scaglia beds. They are either concordant at the platform–basin transition, or follow after a hiatus (Engel 1970). Locally, megaturbidite beds with nummulitic breccias predominate (Engel 1974; Mikuž & Pavlovec 2002), and larger foraminifers of Middle Cuisian age (SBZ11) can occur in upper beds (De Zanche *et al.* 1967; Drobne & Bačar 2003). In the Vipava Valley, the Middle/Late Palaeocene (NP7, Lijak) to Middle Eocene development is well established (Krašeninikov *et al.* 1968; Buser & Pavšič 1976, 1978; Pavšič 1994; Rižnar 1997). At Postojna, deep-marine sediments with conglomerates containing nummulites indicate a Middle Cuisian transgression (Pavlovec 1981).

Separated from the Gorica area and Vipava Valley, several isolated exposures of deep-marine sediments occur in southern Slovenia. They are mostly erosional remnants (Premru 1982; Pavšič 1994, 1995).

Biosedimentary Zone 2 (BSZ 2; includes platform and deep-marine sediments from the southern edge of Vipava Valley, Brkini with border areas, and southwards to the northern slopes of Čičarija). The deep-marine basins of Brkini and Trieste-Komen Plateau are part of the Adriaticum. The Brkini Flysch (Cuisian age: Khan *et al.* 1975; Drobne & Pavšič 1991; Pavlovec *et al.* 1991; Pavšič & Premec-Fuček 2000) was deposited in an elongated syncline with gently uplifted limbs of Palaeocene and Eocene limestones. At the NE margin, there is a gradual transition from limestones to deep-marine sedimentation (Drobne 1977; Pavlovec *et al.* 1991; Knez 1992). At the SW side, this change is more rapid and partly tectonically induced. Deep-marine sediments of the Trieste-Komen plateau (Kras area) are only known from a few isolated localities. The entire belt of deep-marine sediments (flysch) was dated as Cuisian (NP12 to NP14), and is underlain by mostly Ilerdian (partly Lower Cuisian) limestones (Pavlovec 1963; Drobne *et al.* 1991*a*, *b*, *c*).

In this biosedimentary zone, Maastrichtian limestones (Vreme Beds) are overlain by K/Pg boundary breccias, followed by Danian to Ilerdain limestones. They were deposited in shallow-marine environments characterized by occasional emersion events. The inner part of the shelf is typified by the formation of coal prior to, and after, the K/Pg boundary (Hamrla 1959, 1988; Hötzl & Pavlovec 1991). The overlying lagoonal and brackish environments were occasionally affected by emersion. Carbonate sedimentation was coeval with a Late Palaeocene transgression, and persisted until the Early Eocene. Carbonate microfacies and palaeontology have been investigated in several studies (Castellarin & Zucchi *et al.* 1966; Drobne 1977, 1991*a*, *b*, 1996*a*, *b*, 2000*a*, *b*, 2003; Pugliese 1995; Brazzatti *et al.* 1996; Accordi *et al.* 1998; Barattolo 1998; Turnšek & Drobne 1998; Gozzi 2003; Sirel 2004).

Biosedimentary Zone 3 (BSZ 3; includes platform and deep-marine sediments from the SE slopes of Čičarija, Rosandra (Glinščica) valley, eastward to Krk Island). Carbonate platform sediments of this zone occur in a narrow belt extending along the Čičarija Brkini and the karst plateau to the NE, and Golež to the SW and towards the coast along the river Rosandra (Glinščica). To the SE, it extends to the NE margin of Krk Island. These sediments were deposited on a carbonate platform subsequent to Cretaceous–Palaeogene emersion. A pecularity of this development was the continued deposition of the Alveolinid-Numulitid Limestones during the Ilerdian and Cuisian as well as a transition into deeper-marine sediments, with glauconite-bearing strata, at the Cuisian–Lutetian boundary (NP14). Alveolinids predominate, especially those of *Alveolina histrica* (i.e. *Alv. rakoveci*), in addition to nummulitids (Drobne 1977; Drobne & Pavlovec 1991; Drobne & Ćosović 1998; Drobne & Trutin 1997).

Biosedimentary Zone 4 (BSZ 4; includes carbonate platform and deep-marine sediments located along the oldest vertical faults on the SW slopes of Čičarija to Savudrija ridge, to Podpićan, and between Krk and Cres islands). The deep-marine sediments of Čičarija and Slovenian Istria can be subdivided into two complexes of different ages, both of which were part of the Adriaticum. The older one includes the NP14 zone and passes from Alveolinid-Nummulitid Limestone into deep-marine sediments with glauconite beds or with macrofossils (SBZ12?; Mikuž & Pavlovec 2004). At Gračišče, crabs, nummulites and nannoplankton revealed an Early Lutetian (NP15) age (Pavlovec & Pavšič 1987). The coastal belt comprises four sedimentary complexes, passing from limestones with alveolinids and nummulitids (Early to Middle Lutetian, NP15–NP16 boundary) into basinal sediments. Sedimentation continued until the NP16 zone (Pavšič & Peckmann 1996).

Alveolinid-Nummulitid Limestones were deposited from the Middle Cuisian to the Middle Lutetian (SBZ11–SBZ14). They overlie deeply eroded rudist limestones, bauxite, coal and bituminous beds and represent restricted shallow-water environments. At the transition to the Lutetian, foraminiferal associations were part of a larger Central Proto-Tethyan fauna, including species typical for SBZ13 and SBZ14 (Serra-Kiel *et al.* 1998). At the transition from the Middle to the Late Lutetian (NP15–NP16), a subsiding deep-marine basin developed (Drobne 1977; Ćosović & Drobne 1998; Krivic 1982, 1988; Marinko *et al.*1994).

Significant stratigraphic boundaries

In the Slovenian Karst region, the K/Pg boundary is present at several localities (Drobne *et al.* 1988*a*, *b*, 1989, 1995, 1996*a*; Pugliese *et al.* 1995; Dolenec *et al.* 1995; Ogorelec *et al.* 1995, 2001, 2005; Delvalle & Buser 1990). Close to the K/Pg boundary the Maastrichtian lagoonal limestone becomes darker and a freshwater influence is indicated by the presence of ostracods and characeans. The K/Pg boundary itself is characterized by a 20 cm (Dolenja Vas) to 2 m (Čebulovica) thick intraformational breccia of intertidal and supratidal origin, indicated by shrinkage pores and the occurrence of *Microcodium*. This short-lived emersion phase lasted, in Dolenja Vas, for at least 40 ka (Hansen *et al.* 1995). In some sections (Sopada) the breccia is less distinct, and only some plasticlasts and mud-cracked laminae are present (Jurkovšek *et al.* 1996*a*).

Extreme depletion in $\delta^{13}C$ (up to 10‰, from +2 to −8‰ PDB in the Dolenja Vas and Sopada sections) suggests global climatic changes caused by an impact body, followed by the destruction and combustion of terrestrial plants. At Dolenja Vas, the iridium values rise from 0.2 ppb below the boundary up to 5.8 ppb (Hansen *et al.* 1995). Mercury enrichment was reported by

Palinkaš *et al.* (1996) and Hansen & Toft (1996). In the deep-marine basins the K/Pg boundary is known from several locations where it is developed as ooze deposits (scaglia) or within the Podsabotin Formation. The hiatus spans the lowermost Palaeogene (Šribar 1965, 1967; Pavšič 1977, 1981, 1994; Perch-Nielsen & Pavšič 1979).

The Palaeocene–Eocene boundary is documented in the deep-marine sediments of Goriška Brda (Pavšič & Dolenec 1995, 1996; Pavšič 1997). The boundary was defined below the first occurrence date of *Rhomboaster bramlettei* (Bronnimann & Stradner). Anomalous contents of various elements, including iridium, occur somewhat below the biostratigraphically defined Palaeocene–Eocene boundary (Dolenec *et al.* 2000*a*, *b*).

The analyses of $\delta^{13}C$ stable isotopes at the Sopada section (part of BSZ2) reveal an excursion at the K/Pg boundary (depletion of 8‰ $\delta^{13}C$) and a longer-lasting event at the Palaeocene–Eocene boundary with a reduction from +2‰ to −4‰ PDB (Drobne *et al.* 2006), which is supposed to correspond to the global carbon isotope excursion. The excursion is situated within the transitional beds between SBZ4 and SBZ5 (Fig. 17.28). It is characterized by *Lacazina blumenthali*, *Thomasella labyrinthica* and also by a change from Operculina to Alveolinid-Nummulitid limestones. A $\delta^{18}O$ excursion was found in younger beds of SBZ6. It reveals a decrease from −2‰ to −8‰ (PDB), and corresponds with the Palaeocene–Eocene thermal maximum (Pujalte *et al.* 2003).

Slovenian Paratethys basins (B.J., H.R., D.S., M.Po., P.K.)

The Paratethyan deposits of Slovenia (Figs 17.29 & 17.31) are preserved in the Alpine and Dinaric foothills in the central and SE part of the country, in the rolling hills and plains in the east and NE, as well as in some Alpine valleys in the north and NW of the country. From the margin of the Pannonian Basins System in the east, several long, narrow depressions, filled by Cenozoic deposits up to 2 km thick, extend towards the Alpine and the Dinaric mountain chains in the west. The westernmost Paratethyan deposits are found in the valleys of the Julian Alps. The Dinaric chain separates the Mediterranean from the Paratethyan deposits.

Tectonic setting and palaeogeography

From the middle Late Eocene to the Pliocene, depositional basins were continuously created, deformed and destroyed by extensional and compressional tectonic regimes (Jelen & Rifelj 2005*a*, *b*). The area under consideration was subaerially exposed after the suturing of Apulia and the Austro-Alpine units. It underwent extension and flooding during the middle Late Eocene. Block faulting is evidenced by scarp breccias, isolated small carbonate platforms, grabens of different depths, and the diachronous drowning of carbonate platforms. The termination of extension is not known, but the architecture of Early Egerian sediments overlying the Kiscellian deposits is interpreted as a tectonic load-flexural basin sequence.

After Egerian times, the process of post-collisional tectonic escape (Ratschbacher *et al.* 1991) controlled the development of the area. The lateral extrusion process of the Alcapa crustal block from the Eastern Alpine orogenic belt along the Periadriatic Lineament (PAL) fault system and the NW part of the Dinaric orogenic belt (Haas & Kovács 2001; Jelen *et al.* 2001) along the Zagreb-Zemplin Lineament (ZZL) deformed and offset the presumably uniform Palaeogene basin into the present-day Slovenian Palaeogene Basin and the Hungarian Palaeogene Basin (Csontos *et al.* 1992; Jelen *et al.* 1992, 1998; Fodor *et al.* 1998).

Subsequent continental extension (rifting) in a backarc setting resulted in the formation of rapidly subsiding grabens/sub-basins from Karpatian until Middle Badenian times. Synrift subsidence ceased near the Middle–Late Badenian boundary. This was followed by subsidence, which was caused by the post-rift collapse of the extensional basins.

The area to the east of the Julian Alps and NW Dinarides is the western margin of the Pannonian backarc rift. The Venetian Basin to the west of the Dinarides underwent a low rate of ENE–WSW Dinaric compression from the Chattian to the Langhian and was not incorporated into the NNW–SSE South Alpine compression prior to the Serravallian (Massari *et al.* 1986*a*).

Towards the end of the Middle Miocene, minor compression occurred, resulting in uplift and erosion. In the early Late Miocene, the change from uplift and erosion to flooding coincided with extensional reactivation of the Karpatian and Badenian normal faults. In the Pliocene, the climax of compressional activity caused intense folding, accompanied by basin inversion, and strike-slip faulting, accompanied by the formation of small pull-apart and transtensional basins.

Sedimentary and stratigraphic development

Jelen *et al.* (1992) defined four tectonostratigraphic units that correspond to the sedimentary units used herein. These are described from north to south (Fig. 17.31). (1) Unit A1 extends over a wide area north of the PAL (Northern Karavanke mountain range, Pohorje Mountains, Pannonian plain of NE Slovenia) and continues in the Styria (Austria) and in the Zala area and Little Hungarian plain. (2) Unit A2 extends in a very narrow belt between the PAL and the Donat Tectonic Zone (Southern Karavanke mountain range and the Haloze Hills towards the Čakovec area in Croatia). (3) Unit B1 extends in a narrow belt from the Julian Alps in the west and eastwards between the Donat Tectonic Zone and the Sava-Celje Tectonic Zone into the Varaždin area in Croatia. (4) Unit B2 occupies a wide area of central and SE Slovenia between the Sava-Celje Tectonic Zone and the ZZL and continues eastward to Croatia.

Palaeogene Tethys. The oldest Cenozoic deposits developed prior to the creation of the Paratethys Sea in the area under consideration are described above. Other pre-Paratethys remnants are situated to the north of the PAL in unit A1. Two small tectonic enclosures inside the Northern Karavanke overthrust, and south of Kotlje, contain bedded Middle Cuisian to Lower Lutetian limestones (Drobne *et al.* 1977; Mioč 1983). Two further remnants are preserved west of Zreče. There, the Late Cretaceous Gosau beds are unconformably overlain by coal-bearing platy and laminated marly limestones grading into shales (Hamrla 1988). This deeply eroded sequence is covered by fluvial conglomerates, which are interbedded with finer-grained clastic sediments and coal up-section. Alveolinid and nummulitid limestone pebbles indicate a post-Middle Eocene age of formation. These deposits correlate with the Early and Middle Eocene sequences of the Guttaring Group of the Eastern Alps.

Palaeogene Paratethys. Following the suturing of Apulia and the Austro-Alpine units, the area under consideration underwent subaerial weathering, and some bauxites were deposited. Subsequently, depositional basin developments began in Unit A2 within the nannoplankton zone NP19–20 (Jelen *et al.* 2000). These deposits crop out sporadically between the area north of

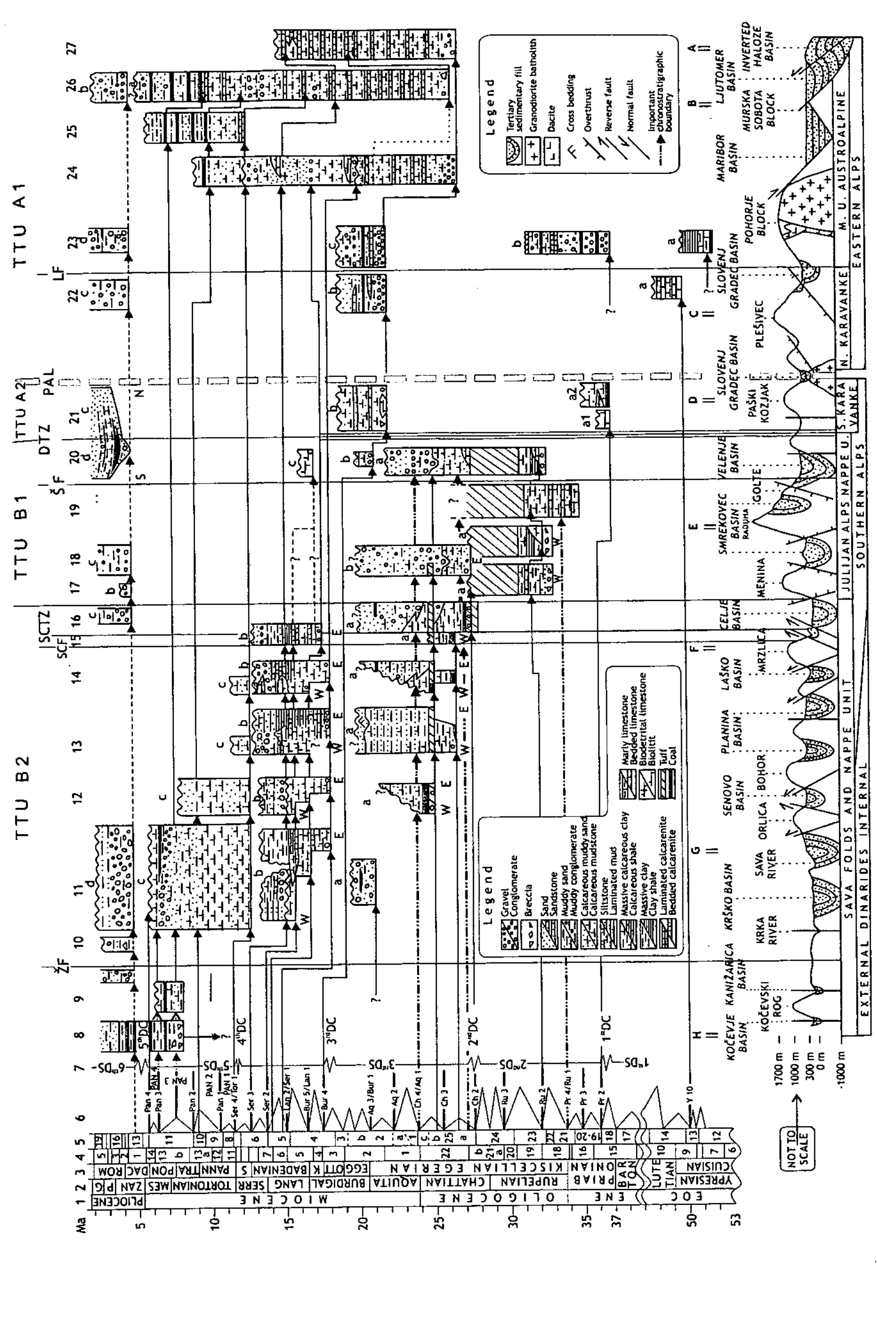


Fig. 17.31. Model of the stratigraphic relations in the Paratethys sedimentary area in Slovenia. TTU, Tertiary tectonostratigraphic units as defined by Jelen *et al.* (1992), corresponding to sedimentary units described in the text. Columns 1–6, chronostratigraphic subdivisions. 1, 2, Standard subdivisions (AQUITA, Aquitanian; LANG, Langhian; SERR, Serravallian; MES, Messinian; ZAN, Zanclean; P, Piacenzian; G, Gelasian). 3, Paratethyan subdivision (EGG, Eggenburgian; OTT, Ottnangian; K, Karpatian; BADEN, Badenian; S, Sarmatian; PANNON, Pannonian; TRA, Transdanubian; PON, Pontian; DAC, Dacian; ROM, Romanian). 4, Planktonic foraminifera biozones. 5, Nannoplankton biozones. 6, Chronosequences after Hardenbol *et al.* (1998), for the Late Miocene after Sacchi (2001); Ru 2, chronosequence boundary. 7, Stratigraphical development stages (DS) and changes in development (DC). Other abbreviations: DTZ, Donat transpressive fault; LF, Labot (Lavantal) fault; PAL, Periadriatic line; SCF, Sava-Celje fault; SCTZ, Sava-Celje tectonic zone; ŠF, Šoštanj fault; ŽF, Žužemberk fault.

Jesenice and the Slovenian–Croatian border (Mikuž 1979; Drobne *et al.* 1985; Šimunić *et al.* in Jelen *et al.* 2000; Jelen & Rifelj 2002). For the Socka area, the co-existence of a platform and a basin can be assumed. To the north, the sedimentary environments changed rapidly up-section, from freshwater to marginal-marine, to outer-shelf, and finally to bathyal environments. The freshwater, coal-bearing strata and brackish calcareous mudstones termed the Socka beds are well-known for their rich flora (Unger 1850). They are overlain by calcareous mudstones deposited on the shelf and slope. Nummulitid-discocyclinid limestones were deposited on the platform to the south, (Jelen *et al.* 2000). During tectonic escape, the basins were inverted and incorporated into the Southern Karavanke shear zone between the PAL and the Donat Tectonic Zone (Fodor *et al.* 1998).

During the latest Priabonian (earliest NP21 or latest NP19–20 zone; Baldi-Beke in Jelen & Rifelj 2002), the extensional basin-forming processes reached unit B1. Unit B1 deposits crop out in various locations between the Julian Alps in Slovenia and the Ravna gora in Croatia (Jelen & Rifelj 2002). The succession of the Smrekovec Basin, with a depocentre in the Luče area, begins with algal-miliolid, nummulitid and discocyclinid limestones. These were suffocated by the oxygen minimum zone. Carbonate-depleted mud was deposited on the outer shelf and upper slope during the Eocene–Oligocene transition. Frequent deep-marine carbonate gravity flows interrupted the low suboxic slope calcareous mud deposition. The latter changed, within the Oligocene part of NP21 zone, to suboxic basin-plain deposition with rare distal calcarenitic turbidites. Its topmost part belongs to NP22. The deepest sediments are the non-calcareous siltstones of the early NP23, in which deep-water agglutinated foraminifera faunas dominate (Jelen & Rifelj 2002).

South of the Luče depocentre, in the Gornji Grad area, delta-plain deposition occurred before late NP22, when a transgression commenced (Báldi-Beke in Jelen & Rifelj 2002). The Gornji Grad Beds (= Oberburg Beds) comprise brackish siliciclastics and marine limestones rich in coralline algae, bryozoa, corals, larger foraminifera and molluscs (e.g. Hauer & Morlot 1848; Reuss 1864; Barta-Calmus 1973; Nebelsick *et al.* 2000). They are overlain by shelf-muds and bathyal turbiditic muds (Scherbacher 2000; Schmiedl *et al.* 2002, and references therein). East of the depocentre, in the Mozirje area, the transgression began in early NP23 (Báldi-Beke in Jelen & Rifelj 2002). Massive calcareous mudstones, in places rich in molluscs, as well as laterally adjacent nummulite limestones, are overlain by a black laminated mud containing fish and plant remains (fish slates of Teller 1896). In the shallower oxygenated grabens, a mud similar to the Kiscell Clay of Hungary was deposited. In the deep sub-sill grabens, deposition of black laminated muds, interrupted by carbonate gravity flows, continued. The locality is historically known for its endemic fauna (Rolle 1858), which is now considered important for the biostratigraphic correlation of the central and eastern Paratethys (Báldi 1984). In the depocentre, non-calcareous silts were interbedded with volcaniclastics derived from the Smrekovec volcanism which occurred within the earliest NP23. The Smrekovec volcaniclastic deposits originate mostly from gravity flows and reach an estimated thickness of 800–1000 m (Mioč 1983; Mioč *et al.* 1986).

The Rupelian foraminifera fauna (Herlec 1985; Drobne *et al.* 1985; Pavlovec 1999), and its position within the Julian Alps nappe structure, suggested that unit B1 continues in the sporadic Cenozoic deposits in the Julian Alps nappe. These deposits, with a total thickness of more than 600 m, consist mainly of alternating various siliciclastics, freshwater limestones and coal lenses. The freshwater succession is topped by a *c.* 20 m thick succession of calcareous siliciclastics and limestones with a rich shallow-water fauna (Herlec 1985; Pavlovec 1999).

Egerian to Eggenburgian Paratethys. Within units B1 and B2, a complex tectonic load-flexural basin, with an inner thrust sheet within the present-day Sava-Celje Tectonic Zone, was formed during the Early Egerian. On the surface, these deposits extend continuously from the Ljubljana Basin to the Slovenia–Croatia border to the east and SE and continue to Croatia. The Celje foredeep basin, in front of the thrust sheet, reached an upper bathyal depth within less than one million years. The basal filling shows that flexure began while the Smrekovec volcanism was still active. The succession grades into marsh, brackish, shallow- and deep-marine environments. The transition to marine mud, (several hundred metres thick) of the underfilling stage took place near the Kiscellian–Egerian boundary (P21–P22 boundary) (Rögl in Jelen & Rifelj 2002). These muds are similar to the Kiscell Clay in Hungary, and are occasionally interrupted by submarine fans. A second volcaniclastic sequence within the mudstones was dated as late Early Egerian age (late NP25) (Báldi-Beke in Jelen & Rifelj 2002).

The Laško back-bulge basin contains economically important coal deposits that have been intensively studied since the middle of the nineteenth century (Ettingshausen 1872, 1877, 1885; Bittner 1884; Papp 1955, 1975; Kuščer 1967; Mihajlović & Jungwirth 1988; Placer, 1999*b*). Volcaniclastic intercalations in coal were dated to 25 ± 1 Ma (Odin *et al.* 1994) and are therefore correlatable with the second volcaniclastic sequence.

A basin north of the inner thrust sheet (i.e. in unit B1) also underwent tectonic load subsidence. Lenses of corallinacean limestones in the shallower part of the inner thrust sheet, and submarine gravel fans in the deeper part, were deposited on the Smrekovec volcanics. Further to the north, there is a transition from volcaniclastic to siliciclastic sedimentation. During the underfilled stage, the northern basin was filled by about 600 m of Early Egerian sediments similar to the Egerian Szécsény Schlier in the Hungarian Palaeogene Basin.

Channel formation occurred prior to the deposition of the second volcaniclastic sequence. They indicate a change in the flexural basin development. Subsequently, the sedimentation rate increased relative to the rate of subsidence and an overfilled basin developed. The interplay of subsidence, sea-level fluctuations and sediment supply resulted in the accumulation of >300 m of sediments. These deposits are best preserved to the east of Celje and Laško. The delta-front shelf tilted towards the north and alternations of sand and mud accumulated rapidly. At the shelf break, glauconitic sand accumulated. Cross-bedding and coal occur at the top of the stacking.

The Adria push and the clockwise rotation of the Tisza Unit, led to inversion, dextral strike-slip faulting and displacement in the Eocene and Oligocene basins. The premise for this displacement is that the NW strike of the Sava-Vardar ophiolite zone changes to a NE strike along the Zagreb–Zemplin Lineament and crops out again in NE Hungary (Bükk Mountains) along with the Szépvölgy Limestone, the Buda Marl, the Tard Clay, the Recsk volcanics and Egerian deposits, all similar to deposits in the Slovenian Palaeogene Basin (Jelen *et al.* 1998, 2001; Haas & Kovács 2001).

Karpatian to Sarmatian Paratethys. A phase of east–west to NE–SW directed backarc extension (rifting) created accommodation space in unit A1 and a NNE–SSW oriented extension created space in unit A2. This resulted in the formation of the Mura-Zala Basin (Fodor *et al.* 2002), which was part of the Styrian extensional wedge of Ratschbacher *et al.* (1991) and of the Raba extensional corridor of Tari (1994). It includes the Radgona-Vas, the Mureck (Cmurek), the Slovenj Gradec, the Maribor, the Eastern Mura-Őrség and the Haloze-Ljutomer-Budafa grabens/sub-basins as well as the Pohorje and the Murska Sobota extensional blocks.

During the first phase of extension in the Karpatian, benthic and planktonic foraminifers indicate deep-marine, restricted environments. Proxy equations of van der Zwaan *et al.* (1990) suggest that a water depth of 840 m (±20%) was reached very quickly in the grabens. They were filled by a mud/sand-rich system of gravity flows, and by dacite/andesite volcaniclastics, with a sedimentation rate of 1000–2000 m/Ma. Sediments were derived from a fluviodeltaic system of muddy flooding rivers and alluvial fans. These processes coincided with the formation of a significantly high relief in the Eastern Alps and the activity of a number of strike-slip and normal faults (Kuhlemann *et al.* 2002). We assume that towards the end of this first phase, the extension established a connection between the Paratethys and the Mediterranean. This happened during the falling stage systems tract between the Bur4 and Bur5/Lan1 sequence boundaries of Hardenbol *et al.* (1998) as indicated by the contemporaneous leap in abundance of the benthic and planktonic foraminifera and the first appearance of species immigrating from the Mediterranean (Jelen & Rifelj 2003). The Karpatian marine sediments are limited to the Styrian extensional wedge and to the marginal basins/feeding canyons south of the Donat Tectonic Zone. The Karpatian age of the brackish water, fluvial and mire deposits west of the wedge has been assumed on the basis of the observed interdigitation (Mioč & Žnidarčič 1989).

The Karpatian–Badenian boundary in Slovenia corresponds with the Styrian unconformity (Mioč & Žnidarčič 1989) and thus with the Bur5/Lan1 sequence boundary. Recent studies (Rögl & Rifelj, unpublished data) suggest, however, a position beneath the Styrian unconformity, corresponding with the increasing abundance of foraminifera and Mediterranean invaders. This finding is in accordance with the observation that the cooling events Mi-2 and MLi-1 are younger than the proposed age of the Burdigalian–Langhian boundary at the first occurrence of *Praeorbulina sicana* (Abreu & Haddad 1998). The Styrian unconformity in this area is a product of tectonic activity and a eustatic sea-level fall.

The Early Badenian was a time of significant paleogeographic changes caused by a second, strong extensional pulse and the Langhian eustatic sea-level rise of Hardenbol *et al.* (1998). Between the Donat Tectonic Zone and the ZZL, the Kozjansko crustal block underwent normal faulting and present-day east–west trending faults were reactivated as scissors faults. During the concurrent range zone of *Orbulina suturalis* and *Preorbulina circularis*, flooding created accommodation space along the boundary-normal faults in the eastern part of the Kozjansko Block and westward along the deep graben, following the central scissor transfer fault. The subsidence rate was very high and the water depth may have reached 500 m. The late Early Badenian transgression was accompanied by the formation of organic buildups composed of corallinacean algae, bryozoans and corals. Rhodoliths with diameters of up to 15 cm occur (Aničić & Ogorelec 1995). The short-lived carbonate platforms were soon tectonically tilted and destroyed (Kázmér *et al.* 2005), followed by the deposition of bathyal mud and calcareous/siliciclastic turbidites.

Continuous extension, which masked the eustatic signal, was followed by the late Middle Badenian transgression. The Late Badenian extensional collapse (near the Middle–Late Badenian boundary) and the Late Badenian flooding of Hardenbol *et al.* (1998) resulted in the maximum extension of the Badenian sea. Its westernmost remnant was found at Kamnik (Premru 1983). The collapse produced confined basins, calciruditic submarine fans, turbidites and laminated muds, with blooms of benthic foraminifera tolerant to oxygen depletion.

Most of the Senovo and Krško basins comprise Late Badenian mainly shallow-subtidal limestones and some calcareous mudstones. From the Krško Basin, Mikuž (1982, 2000) described a rich gastropod fauna. The succession could be dated by diatoms and silicoflagellates (Horvat 2003).

In the Mura-Zala Basin, the combination of the second, strong extensional pulse and a major eustatic flood resulted in water depths of 880 m (±20%) during the late Early Badenian. The lowstand wedge is overlain by transgressive coralline algal limestones with *Orbulina suturalis* and *Preorbulina circularis*, followed by deep-water calcareous muds and rare turbidites. A sudden change to sand-rich turbidites close to the final occurrence of *Preorbulina circularis* may have been related to the eustatic sea-level fall at the Lan2–Ser1 boundary of Hardenbol *et al.* (1998); this was followed by an extensional collapse creating confined basins.

A sea-level fall at the Badenian–Sarmatian boundary, which is correlated with the Ser3 sequence boundary of Hardenbol *et al.* (1998) and the MSi-3 isotopic event (Abreu & Haddad 1998), caused further facies differentiation. On the Kozjansko Block, shelf areas were temporarily subaerially exposed and subsequently flooded by the Early Sarmatian eustatic sea-level rise (Rižnar *et al.* 2002; Horvat 2003; Rifelj *et al.* in press). Various muds and laminated to bedded calcarenites contain diatoms, silicoflagellates, molluscs and plant remains (Horvat 2003, and references therein). In shallow basins, discrete occurrences of Badenian/Sarmatian limestones of various thicknesses, and rich in coralline algae, conformably overlie Badenian mudstones. Limestones and mudstones are overlain by laminated calcareous shales or calcareous mudstones and calcarenites. Sandstones, conglomerates, clays and corallinacean limestone, containing brackish molluscs (Bittner 1884), and in the deepest basins turbiditic sandstones, cover the calcareous mudstones and calcarenites. During the Sarmatian, episodes of marine/brackish conditions alternate with brackish/freshwater conditions (Horvat 2003; Rifelj *et al.* in press). The westernmost Sarmatian deposits are known from Kamnik (Premru 1983), where calcareous mudstones with brackish-water molluscs (Kühnel 1933) and diatoms (Horvat 2003) are overlain by poorly consolidated siliciclastics.

In the shallow parts of the Mura-Zala sub-basins, lowstand coarse clastics are overlain by sporadic corallinacean limestones, mudstones and siltstones with molluscs and plant remains. In the upper part of the succession, poorly consolidated sandstones and conglomerates predominate (Mioč & Žnidarčič 1989). Fan deposits and distal turbidites are the dominant sediment types in the deeper parts of the sub-basins. During the Late Sarmatian, the Neogene sedimentary sequence was locally deeply eroded due to tectonic uplift and a major eustatic sea-level fall at the Ser4/Tor1 boundary of Hardenbol *et al.* (1998).

Pannonian *sensu stricto* to Pontian Paratethys. The Late Sarmatian uplift and erosion was followed by the Early Pannonian subsidence and flooding. On the flanks of the Mura-Zala sub-basins, Pannonian mudstones interspersed with sandstones onlap eroded Sarmatian and Badenian rocks and the metamorphic basement of the Murska Sobota extensional block. Turbidite deposition continued in deeper parts of the basins. On the

Kozjansko Block, Early Pannonian deposits transgrade over Sarmatian, Badenian, or pre-Cenozoic rocks (Poljak 2004; Aničić 1991). Sediments of the Kozjansko Block are well studied in the Krško Basin (Poljak 2004; Škerlj 1985; Stevanović & Škerlj 1985, 1990) and the succession can be correlated with the seismic sequences of Sacchi (2001) as follows. Aggradational calcareous muds containing *Congeria czjzeki* and progradational sands containing *C. praerhomboidea* are equivalent to the PAN-2 (Fig. 17.31) seismic sequence of Sacchi (2001). The overlying aggradational sandy calcareous muds with *C. rhomboidea* and the following progradational calcareous muddy sands with a coal deposit on the top is equivalent to the PAN-3 seismic sequence. The youngest deposits are calcareous muds containing brackish-water ostracods and are equivalent to the PAN-4 seismic sequence. The *C. czjzeki* aggradational unit corresponds with the Pannonian *sensu stricto (sensu* Stevanović 1951, 1990). The *C. praerhomboidea* progradational and the *C. rhomboidea* aggradational units correlate with the Transdanubian (*sensu* Sacchi 2001). The second progradational unit plus PAN-4 are correlatable to the Pontian *sensu stricto* (*sensu* Sacchi 2001). Other remnants along the Slovenia–Croatia border are Pannonian marls and marly clays interspersed with sands and sandstones (Aničić 1991).

In the Mura-Zala Basin, the delta began to prograde into the basin during the Sarmatian, and in the Krško Basin during the early Transdanubian. By the end of the Pontian, the delta plain extended over both basins. Transdanubian sediments attain a thickness of 350 m in the Krško Basin and 500 m in the Mura-Zala Basin; the thickness of Pontian sediments in the two basins is 1100 m and 1700 m, respectively. It is not clear, whether this was the result of post-rift subsidence or post-rift activity coupled with the increasing intraplate stress caused by the Adria microplate CCW rotation (Márton *et al.* 2003*b*). This movement began at *c.* 9 Ma, which is the time of the opening of the Tyrrhenian Sea (Jelen & Rifelj 2005*a*, *b*).

Pliocene. Intense folding and the development of pop-up structures supplied material to alluvial fans and fluvial systems; e.g. the Haloze part of the Ljutomer-Haloze-Budafa sub-basin was positively inverted and about 2000 m of sediments were eroded (Sachsenhofer *et al.* 2001; Fodor *et al.* 2002; Márton *et al.* 2002*a*). Coeval strike-slip faulting created small pull-apart and transtensional basins (Vrabec 1994, 1999) filled by lake sediments.

The Velenje Basin includes some of the best studied lake sediments in the region. The basin was presumably formed during the Middle Pliocene as a transtensional basin with a half-graben geometry (Brezigar *et al.* 1987; Vrabec 1999). Pliocene sediments of this basin include basal silts with boulders, marsh and mire lignite, shallow lake marls and clays as well as massive and laminated clays with pebbly sands. At the depocentre, near the Šoštanj Fault, these sediments are about 1000 m thick.

A change from the *Taxodium* to *Fagus* palynoflora in the laminated clays represents the Middle–Late Pliocene boundary. The Pliocene–Pleistocene boundary is transitional within the transitional shallow lake/terrestrial silts and sands. Shallow-water lake deposits are rich in molluscs, mammals and plants (Brezigar *et al.* 1987). Lamination is interpreted as being related to planktonic diatom stratification in a eutrophic lake. The geology and petrology of the 160 m thick uniform lignite seam was studied by Brezigar (1987) and Markič & Sachsenhofer (1997).

Volcanism

Cenozoic volcanic activity in NE Slovenia is closely related to the tectonic evolution of the Pannonian Basin within the Alpine–Dinarides–Carpathian orogenic belt. This commenced with the subduction of the European Plate below Africa and continued with the collision, post-syncollisional transpression, separation and eastward escape of the Pannonian fragment, and finally basin extension. Calc-alkaline volcanism apparently commenced in the Eocene and reached its climax during the Late Oligocene and the earliest Miocene. It ceased during the Badenian. The youngest volcanic rocks are alkaline basalts that extruded after the main extension of the Pannonian Basin during the Upper Pliocene–Romanian.

The Oligocene to Early Miocene magmatic period along the easternmost sector of the PAL was very intense. It yielded Karavanke and Pohorje tonalite and granodiorite intrusions (Altherr *et al.* 1995; Pamić & Palinkaš 2000) and volcanic rocks of the Smrekovec suite which encompass the Smrekovec volcanic complex and the occurrences of pyroclastic and volcaniclastic rocks extending from NW Slovenia (Peračica) via Sava Folds to the Celje, Laško and Mura basins (Hinterlechner-Ravnik & Pleničar 1967; Kralj 1996, 1999). Volcanic rocks of the Smrekovec suite range in composition from basaltic andesite to rhyolite, and mainly exhibit a medium-K affinity.

Three main volcanic lithofacies groups were recognized: coherent, autoclastic and volcaniclastic types (Kralj 1996). Coherent rocks are developed as lavas, shallow intrusive bodies or volcanic vent fillings. Marginal parts of lavas and shallow intrusive bodies are commonly autobrecciated with a tendency to grade into hyaloclastite breccias and peperitic breccias. Peperites are less common.

Volcaniclastic deposits are the most widespread lithofacies group. Submarine pyroclastic flow deposits of dacitic to rhyolitic composition can be >100 m thick, and consist of pumice and volcanic glass shard-rich tuffs (Kralj 1999). Secondary volcaniclastic deposits are abundant in the Smrekovec volcanic complex and comprise volcaniclastic debris flows and turbidity flows. They form internally stratified, fining-upward sequences.

Late Pliocene alkaline basaltic volcanism occurred in the northwesternmost margin of the Mura Basin in the Late Pliocene (Romanian). This volcanism is related to the extension of the Pannonian fragment and upwelling of the asthenosphere (Embey-Isztin & Kurat 1996). Alkali basaltic volcanism occurred in a fluvial environment characterized by rapid sedimentation (Kralj 1995). Initial magmatic eruptions created a cinder cone with minor lava flows. Occasionally, the style of eruptions became essentially hydrovolcanic, producing pyroclastic surge deposits, and in the final stage, large lahar deposits. Alkali basaltic volcanism can be correlated with the final stage of volcanic activity in the neighbouring Styrian Basin (Poulditis 1981; Poschl 1991), Little Hungarian Plain and the Bakony–Balaton highland (Martin & Németh 2004). Palaeogene volcanism of the Smrekovec Basin was recently studied by Hanfland *et al.* (2004).

Dinarides in north Croatia and Bosnia (D.P., M.B.)

The area of Croatia can be divided into three major units (Fig. 17.26). (1) The Pannonian and Peri-Pannonian area comprises the lowland and hilly parts of east and NW Croatia. (2) The hilly and mountainous Dinarides separate the Pannonian Croatia from the coastal region. (3) The Adriatic Area includes a narrow coastal belt separated from the hinterland by high mountains. This is predominantly a karst area.

Tectonic setting

The geological evolution of the areas of central and northern Croatia and Bosnia and Herzegovina differed during the Palaeogene and Neogene. The Palaeogene basin development was mainly controlled by compressional events, which resulted in closure of the Western Tethys and uplift of the Dinarides, while extension, interrupted by minor compressional events, controlled sedimentation in the Neogene. The area of north Croatia and north Bosnia was characterized by changing marine connections, typical for the Central Paratethys.

The formation of the Palaeogene basin in the area of north Croatia and north Bosnia was related to Jurassic/Cretaceous subduction along the north Tethyan margin. This initiated the gradual closure and shortening of the Dinaridic Tethys and the development of a magmatic arc. In the trench associated with this magmatic arc, Late Cretaceous–Palaeogene deep-marine sequences accumulated. Compression at the end of the Eocene was accompained by uplift of the Dinarides (Pamić *et al.* 1998, 2000*a*). This caused the formation of local alluvial environments in eastern Croatia (Halamić *et al.* 1993).

The Neogene basins of north Croatia and north Bosnia were part of the south Pannonian Basins System (PBS), and were formed due to the collision of the European (Tisia-Moesia) and the African plates (Horváth & Royden 1981; Horváth 1995; Kováč *et al.* 1998). Following separation of the Western Tethys into the Paratethys and Mediterranean during the Late Eocene, the northern part of the uplifted Dinarides became emergent. During the Oligocene transpressional phase, a Periadriatic dextral strike-slip fault, controlling the formation of the SW margin of the PBS along relicts of the previous subduction zone, was generated. The passive, lithosphere-generated rifting processes led to the formation of elongated half-grabens during the synrift phase. This began during the Ottnangian and lasted until the Middle Badenian. At the end of the Karpatian, uplift commenced as a consequence of the rotation of fault blocks around a horizontal axis. The uplift was contemporaneous with sinistral NE–SW strike-slip faulting, which was transverse to oblique to the master WNW–ESE elongated structures causing CCW rotation and the destruction of the elongate half-graben structures. This locally reduced the effects of uplift (Jamičić 1983; Pavelić *et al.* 1998, 2003; Márton *et al.* 1999*b*, 2002*b*; Pavelić 2001).

The post-rift phase lasted from the Middle Badenian to Recent, and was characterized by thermal subsidence, which was interrupted by two compressional phases generated by intraplate stress. The onset of the first period of intraplate stress took place at the end of the Sarmatian and may have initiated the uplift of blocks, resulting in base-level fall and partial basin inversion. The second intraplate stress phase affected the basin during the Pliocene, causing overall compression and structural inversion across the North Croatian Basin and north Bosnia. In these areas it was characterized by the formation of several compressional structures, subsidence, the uplift of basement blocks, CCW rotations, and erosion (Márton *et al.* 1999*b*, 2002*b*; Pavelić 2001; Tomljenović & Csontos 2001; Pavelić *et al.* 2003; Saftić *et al.* 2003).

Tectonic activity was the main external control on sedimentation in the Neogene intramontane basins of the Dinarides and it can be assumed that both compressional and extensional events influenced basin evolution. The Dinaridic intramontane basins were formed after the uplift of the Dinarides due to compression. Most of the Dinarides became emergent during the Palaeogene and normal faulting may have caused the formation of small tectonic depressions during the Oligocene and Neogene. Tectonically-controlled subsidence facilitated high sedimentation rates, although Pleistocene tectonics may have generated differential uplift of the Dinarides (Soklić 1970; Herak 1986). In addition to compression, extension was also active during the Pleistocene (Jamičić & Novosel 1999). In the Pleistocene, the uplift of the Dinarides to Recent elevations caused erosion of the Neogene sediments and reduced the dimensions of the freshwater basins. Frequent earthquakes along the eastern Adriatic coast suggest that compressional tectonics is still active today (Prelogović & Kranjec 1983; Papeš 1985; Herak 1986; Blašković 1999; Dragičević *et al.* 1999).

Sedimentary and stratigraphic development

Palaeogene. In the north Dinarides (i.e. north Bosnia), Palaeogene sediments form part of a Late Senonian–Palaeogene sedimentary sequence. The K/Pg boundary is frequently characterized by a significant biostratigraphic discontinuity (Polšak 1985). The uppermost Cretaceous–Palaeogene sedimentary sequence consists mainly of turbidites deposited in a narrow basin (Jelaska 1978). The deep-marine sedimentation, starting in the Maastrichtian, consists of individual graded sequences. It is dominated by sandstones and shales during the Maastrichtian and Palaeocene, while calcareous shales and sandstones, sandy limestones and limestones prevail during the Early and Middle Eocene. These deep-marine sediments are conformably overlain by late Middle Eocene limestones. This succession reflects the Palaeogene termination of subduction and the convergence of stable Africa and Eurasia in the area of the northern Dinarides (Jelaska 1978; Polšak 1985; Pamić *et al.* 1998). Tectonic uplift and erosion of the Dinarides took place at the end of the Late Eocene and Early Oligocene.

The Palaeogene sediments occur in scattered outcrops and in the subsurface of NW and central Croatia. Palaeocene sediments include marine clastics and limestones such as algal and coral limestones (Jelaska *et al.* 1970; Šikić *et al.* 1979; Pikija 1987). At the beginning of the Eocene, sedimentation of the Cretaceous–Palaeogene deep-marine succession was interrupted, and initial emergence occurred in NW Croatia. The emergence phase lasted until the Middle Eocene, when terrestrial clast accumulations were flooded by a marine transgression (Šimunić *et al.* 2000). Several Eocene sediment types occur in NW Croatia: Early Eocene limestone breccias, calcarenites, shales and conglomerates pass upward into reddish-brown gravels, sands and tuffitic clays, and Early Eocene bauxites; Middle Eocene coarse-grained clastic limestone breccias and limestone-dolomite breccias; shallow-marine Middle to Late Eocene limestones (coral biolithites, algal-foraminiferal biomicrites and biomicrudites) (Šimunić *et al.* 2000).

Marine sediments are also known from the Oligocene of NW Croatia. They are characterized by marls with sand intercalations (Šimunić 1992). At Mount Požeška (east Croatia), Oligocene coarse-grained clastics were deposited in an alluvial environment, thus reflecting the existence of continental conditions (Halamić *et al.* 1993).

Neogene of north Croatia and north Bosnia. Neogene rocks cover a large area of north Croatia and north Bosnia (Fig. 17.32). The thickness of Neogene deposits is highly variable and is more that 6500 m in the Drava depression. Egerian–Eggenburgian brackish-water to marine sedimentation was restricted to the Hrvatsko Zagorje Basin, i.e. to the area of the PAL (Šimunić *et al.* 1990; Pavelić *et al.* 2001). Marine deposition continued into Ottnangian times.

South and SE of the Hrvatsko Zagorje Basin (i.e. south Pannonian Basins System), alluvial and lacustrine sedimentation commenced in the earliest synrift phase (Pavelić & Kovačić 1999; Pavelić 2001; Saftić *et al.* 2003). The occurrence of

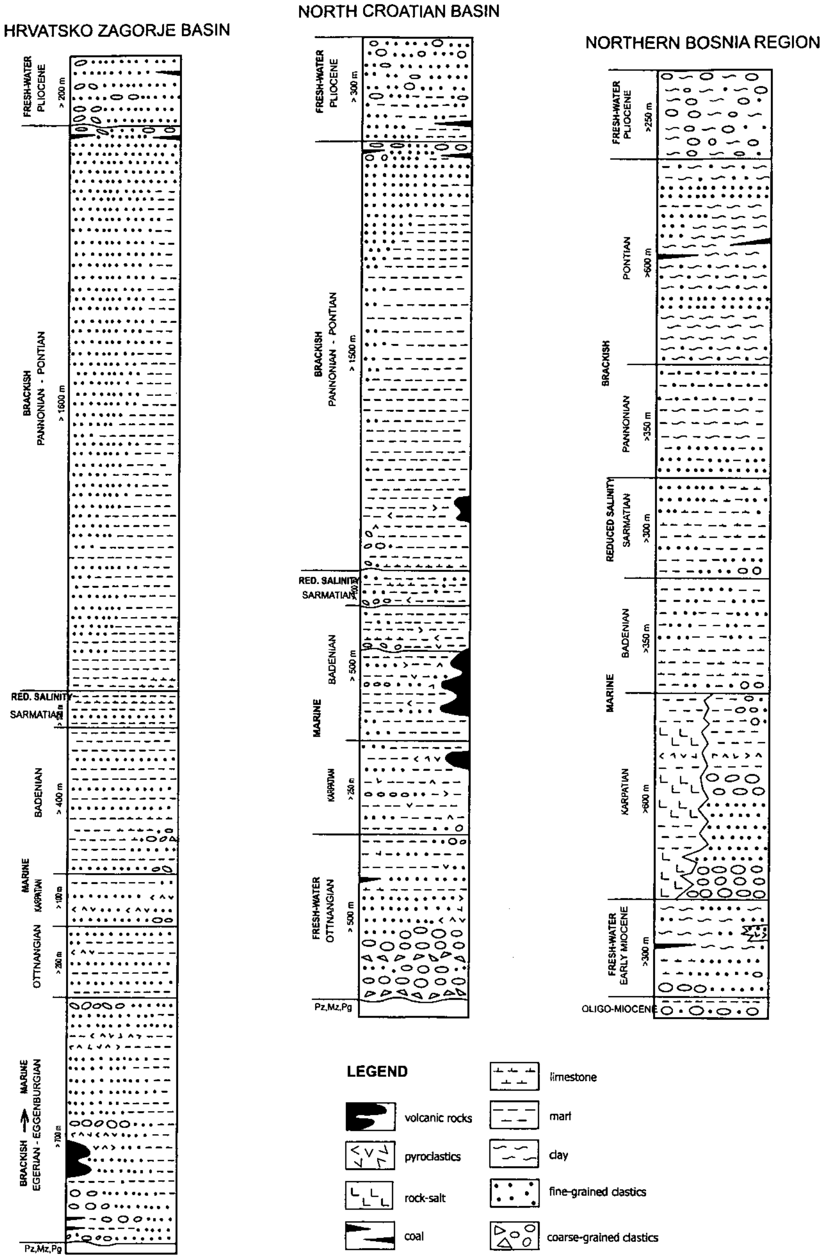


Fig. 17.32. Geological columns of the southern Pannonian Basins System.

Mastodon angustidens indicates a non-marine development during the Early Miocene of the Tuzla Basin (Soklić & Malez 1969). During the Late Ottnangian, a hydrologically open lake covered the whole area. Lacustrine sedimentation was accompanied by explosive rhyolitic volcanism (Mutić 1980; Vrabac 1999; Pavelić 2001). Marine environments commenced during the Karpatian (Pavelić 2001; Bajraktarević & Pavelić 2003; Saftić *et al.* 2003). They are characterized by calcareous siltstones with intercalations of clastics. The Karpatian sediments in the Tuzla sub-basin contain rock-salt and pyroclastic layers (Vrabac *et al.* 2003).

Due to a relative sea-level fall at the end of the Karpatian, some blocks were exposed, which locally resulted in the complete erosion of Early Miocene deposits and the exposure of basement rocks. The eroded siliciclastics were transported into high-energy, shallow-marine environments (Pavelić *et al.* 1998; J. Velić *et al.* 2000), and also into the relatively deeper sea (Pavelić *et al.* 1998). A deepening event during the Early Badenian caused the deposition of marls and gravelly calcarenites in the offshore areas (Pavelić *et al.* 1998). During the Late Badenian, the transgression flooded the peaks of the exposed blocks, which had formed isolated islands during the Early Badenian. Marine sedimentation commenced with the deposition of gravels, which are overlain by coralline algal beds. Further deepening resulted in the deposition of marls. In the Early Sarmatian, the salinity of the sea decreased, and the environment became mesohaline. The isolation of the basin caused a sea-level fall during the latest Badenian, which resulted in the resedimentation of older Badenian faunas (Pavelić 2001; Saftić *et al.* 2003). A subsequent transgression resulted in widening of the basin. Laminated marls (similar to varves) and massive marls dominated in this period, and the episodic input of sands took place by sediment gravity flows. The excellent preservation of the lamination may reflect anoxic conditions (Pavelić 2001).

The ecological conditions changed during the Early Pannonian. The environments became brackish (oligohaline) and locally freshwater, which caused the development of endemic molluscs and ostracods. Early Pannonian deposits overlie Sarmatian sediments almost conformably and include lacustrine platy and thin-bedded, littoral limestones. Resedimented Badenian fossils are abundant in the middle Pannonian deposits, reflecting a short-lived latest Sarmatian period of emergence and erosion. A subsequent lake-level rise affected the dominance of marls with occasional siliciclastic influx. During the Pontian, gradual shallowing is reflected by the increased terrigenous sedimentation within a prodelta environment. Sedimentation terminated during the Pontian with sandy delta progradation and the formation of peat bogs (Vrsaljko 1999; Pavelić 2001; Saftić *et al.* 2003; Kovačić *et al.* 2004).

Late Miocene deposits are overlain by Pliocene siliciclastic sediments accumulated in small freshwater lakes, swamps and rivers. Pleistocene deposits are comparable to those of the Pliocene, except for the remarkable amounts of aeolianites (J. Velić & Durn 1993; J. Velić & Saftić 1999; Saftić *et al.* 2003).

Neogene intramontane basins of the Dinarides. In the Dinarides, sedimentation of Neogene freshwater sediments began within intramontane depressions that had started to form during the Late Palaeogene (Fig. 17.33). Neogene sediments overlie the pre-Neogene basement or Oligocene deposits, and are characterized by frequent lateral and vertical facies variations as a result of independent local basin developments. Throughout Bosnia and Herzegovina as well as central and south Croatia, earliest Miocene sedimentation was restricted to freshwater basins, except for the SE part, which was probably flooded from the Adriatic Sea. The thicknesses of the Neogene deposits in these basins vary from a few hundred to 1900 m in the Livno-Duvno Basin and more than 2400 m in the Sarajevo-Zenica Basin. However, the stratigraphy of these freshwater deposits is still problematic (Pavelić 2002).

The oldest Neogene sediments belong to the upper part of the Oligo-Miocene series. Siliciclastics include conglomerates, sandstones, marls and clays. Characacean limestones and coal beds are sometimes found. Similar conditions occurred in the Early Miocene, and were followed by sporadic volcanic activity. Freshwater endemic bivalves *(Congeria)* had their main phase of evolution at this time (Kochansky-Devidé & Slišković 1978). Middle Miocene deposits are similar to those of the Early Miocene. Marls, limestones, sandstones, clays and conglomerates predominate. Pyroclastics occur in some areas. Thick marly limestone units with occasional coal seams are Middle and Late Miocene in age. The Upper Miocene and Pliocene deposits are characterized by marls, clays, siltstones, sandstones, conglomerates, limestones and coal seams. Pliocene sediments are known from a few localities, such as a series with lignite in the Livno-Duvno Basin and the Sarajevo-Zenica Basin. The youngest Pliocene to Pleistocene sediments are found in the intramontane basin in the NW Dinarides (Jurišić-Polšak *et al.* 1997).

Volcanism

Palaeogene magmatic rocks in north Croatia and north Bosnia were generated by the collision of the NE parts of the Apulian Plate and the SW margin of the Eurasian Plate. The final stage of this subduction is recorded by the sedimentary, magmatic and metamorphic units of the Prosara and Motajica mountains in north Bosnia, which are interpreted as remnants of a subduction-related magmatic arc (Pamić 1977, 1993; Lanphere & Pamić 1992). In north Bosnia, granitoids occur both on the surface and the subsurface. They occur as veins and small- to medium-sized synkinematic plutons in medium-pressure metamorphic rocks. Isotopic ages of Mesoalpine granitoids, which are associated with andesites and dacites, range from 48 to 30 Ma. In the basement of the Neogene of the north Croatian Basin, rhyolites of the ophiolite complex yielded ages of 67–47 Ma (Pamić 1993; Tari & Pamić 1998).

Neogene volcanism is generally related to extensional processes, except for the oldest volcanic units. Volcanic rocks occur mostly in the area of the Hrvatsko Zagorje Basin, and in the northern part of the North Croatian Basin, i.e. in the Drava depression. These rocks are divided into several volcanic formations: the Egerian–Eggenburgian dacite–andesite formation, the Ottnangian pyroclastics, the Karpatian trachyandesite (latite) formation, the Badenian andesite–basalt formation with subordinate dacites and rhyolites, and the post-Badenian basalt–alkali basalt formation (Pamić 1997; Pavelić 2001).

The rocks of the Egerian–Eggenburgian dacite–andesite formation are known only from the Hrvatsko Zagorje Basin and from neighbouring Slovenia. They may be associated with brackish-water and marine sands, sandstones, marls, breccias and conglomerates (Šimunić & Pamić 1993). The main phase of Alpine deformation (i.e. the Pyrenean phase), which was followed by a transpressional phase, controlled this magmatic activity (Laubscher 1983; Pamić 1997).

The Ottnangian was characterized by very low rates of volcanic activity. Lacustrine sedimentation was accompanied by explosive volcanism, resulting in the deposition of tuffs and tuffites, reflecting the commencement of rifting (Mutić 1980; Ščavničar *et al.* 1983; Pavelić 2001).

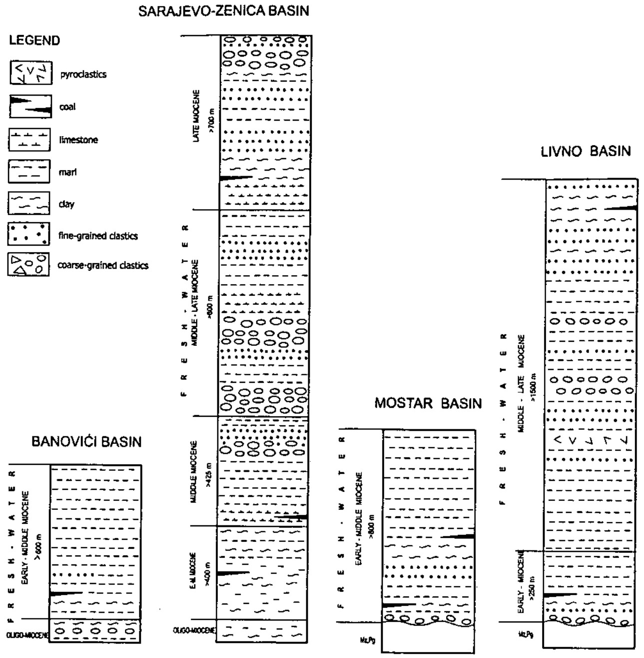


Fig. 17.33. Geological columns of the Dinarides intramontane freshwater basins.

The Karpatian volcanics are known only from east Croatia. They comprise a volcanic body of 5 km^2 on Mount Krndija made up of trachyandesites and tuffs interlayered with marine clastic sediments. These volcanics may have originated from the partial melting of upper mantle rocks enriched in MgO (Golub & Marić 1968; Pamić *et al.* 1992/1993; Pamić 1997).

The Early Badenian is characterized by a climax in volcanic activity. The rocks of the Badenian andesite–basalt formation, with subordinate dacites and rhyolites, are the most widespread volcanics in the North Croatian Basin (Lugović *et al.* 1990; Pamić 1992). Peperites and pillow lava also occur (Belak *et al.* 2000). The thickness of the volcanics varies and is more than 1000 m in some parts of the Drava depression. They are also found in the Hrvatsko Zagorje Basin. The volcanics are frequently intercalated with marine clastics and marls. In some places pyroclastics occur within algal limestones (Belak *et al.* 1991). The rocks of this formation probably originated from the partial melting of the heterogeneous lower crust (Pamić 1997), and are a result of the final stage of the synrift phase of basin evolution (Pavelić 2001).

The rocks of the post-Badenian basalt–alkali basalt formation are known only from a few wells in the Drava depression. K–Ar measurements carried out on basalts yielded isotopic ages of 11.6–9.4 Ma (Pamić 1997). The occurrence of these rocks has been interpreted as being related to a short-lasting period of volcanic reactivation during a post-rift phase of basin evolution (Pavelić 2001).

In the southern Dinarides, dacite–andesite tuff beds found in the Sinj Basin are interpreted as being derived from volcanic sources in Bosnia (Šušnjara & Šćavničar 1974).

Outer Dinarides: eastern Adriatic coast (V.Ć., T.M., K.D., A.M.)

The eastern Adriatic coast includes two units: the autochthonous part (Adriatic carbonate platform) and the overthrust nappe system (Dinaric thrust front), which is situated between the Dinaric Mountains and the Apennines, and includes convergent major structural units. Four particular domains can be distinguished (Figs 17.24 & 17.34): (1) Istria with the Ćićarija Mountains at the NW part of the coast; (2) the northern Adriatic (the coast with the islands of Krk, Cres, Lošinj, Rab, and Pag); (3) Ravni kotari (northern Dalmatia with the island of Dugi Otok and the Kornati archipelago); and (4) central to southern Dalmatia (the coast and the islands of Hvar, Brač and Korčula).

Palaeogeography

In terms of palaeogeography, the studied region belongs to two different tectonic settings: the Adriatic Carbonate Platform (AdCP) (*sensu* Herak 1986, 1999; Tari 2002) and the Dinaric nappe realm (Herak 1986, 1999; or frontal thrust of the western thrust belt of Tari 2002). According to the Eocene palaeogeographic reconstruction of Butterlin *et al.* (1993), the AdCP was situated at the northern fringe of the global desert belt, where

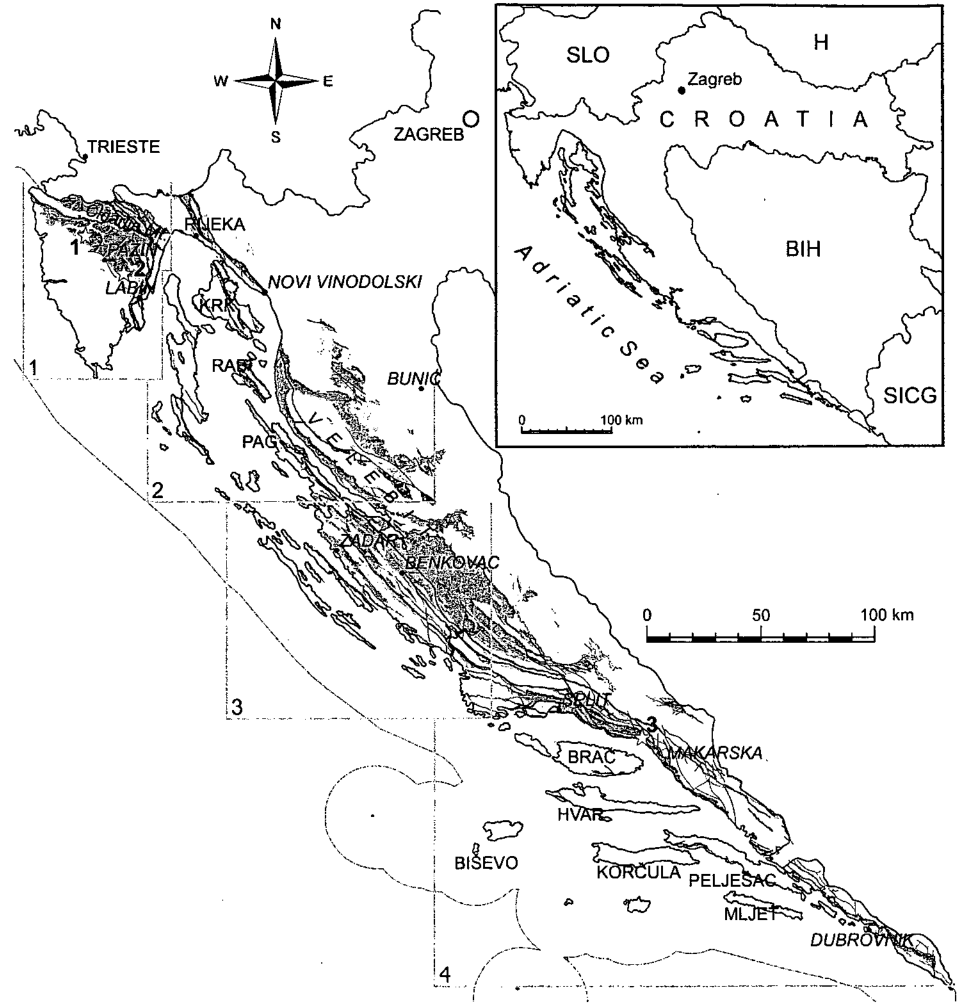


Fig. 17.34. Geological sketch map of the Croatian Outer Dinarides showing the location of principal Cenozoic zones (modified after Geologic map of Croatia 1:300 000, 1995). Zones of the Outer Dinarides: 1, Istrian peninsula and Ćićarija Mountains; 2, northern Adriatic coastal region with adjacent larger islands Krk, Cres, Lošinj, Rab and Pag; 3, Ravni kotari region with adjacent Kornati islands and Dugi Otok; 4, central to southern Dalmatian coastal area with adjacent islands Hvar, Brač and Biševo.

continental runoff was low. Larger benthic foraminifers indicate oligotrophic, warm-water conditions during Early–Middle Eocene times, and their development reflects the Eocene to Oligocene 'greenhouse'/'icehouse' transition (Moro & Ćosović 2002) The global Middle to Late Eocene cooling trend is shown by the planktonic foraminiferal associations of the deep-water deposits (Premec-Fućek & Živković 2005). Fossil flora (Jungwirth 2003) and coals from the Promina beds (Eocene–Oligocene) indicate wet vegetation conditions on land.

Carbonate platforms, such as the extensive AdCP, are an important element in the Outer Dinarides. Shallow-marine carbonate deposition commenced in the Late Triassic and lasted until the Lutetian transgression (Velić *et al.* 2003; Drobne 2003*b*, Vlahović *et al.* 2005). Carbonate deposition was terminated by the Eo-Alpine tectonic phase, which caused uplift and emergence of the Dinarides, resulting in the formation of significant bauxite accumulations (Istria, Ravni kotari). Post-Eo-Alpine sedimentation commenced diachronously from Ilerdian to Cuisian times (Bignot 1972; Drobne 1977; Ćosović *et al.* 1994; Matičec *et al.* 1996; Marjanac *et al.* 1998).

The Cenozoic history of the east Adriatic coast reflects the transition from a Cretaceous passive-margin setting (Dercourt *et al.* 1993) to a foreland basin evolution (Tari 2002). During the Early Eocene, a homoclinal carbonate ramp, the AdCP, developed. Palaeogene carbonate ramps were subject to short-term tectonic controls, including subsidence and sea-level changes. Inner and middle ramp deposits (Liburnian formations and foraminiferal limestones) are widespread along the coast, while outer ramp deposits are rare in the north Adriatic region and north Dalmatia. Deep-marine sedimentation commenced after a significant drowning of the Eocene carbonate platforms.

Tectonic setting

Numerous studies have examined the present-day structural architecture, tectonic evolution, and palaeogeography of the east Adriatic coastal region (Kossmat 1924; Petković 1958; Grubić 1975; Dimitrijević 1982; Pamić *et al.* 1998; Herak 1986, 1999; Picha 2002; Tari 2002). Although there is general agreement that the east Adriatic coast consists of (an) autochthonous platform(s) overthrust by nappes, the tectonic evolution is controversial. For instance, Pamić *et al.* (1998) assumed a uniform Adriatic–Dinaridic carbonate platform, while Herak (1986, 1999) distinguished the Adriatic and Dinaric platforms, and this is supported by biostratigraphic data (Drobne 1977, 2003*b*). Tari (2002) distinguished a gently tectonized Adriatic carbonate platform (Istria) with an imbricated margin (the so-called imbricated Adria) and the Dinaric nappe pile.

The Palaeogene development was characterized by several tectonic events. The first of these caused an unconformity between the Upper Cretaceous and Palaeogene strata (Pamić *et al.* 1998), and period of tectonic movements must have been responsible for the SW vergence at the east Adriatic coast. According to Pamić *et al* (1998), the second period of deformation took place during the latest Eocene to Early Oligocene. Picha (2002) related NW–SE trending strike-slip faults to the opening of the post-Messinian Albanian foredeep in the Pliocene–Holocene (the third deformational event), which indicates the existence of escape tectonics in the region. Tari (2002) suggested that folding and thrusting began in the Oligocene, when collision and progressive underthrusting of Adria beneath the Dinaric carbonate platform created imbricated structures at the margin of Adria; this process continued during the Miocene and thereafter. The typical NW–SE striking Dinaric structures were refolded as a result of neotectonic deformation (Marinčić 1997), producing west–east orientated structures, which are most prominent in the Central Dalmatian islands.

Palaeomagnetism

Palaeomagnetic studies were performed on Cuisian to Lutetian platform carbonates from Istria and central to south Dalmatia (Márton *et al.* 1995*a*, 2003*b*) as well as on deep-marine sediments from Istria. These proved the existence of an Adriatic microplate and demonstrate counterclockwise (CCW) rotation with respect to both the African and European plates in post-Eocene times. Istria must have rotated by 30° CCW, relative to Africa and stable Europe. The latest Miocene to early Pliocene CCW rotations observed in the Adriatic region were driven by that of the Adriatic foreland. This implies that the studied regions were rigidly connected and together participated in post-Eocene CCW rotation.

Sedimentary and stratigraphic development

Generally, the oldest Palaeogene deposits of the area (Fig. 17.35) are breccias overlain by freshwater or brackish and paralic limestones (Kozina-type limestones, part of the Liburnian Formation; Drobne & Pavlovec 1991) with local coal accumulations (Hamrla 1959). Fully marine conditions were established during the Ilerdian to Lutetian (Drobne 1974, 1977), when foraminiferal limestones were deposited on carbonate platforms (Magaš & Marinčić 1973; Korolija *et al.* 1977; Košir 1997; Ćosović & Drobne 1998; Marjanac *et al.* 1998; Drobne 2000; Velić *et al.* 2003; Ćosović *et al.* 2004*a*). These are overlain by the so-called Transitional Beds with the basal Marls with crabs and the overlying *Globigerina* ('*Subbotina*') marls. These sediments reflect the regionally diachronous deepening, which commenced during the Early to Middle Lutetian in Istria (Šikić 1965; Bignot 1972; Drobne 1977; Benić 1991; Ćosović & Drobne 1995, 1998; Ćosović *et al* 2004*a*) and during the Late Lutetian in Dalmatia (Pavšič & Premec Fuček 2000). Varying thicknesses were caused by orogenic processes (Šikić 1965, 1968, 1969). Clastic Eocene sediments of this area are traditionally referred to as flysch deposits (e.g. Marinčić 1981; Marjanac *et al.* 1998). The onset of deep-marine (flysch) sedimentation is diachronous, ranging from Cuisian–Middle/Late Lutetian (Živković 2004; Schweitzer *et al.* 2005) to Late Lutetian (Benić 1983), Bartonian (Benić 1983), Bartonian–Priabonian boundary (Krašeninikov *et al.* 1968; Marinčić 1981; Marjanac *et al.* 1998) and to Priabonian (Pavšič & Premec Fuček 2000). Sedimentation lasted until the Priabonian (NP18; Benić 1991), Bartonian (Benić 1983) and Late Oligocene–Early Miocene (Puškarić 1987; De Capoa *et al.* 1995). Deep-marine sediments are mostly overlain by the Eocene–Oligocene continental to shallow-marine sediments of the Promina Beds (Šikić 1965; Komatina 1967; Babić & Zupanič 1983), which are overlain by the Jelar breccia (Oligocene and younger) (Bahun 1974; Herak & Bahun 1979; Vlahović *et al.* 1999). During the Miocene, lacustrine sediments were deposited (Jurišić-Polšak 1979; Jurišić-Polšak *et al.* 1993; Šušnjara & Sakač 1988).

Istria. The geological literature on Istria (part of the AdCP) is plentiful, including two complex reviews by Bignot (1972) and Drobne (1977), and many overviews (Drobne 1979; Drobne *et al.* 1979; Hagn *et al.* 1979; Drobne & Pavlovec 1991; Drobne & Pavšič 1991; Ćosović 1991, Benić 1991, Marjanac 1991; Velić *et al.* 1995, 2002, 2003; Šparica *et al.* 2000). The carbonate depositional regime, which prevailed throughout the Mesozoic, terminated during the Early to Late Cretaceous (Drobne 1977; Matičec *et al.* 1996). Different Eocene successions transgraded over various Cretaceous units, which resulted in a high degree of lateral and vertical facies variability. The Liburnian Formation was deposited in the deepest parts of the palaeorelief. It is characterized by alternating freshwater to brackish, lagoonal Early Eocene deposits up to 80 m thick (Ilerdian to Cuisian; Drobne 1977). These are thin-bedded mudstones rich in organic matter and/or clayey mudstones, sporadically with stromatolites and coal intercalations (Labin Basin). Their fossil content, well-known since Stache (1889), includes Charophyta and molluscs.

The early Eocene to Middle Lutetian foraminiferal limestones in Istria (up to 200 m thick) were deposited in environments ranging from the restricted inner parts of the carbonate platform/

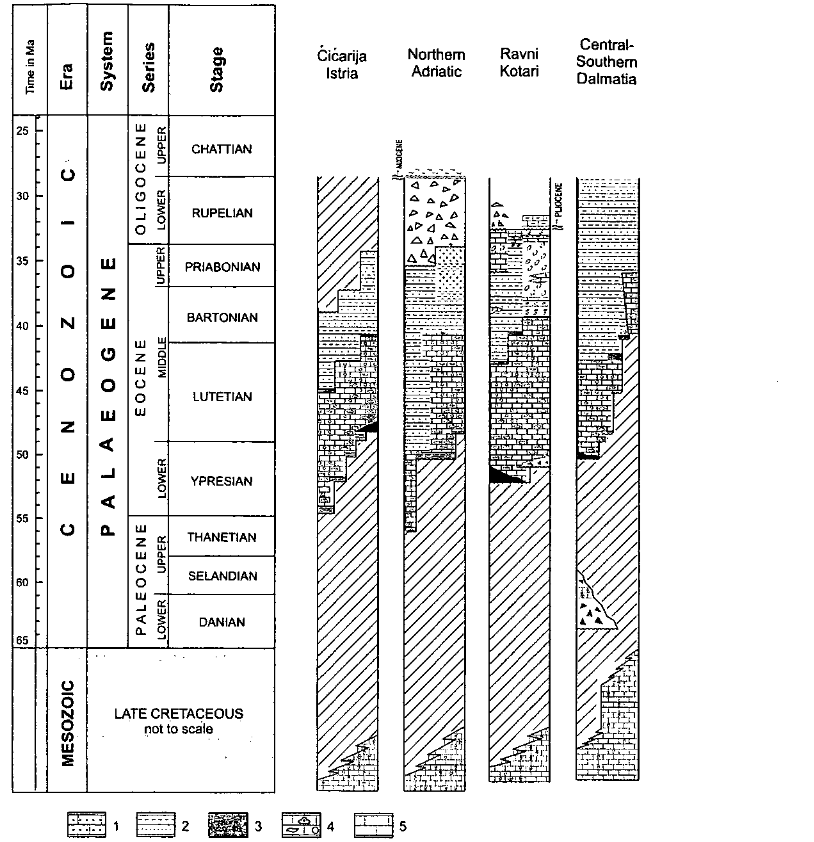


Fig. 17.35. Correlation chart of lithostratigraphic units from the Palaeocene to Pliocene in the different zones of the Croatian Outer Dinarides. Columns correspond to the zones of Figure 17.34. 1, Promina beds; 2, Flysch; 3, Transitional beds; 4, Shallow-marine carbonates (including Kozina-type of limestones, Foraminiferal limestones); 5, Shallow-marine limestones (zone 1 after Bignot 1972; Drobne 1977; Drobne *et al.* 1979; Benić 1991; Drobne & Pavlovec 1991; Ćosović *et al.* 2004*a*; Živković 2004; zone 2 according to Magaš 1968; Mamužić 1968; Mamužić *et al.* 1969; Bignot 1972; Drobne 1974; Benić 1983; Márton *et al.* 1995*a*; zone 3 according to Mamužić *et al.* 1970; Ivanović *et al.* 1973; Benić 1983; Drobne *et al.* 1991*c*; zone 4 according to Magaš & Marinčić 1973; Borović *et al.* 1976, 1977; Korolija *et al.* 1977; Benić 1983; Šušnjara & Sakač 1988; De Capoa *et al.* 1995; Marjanac *et al.* 1998; Pavšič & Premec Fuček 2000; Jelaska *et al.* 2003; Ćosović *et al.* 2004*b*).

ramp (Miliolid limestones), to shallower and deeper shoreface environments (with conical agglutinated foraminiferal, *Alveolina* and *Nummulites* limestones), and to deeper parts of relatively open carbonate ramps (Orthophragminae limestones). Facies patterns reflect a general deepening-upward trend. This was the consequence of intense synsedimentary tectonics providing appropriate accommodation space and a relatively low sedimentation rate. Sedimentation commenced during the Middle Ilerdian (SBZ7; Serra-Kiel *et al.* 1998) in the Ćićarija region, during the Cuisian (SBZ10–SBZ12) throughout Istria, and during the Early Lutetian (SBZ13) in its eastern part (Labin Basin; Bignot 1972).

The Transitional Beds range from shallow to deep-marine deposits. The basal Marls with crabs comprise <5 m thick nodular clayey limestones and calcitic marls (Schweitzer *et al.* 2005; Tarlao *et al.* 2005), and the upper parts consist of thick (a few to several tens of metres), massive *Globigerina* (*Subbotina*) marls with occasional intercalations of thin sandstone beds. These marls are rich in planktonic foraminifera and glauconite grains. They were deposited in Middle to Late Lutetian deeper-marine environments (bathyal hemipelagic deposits) (Ćosović *et al.* 2004*a*).

Deep-marine deposits crop out in the Trieste-Pazin Basin, Labin Basin, Mount Učka, and partly at Mount Ćićarija. These bathyal deposits (600 to 1200 m water depth; Živković & Babić 2003; Živković 2004) are up to 350 m (Marinčić *et al.* 1996). They were deposited in a structurally controlled basin, which was fed from the NW, NE, south and SE. The palaeotransport pattern was complex, although the central Pazin Basin shows a predominantly longitudinal transport direction with respect to the basinal axis (Babić & Zupanič 1996). The deposits are characterized by an alternation of hemipelagic marls and gravity flow deposits (Magdalenić 1972). The prevailing turbidite succession of hybrid carbonate/siliciclastic sandstones and marls is randomly intercalated with thick carbonate beds of debrite origin, i.e. megabeds (Hagn *et al.* 1979). The succession is characterized by two distinct stratigraphic units: the lower marl-rich unit with subordinate arenite interbeds, and an upper sandstone-rich unit. The deep-marine sediments have a stratigraphic range covering the Middle to Upper Eocene planktonic foraminiferal zones P11 to P15 (Berggren *et al.* 1995) or E9 to E15 (Berggren & Pearson 2005).

Northern Adriatic. The north Adriatic region (AdCP and Dinaric thrust front) includes the north Adriatic islands (Krk, Rab, Pag) and the adjacent coastal region (Rijeka hinterland and Vinodol area). Geological overviews are given in the geological maps of Ilirska Bistrica (Šikić *et al.* 1972), Delnice (Savić & Domazet 1984), Labin (Šikić *et al.* 1969), Crikvenica (Šušnjar *et al.* 1970), Cres (Magaš 1968), Rab (Mamužić *et al.* 1969), Lošinj (Mamužić 1968), Silba (Mamužić *et al.* 1970) and Gospić (Sokač *et al.* 1974).

The oldest Palaeogene sediments are the Liburnian deposits, followed by foraminiferal limestones and marly limestones with crabs. The Liburnian deposits, and the freshwater and brackish to marine limestones, are of Palaeocene age (Thanetian, SBZ 4 in Klana environs; Drobne 1974; Drobne & Ćosović 1998) and show a gradual transition to carbonate platform limestones composed of alveolinids, nummulitids, assilinids, operculinids and orthophragminids (SBZ11–SBZ13; Krk island and environs of Rijeka: Bignot 1972; Drobne 1977; Schaub 1981; Ibrahimpašić & Gušić 2000; Klepač 2003). They are overlain by mainly deep-marine sediments, the deposition of which started diachronously in Early Eocene (Rijeka hinterland; Bignot 1972), and Late Lutetian (P12 or E11, Rab and Krk islands; Benić 1983) times. The Lopar (Rab) sandstones are unique in the east Adriatic realm, because they represent shallow-marine siliciclastics. The facies patterns document sea-level changes and comprise sandy marls, heterolithic packages (sandstone–marl alternation), sandstones, calcarenites, conglomerates and slumps. The age of this succession is uncertain, because specimens resedimented from older strata prevail among the nannofossils (Benić 1983). The deep-marine sediments are overlain by the Jelar breccia of unknown age (Oligocene and younger) and by the Promina beds.

Towards the mainland, in the Lika region, Palaeogene outcrops are rare. The largest occurrence of Eocene limestones and deep-marine sediments is located at Bunić village (tectonic window after Herak 1986). Studies on alveolinids show a Middle to Late Cuisian age (SBZ11 and SBZ12; Drobne & Trutin 1997). Lacustrine sediments of Miocene age are found in the north Adriatic region (Pag island, Krbavsko polje; Jurišić-Polšak 1979; Jurišić-Polšak *et al.* 1993).

Northern Dalmatia. The area of Ravni kotari in north Dalmatia (AdCP and Dinaric thrust front) comprises Mesozoic platform carbonates transgressively overlain by Eocene foraminiferal limestones (Middle Cuisian to Middle Lutetian, SBZ10–SBZ14; Drobne *et al.* 1991*c*). Their contact is marked by breccias and locally by bauxite accumulations (Marković 2002). They are overlain by Transitional Beds, which are overlain by *c.* 850 m of deep-marine sediments (Ivanović *et al.* 1973). The Ravni kotari deep-marine deposits are Late Lutetian (NP16) to Bartonian (NP17; Benić 1983). The Bekovac section indicates that deposition extended from the Bartonian (plankton zone P13, E12) to the Priabonian (P16/17, E16: Drobne *et al.* 1991*c*) and was deposited in a deep-marine basin with mainly transversal palaeocurrent patterns. A close relationship of sandstone bodies (depositional lobes) with channels indicates a lower-slope or base-of-slope environment (Babić & Zupanič 1983). The latter is also indicated by the presence of olistolite-bearing megabeds with huge shallow-marine limestone clasts. The deep-marine sediments of north Dalmatia are overlain by the Promina Beds (sometimes referred to as a 'formation'; Ivanović *et al.* 1976) which are characterized by progressive shallowing of the basin. Babić & Zupanič (1983) located its base in a deep-marine olistostrome, and assumed an Eocene–Oligocene age for the entire formation. The lower part of the Promina Beds is characterized by an alternation of marls, sandstones, conglomerates, limestones and cherts, with freshwater to brackish-water fauna. Sedimentation of the upper Promina Beds is characterized by significant shallowing, which caused progradation of coarse alluvial sediments. These conglomerates contain marly beds with coal and plant remains. The Jelar breccia is interpreted as the youngest Cenozoic sedimentary unit in the area, but there are no direct contacts between the deep-marine units, the Promina Beds and the Jelar breccia. The age of the latter is uncertain, being possibly Middle Eocene–Early Oligocene (Herak & Bahun, 1979), Palaeogene–Neogene (Ivanović *et al.* 1973) or Late Lutetian–Bartonian (Sakač *et al.* 1993). Palaeogeographically, the Jelar breccia is an element of the Dinaric thrust front.

Central and Southern Dalmatia. Palaeogene deposits of this region (AdCP and Dinaric thrust front) primarily occur in tectonic contact with Mesozoic units, except for a few localities (Solin and Trogir hinterland). Previous studies were conducted in the framework of regional mapping (Borović *et al.* 1976, 1977; Herak *et al.* 1976; Korolija *et al.* 1977), stratigraphy and tectonics (Blanchet 1972, 1974; Chorowitz 1969, 1975, 1977; Šikić 1965) as well as sedimentology and stratigraphy (Benić 1983; Drobne *et al.* 1988*b*; Puškarić 1987; De Capoa *et al.* 1995; Marjanac 1996; Marjanac *et al.* 1998, Drobne *et al.* in press).

The oldest Palaeogene deposits are known from the Split–Makarska hinterland. These are Danian to Early Thanetian (P1–P3) limestones (Chorowicz 1977; Jelaska *et al.* 2003; Ćosović *et al.* 2006) which are overlain by Early–Middle Eocene (Ypresian to Lutetian) clast-supported breccias and limestones with larger planktonic foraminifers.

The Kozina-type limestones were deposited in more-or-less brackish, protected lagoons and bays during the Middle Cuisian (SBZ11), which were partly affected by severe subaereal exposure causing karstification. They comprise discorbidal limestones, miliolid limestones, stromatolites and breccias. Discorbidal limestones and stromatolites predominate in Hvar, while miliolid limestones are prominent on the Pelješac Peninsula.

The Kozina-type limestones were transgressively overlain by low-energy, open-shelf foraminiferal limestones. Their age ranges from Middle to Late Cuisian (SBZ11–12) in the central Dalmatian coastal region (Drobne 1985; Hottinger & Drobne 1980), from early Middle Lutetian (SBZ14) to early Late Lutetian (SBZ16) in the central Dalmatian islands, and from Early Lutetian (SBZ13) to Late Lutetian/Bartonian (SBZ17) in south Dalmatia. Massive Nummulite–Orthophragmina limestones indicate deposition in a quiet, well-aerated subtidal environment. Cross-bedded skeletal sands and imbricated larger foraminifers indicate the influence of major storm events.

The foraminiferal limestones are overlain by the so-called Transitional Beds, which comprise Marls with crabs and *Globigerina* (*Subbotina*) marls. Their thickness ranges from 11 m (central Dalmatian islands) to 15 m (south Dalmatia). The deposition of the Transitional Beds extends from the Late Middle Lutetian to the Late Lutetian and reflects a progressive deepening of the sedimentary environment.

Several hundred metres of deep-marine sediments were deposited above the Transitional Beds. The contact can be transgressive or transitional (Marinčić 1981; Šikić 1965). The stratigraphic range is probably from the Bartonian (NP17: Benić 1983) up to the Upper Miocene (De Capoa *et al.* 1995). A Priabonian age has been proven for the south Dalmatian deposits (Pavšič & Premec Fuček 2000). In central Dalmatia, the sediments are divided into Lower, Middle and Upper Flysch zones. The Lower Flysch Zone is about 750 m thick and dominated by megabeds. The Middle Flysch Zone is a *c.* 200 m thick olistostrome (Marjanac 1996). Within the 850 m thick Upper Flysch Zone, megabeds are restricted to the lower part; conglomerates interfingering with turbidites and alternations of thin-bedded calcarenites and calcirudites and marls occur throughout this unit. Coeval with the upper parts of the Upper Flysch Zone are Miocene lacustrine sediments on the island of Pag and inland in most of the karst poljes (Jurišić-Polšak 1979; Kochansky-Devidé & Slišković 1981; Šušnjara & Sakač 1988; Jurišić-Polšak *et al.* 1993).

Biševo Island extends far out into the Adriatic Sea and its fossil content represents a geological peculiarity. The Cretaceous–Cenozoic stratigraphy and fossil content of the island differs from those of the AdCP; instead the transgressive Oligocene sediments of the island (Drobne *et al.* 2000*a*) are similar to those of Cephalonia, Greece (Accordi *et al.* 1998). The Cenozoic succession of the island begins with a characteristic interval of reddish to yellowish clayey marls, alternating with thin layers of nodular limestones. The marls are overlain by limestones with *Nummulites fichteli* (SBZ 22, Rupelian–Chattian, *sensu* Cahuzac & Poignant 1997).

Dinarides and South Pannonian Basin in Bosnia and Herzegovina (S.Č., R.Red., H.H.)

The area of Bosnia and Herzegovina (Figs 17.2 & 17.26) is mostly mountainous, encompassing the central Dinarides. The NE parts extend into the Pannonian Basin, while the south partly borders the Adriatic Sea. Isolated Cenozoic sediments crop out all across the region. Sedimentary rocks prevail, but eruptive masses of dacite–andesite rocks occur in east Bosnia. During the Palaeogene, shallow-marine carbonates were dominant. They occur in two separate regions: in the Herzegovina area (Outer Dinarides) to the south and in Bosnia to the north. The Inner Dinarides (Bosnia) are composed of deeply weathered clastic, metasedimentary, metamorphic and igneous rocks. They included mostly Palaeozoic–Triassic rocks and the Dinaride Ophiolite Zone. The South Pannonian Basin (SPB) covers the Tuzla area (north Bosnia), while the Pannonian Basin *sensu lato* covers a large area at the southern margin of the Posavina area.

Palaeogeography and tectonic setting

The evolution of the region is related to the Palaeogene Eo-Alpine, Pyrenean and Sava orogenic phases, and subsequent Neogene neotectonic phases. Marine sedimentation prevailed until the earliest Oligocene in the Inner and Outer Dinarides, and was probably terminated by the orogenically induced rise of the middle Dinarides. A good example for the tectonic control is the synorogenic Promina Formation (Late Eocene–Early Oligocene) of Herzegovina, which developed during the Pyrenean Orogeny. The Eocene final deformation of the Dinarides resulted from underplating of Apulia beneath Tisia (the present-day Pannonian Basin terrains), which began during an Oligocene (32–28 Ma) period of transpression. This deformation gave rise to the final structure of the Dinarides. To the north of the uplifted Dinarides, strike-slip faulting produced a system of smaller and larger transpressional depressions, mainly orientated NE and ESE. During the Oligocene, shallow- to deep-water, marine and brackish to freshwater sedimentation took place in these depressions. Along the northern Dinarides margin, this dextral strike-slip faulting is seen in the incipient Sava and Drava transpressional faults, which apparently represent the ESE prolongation of the PAL (Pamić *et al.* 1998). Strike-slip faulting was accompanied by Early Oligocene volcanic activity, which took place in the northern Oligocene transpressional basins and along the northern margin of the Dinarides (e.g. area of Maglaj and Srebrenica). Oligocene strike-slip faulting was very active across the Dinarides area and is evident in the presence of transpressional depressions as precursors of Neogene freshwater basins. The largest transpressional faults are the Bosovača Fault, which controlled the origin of the Sarajevo-Zenica depression, and the Vrbas-Voljevac Fault, which controlled the origin of some smaller Cenozoic depressions. Within the AdCP, the Oligocene strike-slip faulting produced several larger and smaller karst valleys.

In the SPB, NE of the uplifted Dinarides, geodynamic processes changed fundamentally after the Oligocene transpressional deformation. Diapirism of the upper mantle and resulting attenuation of the lower continental crust are manifest in the extensional processes that gave rise to the evolution of the Pannonian Basin (18–17 Ma; Royden 1988). The extensional evolution of the SPB was mainly predisposed by the original Drava and Sava fault system, created during the Oligocene transpression. As a result of the extension, marine transgression occurred, and the first phases of the evolution of the Pannonian Basin took place under rift-related, mainly marine environments. This rift-related Early/Middle Miocene filling of the Pannonian Basin was accompanied by synsedimentary volcanic activity, which took place during the Karpathian and Badenian. The first rift-related stage of the evolution of the Pannonian Basin terminated by the end of the Sarmatian. This was followed by the second evolutionary stage, when continuous Late Miocene and Pliocene freshwater sedimentation occurred in the Pannonian Basin.

Palaeogene of the Outer Dinarides

In this zone, the Palaeogene rocks are found around Bihać, in Dinara, and between Livno, Mostar and Trebinje. Palaeocene sediments are restricted to the syncline core of Čardak Livada, in the Votorog Mountains (Papeš 1985). They are up to 400 m thick, contain characteristic Palaeocene foraminifers and algae, and are composed of breccias, marls, conglomerates and thick-bedded calcirudites.

Palaeocene–Eocene carbonates of the 50–300 m thick Liburnian Formation (= Liburnian Limestone; Čičić 1977) occur in the vicinity of Bihać, between Livno and Mostar, and from Stolac and Čapljina up to Trebinje. They are dark grey to black coloured beds. Locally, basal breccia limestones and calcarenites occur, while the upper levels are characterized by marly and platy, compact limestones. They discordantly overlie Senonian rudist limestones, separated by a bauxite unit.

Early and Middle Eocene sediments appear in Dinara, in the vicinity of Duvno and Lip Mountain, as well as in the area of Mostar and further towards the east. Sediments in the area of Duvno and Lip Mountain comprise *c.* 400 m of thick deep-marine sediments with larger planktonic foraminifers (Papeš 1985). At other localities, 100–600 m thick alveolinid-nummulitid limestones of the same age overlie the Liburnian Formation. Both of these successions are found in narrow zones related to reverse faulting and block overthrusting.

Middle and Late Eocene sediments of Herzegovina can be separated into two distinct successions: (1) marls, conglomerates, sandy stones, clays, slates, breccia limestones and limestones; and (2) limestone breccias, marls and sandstones with a thickness of 100–600 m and mainly deep-marine. Sediments of the first type are known from large areas on both sides of the Neretva river, from Posušje and Čitluk to Dabrica, and in the area of Lukavičko. They discordantly overlie the Liburnian Formation. Bauxite deposits of significant economical importance occur in the areas of Posušje and Dabrice. Mollusc lumachelles containing larger foraminifera are also found (Čičić 1977). Sediments of the second type are typically developed around Posušje, Tribistovo and Konjovac. They discordantly overlie alveolitid-nummulitid limestones and bauxites, and contain a rich mollusc, coral and echinid fauna. The succession is up to 200 m thick and considered as being coeval with the first succession.

Eocene–Oligocene units comprise conglomerates, calcarenites and marls. According to their typical development in the Promina Mountains (Dalmatia) they are called the Promina Beds (see above). They lie discordantly on Late Cretaceous and early Palaeogene sediments and are interpreted as terrestrial. The age of the Promina Beds is assumed to be Late Eocene to Early Oligocene. Their thickness decreases from Livno (1300 m), towards Posušje and Tihaljina (900 m), and to Nevesinje and Gacko (200–600 m).

Palaeogene of the Inner Dinarides

Various types of igneous rocks are known from the Dinarides (Katzer 1926; Varićak 1966; Pamić 1996). In Motajica, granites create the central part of the mountain. Six types of granite can be recognized: (1) massive, medium-grain-sized normal granite; (2) fine-grained granite with 'frozen edges'; (3) leucocratic granite; (4) greisenized; (5) aplicoide; and (6) kaolinized granites. Isotopic analyses suggested a Cretaceous–Palaeogene age, although Palaeocene–Eocene metamorphosis of the sediment cover suggests an Eocene–Oligocene age. Granite massifs also form significant parts of the base of the Pannonian Basin close to the Sava River.

Carbonate sediments dominate during Palaeocene to Early Eocene times. They are known from the broad area of Doboj and Tešanj, Derventa as well as in the areas of Trebovac and Majevica (Stojčić 1968). The microfauna of this 50–200 m thick succession comprises planktonic, benthic and larger foraminifera. In NE Majevica, sediments of this age are characterized by a 700–1500 m thick succession of slates, sandstones, marls and limestones. Sections from Tavna and north Majevica suggest that these sediments are part of a continuous Maastrichtian to Palaeogene deep-marine succession. The succession was dated by means of foraminifera and algae (e.g. Radoičić 1992). Over large parts of north Bosnia, deep-marine sedimentation continues up into the Middle Eocene. The thickness of these sediments is highly variable, ranging from 550 m (Kozara) to 1100 m (Majevica, Trebovac).

On Majevica and Trebovac, the lower part of these units is characterized by an alternating series of sandstones, slates and marls, with subordinate conglomerates and limestones. Marly sandstones, sandy limestones and limestones prevail at the top of the succession. The macrofauna was studied by Oppenheim (1908), Katzer (1918) and Čičić (1964).

Between the rivers Una and Vrbas, in the crest of Kozara and surrounding terrains, the so-called Kozara Flysch occurs. Palynological data (Čičić 1977) suggests an age of Early to Middle Eocene. It is composed of arkose and subarkose sandstones, as well as quartz-mica siltstones and rare conglomerates.

The Majevica Mountains comprise a 100 m thick Middle Eocene series of deep-marine sandstones, marls and limestones, which are overlain by marls that include a mixed marine and freshwater fauna (Čičić 1964), as well as slates and sandstones with limestone beds. The thickness of this succession in north Majevica is up to 200 m. Coal seams in these sediments are of local economic importance.

During the Middle and Late Eocene, the Majevica and Trebovac mountains comprise a series of 400–1200 m thick flysch-like sediments (Čičić 2002*b*). The lower parts are composed of sandstones and alevrolite, the middle parts comprise siltstones, marls, slates and rare limestones. The upper parts are characterized by thick-bedded marls, slates and sandstones, locally with conglomerate lenses. Sediments from the Vučjak and Trebovac mountains comprise thick-bedded sandstones, siltstones, conglomerates and slates with larger foraminifera, pollen and spores suggesting a continuous succession from the Late Eocene to the Oligocene (Ercegovac & Čičić 1968; Čičić 1977).

Miocene freshwater basins

A large number of lakes existed during the Miocene, which were initiated by tectonic movements during the Sava Phase. These created depressions, in which swamps and lakes formed due to the change from an arid to a humid climate. Milojević (1964) separated the lacustrine basins of Bosnia and Herzegovina into three regions: (1) north Bosnian region (Ugljevik, Mezgraja-Tobut, Priboj, Prnjavor, Lješljani, Banovići, Đurđevik, Seona and Jasenica basins); (2) middle Bosnian region (Miljevina-Rogatica, Sarajevsko-Zenički, Žepački, Teslić, Kotor Varoš, Banja Luka and Kamengrad basin); and (3) Outer Dinarides (Gacko, Mostar, Livno, Duvno and Cazin basins).

The largest lacustrine basins of Bosnia and Herzegovina were created within the main fault zones, e.g. Sarajevo-Zenica Basin in Busovača, Tuzla Basin in Spreča, Kamengrad Basin between the Grmeč and Sana fault, Mostar Basin in the zone of the Neretva Fault. Terrestrial sediments filling these basins have thicknesses of 1000–4500 m. Sedimentation in the larger basins began during the Egerian, but the exact age is unclear in many cases. In Ugljevik and other lacustrine basins of north Bosnia, the lacustrine regime lasted until the Karpatian marine ingression from the Pannonian Sea. In other basins, lacustrine sedimentation continued until the Pontian. In addition to coal seams, these basins contain marls, limestones, clays and conglomerates, and individual units of freshwater limestone can be more than 300 m thick.

Palaeontological information on the lacustrine sediments is sparse, but several biostratigraphic conclusions based mainly on mammals can be drawn, (Laskarev 1925; Milojković 1929; Malez & Slišković 1964 Pamić-Sunarić 1977; Malez & Thenius 1985). A palaeontological overview on Bosnia and Herzegovina has been published by Soklić (2001).

The Tuzla Basin

The Tuzla Basin is part of the South Pannonian Basin, which is situated to the north of the Dinarides mountain chain. This section summarizes the most important papers (Katzer 1918; Soklić 1955, 1959, 1964, 1977, 1982, 2001; Stevanović & Eremija 1960, 1977; Čičić 1977, 2002*b*; Čičić & Papeš 1970; Čičić & Milojević 1975; Čičić & Jovanović 1987; Čičić *et al.* 1988; Čičić & Redžepović 2003; Vrabac 1989, 1991, 1999; Ferhatbegović 2004).

Hydrocarbon exploration between 1934–50 provided data on the pre-Neogene basement of the Tuzla Basin. This consists of: (1) Upper Cretaceous clastics and carbonates; (2) Maastrichtian–Palaeocene–Lower Eocene deep-marine deposits; (3) Middle Eocene with foraminifers, molluscs and corals; and (4) Upper Eocene quartz sandstones and marls. Early Miocene sediments comprise the Slavinovići limestones, the Red and Mottled suites, and the Salt Formation. They are Aquitanian and Burdigalian in age (i.e. Late Egerian, Eggenburgian, Ottnangian and Karpatian).

The 60 m thick Slavinovići Limestone is characterized by desiccation cracks of dry lake-mud and the occurrence of freshwater gastropods. It is overlain by the Red suite, composed of redeposited tropical soils, derived from the peneplanated Dinarides. Silty sediments with secondary conglomerates with chert, peridotite and other silicate pebbles. The overlying Mottled suite, genetically related to renewed tectonic activity, was derived from the same source area. In addition to ophiolite rocks, pebbles of Tithonian–Valanginian limestones also occur. Volcanic activity in the adjacent Podrinje area caused the occurrence of ash layers. Sediments of the Mottled suite are rich in fossils, particularly gastropods. Pelite sediments of the suite are irregularly reddish, greenish and flaky, which is characteristic of muds flooded by alluvial fans. In the area of Soline, both the Red and Mottled suites are about 300 m thick. The *c.* 600 m thick Salt Formation was deposited within a marine lagoonal environment and shows a distinct cyclicity. The cycles begin with banded marls, followed by anhydrite and finally salt (halite and tenardite). Occasional volcanic ash layers can be up to 14 m thick. The Salt Formation is exploited at Tušanj (Tuzla town).

The overlying Karpatian sediments (>300 m thick) were deposited after a marine transgression from the Pannonian Basin (Central Paratethys). In the Ravna Trešnja-1 well near Tuzla, rocks of the Mottled suite comprise coarse-grained marine conglomerates with intercalations of clayey and slightly bituminous marls.

Karpatian sediments are conformably overlain by deep-marine Badenian marls intercalated with fine-grained sandstones. The marls contain foraminifers, pteropods and other molluscs, nautilids and echinoids, while Late Badenian coarse-grained sediments contain larger, thick-shelled, shallow-marine fossils. The total thickness of the Badenian sediments is about 500 m (e.g. deep-well Bukinje). They are covered by up to 300 m of Sarmatian brackish-water sediments comprising marls intercalated with sandstones and conglomerates. In contrast to the Badenian marine fauna, the brackish Sarmatian fauna comprises a low species diversity but a large number of individuals. The Pannonian fossils are characteristic of brackish environments.

In the Kreka Basin, six cyclothems with coal seams were deposited during the Pannonian. The cycles were caused by regressions within the Pontian Sea, which resulted in sandy sequences with increasing compositional maturity up-section. These processes continued until the final filling of the basin, after which freshwater swamps developed. Subsequently, these areas were covered by swampy forests, which were predisposed to coal generation. The accumulation of lignite and cellulose continued until the following marine ingression, which occurred as a result of sudden subsidence or, alternatively, of penetration of the barrier separating the basin from the Pontian Sea.

Pannonian Basin sensu lato

The Pannonian Basin *sensu lato* covers a large area in north Bosnia, between the rivers Una in the NW and Drina in the SE. The Late Miocene to Pontian period contains 200–300 m of brackish and limnic sediments rich in molluscs. The *c.* 100 m thick Pannonian sediments of the Sarmatian Sea comprise molluscs and ostracods. Sediments (170 m thick) of the brackish Sarmatian Sea show a characteristic succession: reefal olithic and sandy limestones, sandy marls, clayish marls with molluscs, foraminifers and ostracods (Soklić 1977, 2001). Further proven ages are Early Miocene–Ottnangian (sandstones and marls), Karpatian (marls and sandstones with Globigerinid foraminifers) and Badenian (sandstones and marls with foraminifers, molluscs and reefal limestones; Soklić 2001).

Summary

Geologically, the Cenozoic represents the period when Africa and Europe were converging, with seafloor spreading taking place in the Atlantic only as far north as the Labrador Sea (between Greenland and North America). Additionally, numerous microplates in the Mediterranean area were compressed as a direct result of Africa–Europe convergence, gradually fusing together. This resulted in a shift in the palaeogeography of Europe from a marine archipelago to more continental environments; this change was also related to the rising Alpidic mountain chains. Around the Eocene–Oligocene boundary, Africa's movement and subduction beneath the European plate led to the final disintegration of the ancient Tethys Ocean. In addition to the emerging early Mediterranean Sea another relict of the closure of the Tethys was the vast Eurasian Paratethys Sea. At the beginning of the Cenozoic, mammals replaced reptiles as the dominant vertebrates.

Central Europe is composed of two tectonically contrasting regions, namely a northern Variscan, and a southern Alpine Europe. The European Plate, a broad area of epicontinental sedimentation, essentially represented the stable European continent during the Cenozoic. The area, extending from the Atlantic shelves of Norway and the Shetland Islands through to eastern Poland and beyond, was separated from the Alpine-Carpathian chain by the Alpine-Carpathian Foreland Basin (= Molasse Basin) and its precursors as a part of the Palaeogene Tethys or Oligocene–Miocene Paratethys. The region encompasses the North Sea Basin, the Polish Lowlands, the Volhyno-Podolian Plate, the Upper Rhine Graben and the Helvetic units. To the south lay the Alps, a chain of mountains which formed during the multiphase Alpine Orogeny (see Froitzheim *et al.* 2008), and which can be traced eastwards into the Carpathians; this latter area is very different from the Alps mainly due to the presence of broad Neogene basins and extensive acidic to calc-alkaline volcanic activity. During the Palaeocene and Eocene, the Alpine system formed an archipelago. The North Alpine Foreland Basin was part of the Alpine-Carpathian Foredeep, a west–east trending basin located in front of the prograding nappes of the Alpine orogenic wedge. The Southern Alps, to the south of the Periadriatic Lineament, represent the northern extension of the Adriatic Microplate (together with the Eastern Alps, i.e. Austro-Alpine units). In Palaeogene times, they constituted the southern continuation of the Eastern Alps archipelago.

The Cenozoic history of Central Europe is chronicled in a series of Palaeogene and Neogene basins present across the region. In addition to the more stable North Sea Basin, the majority of these basins were strongly influenced by compressive forces related to the ongoing evolution of the Alpine chain. These forces resulted in general uplift of Europe during the Cenozoic. The marginal position of the seas covering the region of Central Europe and the considerable synsedimentary geodynamic control resulted in incomplete stratigraphic successions with frequent unconformities, erosional surfaces and depositional gaps. Additionally, during the Palaeocene, Europe–Africa convergence paused and a major hot-spot developed in the Faeroe–Greenland area. This mantle plume caused thermal uplift and associated volcanism across a broad area extending from Great Britain to the west coast of Greenland. At the Palaeocene–Eocene boundary, continental rupture occurred across this thermal bulge and ocean-floor spreading commenced between Greenland and Europe. Thus from Eocene times onwards, NW Europe became part of a thermally subsiding passive continental margin which moved progressively away from the hot-spot that continues today under Iceland.

Cenozoic times were characterized by a gradual long-term fall in global sea levels. This broad pattern was overlain by high-frequency and high-amplitude short-term changes related to polar glaciations and the repeated development of continental icesheets in subpolar areas of the northern hemisphere, which were broadly related to the closure of the Panama Isthmus (Ziegler 1990). Glaciation commenced in Miocene times, with regional icesheets present from the Pliocene. These eustatic changes are broady reflected in the sedimentary record of western and Central Europe. The Oligocene and Miocene deposits of the region are mainly found in the North Sea area in the north, the Mediterranean Sea region in the south and the intermediate Paratethys Sea and its late Miocene to Pliocene successor Lake Pannon. At its maximum extent, Paratethys extended from the Rhône Basin in France towards Inner Asia. Subsequently, it was partitioned into a smaller western part, consisting of the Western and Central Paratethys, and the larger Eastern Paratethys. The Western Paratethys comprises the Rhône Basin and the Alpine Foreland Basin of Switzerland, Bavaria and Austria. The Central Paratethys extends from the Vienna Basin in the west to the Carpathian Foreland in the east where it abuts the area of the Eastern Paratethys. Eurasian ecosystems and landscapes were impacted by a complex pattern of changing seaways and landbridges between the Paratethys, the North Sea and the Mediterranean, as well as the western Indo-Pacific. The geodynamic evolution of the region in Cenozoic times has resulted in marked biogeographic differentiation across the region. This has necessitated the establishment of different chronostratigraphic and geochronologic scales in order to facilitate cross-regional correlation.

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