Permo-Carboniferous climate: Early Pennsylvanian to Late Permian climate development of central Europe in a regional and global context

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Abstract: A well-justified stratigraphical correlation of continental successions and new palaeogeographic reconstruction of Pangaea reveal new insights into the northern Pangaean climate development influenced by palaeogeography, palaeotopography, glacio-eustatic sealevel changes and ocean currents. The overall Permo-Carboniferous aridization trend was interrupted by five wet phases. These are linked to the Gondwana icecap. The aridization and weakening of wet phases over time were not only caused by the drift of northern Pangaea to the arid climatic belt, but also by the successive closure of the Rheic Ocean, which caused the expansion of arid/semi-arid environments in the Lower/Middle Permian. The end of the Gondwana glaciation rearranged ocean circulation, leading to a cold, coast-parallel ocean current west of northern Pangaea, blocking moisture coming with westerly winds. The maximum of aridity was reached during the Roadian/Wordian. The Trans-Pangaean Mountain Belt was non-existent. Its single diachronous parts never exceeded an average elevation of 2000 m. The maximum elevation shifted during time from east to west. The Hercynian orogen never acted as an orographic east–west barrier, and the Inter-Tropical Convergence Zone was widely displaced, causing four seasons (dry summer/winter, wet spring/autumn) at the equator and a strong monsoon system.

The climate history of the European realm during the Late Carboniferous (Pennsylvanian) and Permian is stored in many solitary basins (Fig. 1) within the Hercynian orogen and the foreland basin. The story of Westphalian climate is well known because of the numerous investigations of the coal-bearing Variscan foredeep. The younger Westphalian is characterized by a slight aridization (Abbink & van Kronijnenburg-van Cittert 2003; Opluštil 2004), which was accompanied by an increase of seasonality. Nevertheless, the environment was strongly influenced by the ocean and epi-continental seas with multiple transgression events. The last extensive marine ingression was the Aegir/Mansfield Band (Westphalian B/C). The post-Westphalian climate development is more differentiated and not well known.

The Permo-Carboniferous climate of the central European realm is unquestionably dominated by an aridization trend (e.g. Chumakov & Zharkov 2002), which is not as simple as previously thought. Based on an improved stratigraphical correlation chart, several more humid phases are provable within the Late Carboniferous and the Early Permian. These so-called 'wet phases' are interpreted as a result of the waxing and waning of the Gondwana icecap. This is supported by the correlation of the continental European basins via isotopic ages to the Karoo Basin and adjacent areas. Further literature studies about this glaciation revealed numerous inconsistencies. Most important are the exact age, geographical position and size of the icecap. Nevertheless, the low-frequency cycles of this indubitably large glaciation are reflected in the continental sediments of northern Pangaea. This influence, and thus the strength of the wet phases, decreased during the Early Permian, although no decrease in the amplitude of waxing and waning of the icecap is reported. This discrepancy is solved by the new palaeogeographic reconstruction presented here of Late Carboniferous to Early Permian Pangaea (Fig. 2). The reduction of marine influence, which is essentially coupled to glacioeustatic sealevel, is based on the slow closure of the Rheic Ocean. In contradiction to the common palaeogeographic maps (e.g. Scotese 2001), this ocean persisted until the Middle to Late Permian. A second problem exists: the east–west barrier of the Trans-Pangaean Mountain Belt (Keller & Hatcher 1999) used in every modern (including altitudes) palaeoclimatic model, did not exist as a single orogen. Also, the height of individual mountain elements seems to be overestimated.

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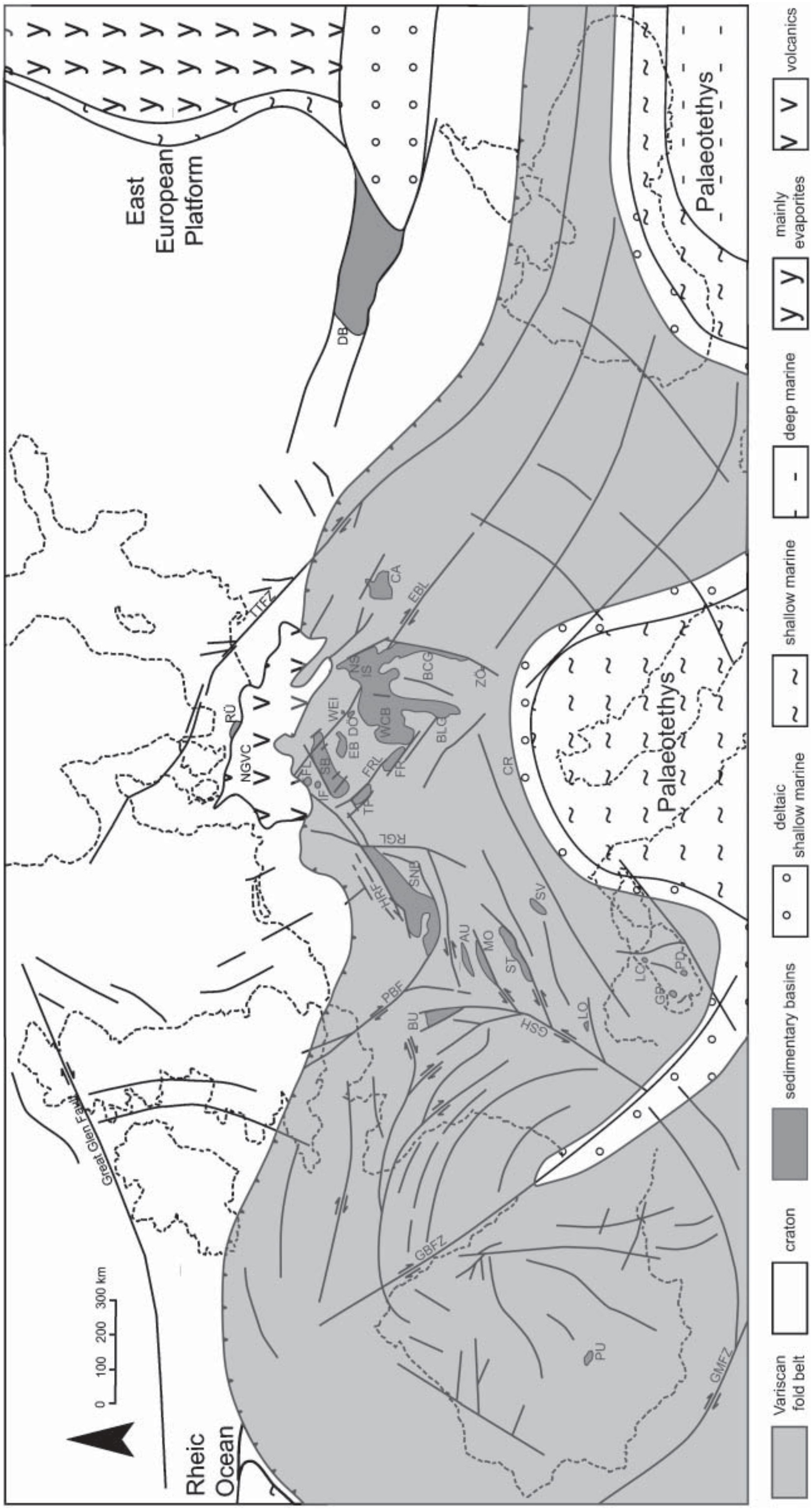
East European Platform
Palaeotethys
Palaeotethys
Rheic Ocean
Great Glen Fault
0 100 200 300 km
DB
CA
EBL
TTFZ
RÜ
NGVC
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NS
IS
BCG
ZO
WCB
EB
DO
FRL
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RGL
CR
SNB
HRF
SV
AU
MO
ST
PBF
BU
GSH
LO
LC
GP
PD
GBFZ
PU
GMFZ
Variscan fold belt
craton
sedimentary basins
deltaic shallow marine
shallow marine
deep marine
mainly evaporites
volcanics

Fig. 2. Palaeogeographic map at 300 Ma, grid not arranged to the palaeo-equator (adapted from Kroner, in Schneider *et al.* 2006).

These problems lead to a different reconstruction of palaeoclimate and its forcing mechanisms. The model presented here is based on analogies and climate sensitive sediments.

The European basins and their sedimentological development

Each of the Late to post-Westphalian European basins has its own tectonic, sedimentological and climatic history. This paper describes climate signals based on our personal research and extensive literature studies. All signals that could give a hint of at least regional climate changes are displayed in Figures 15a, b and 16. For the comparability of these datasets, a well-based stratigraphy (Fig. 15a, b) is indispensable. This is a balanced combination of isotopic ages and litho-, bio-, event and tectono- stratigraphy. The international scale is adopted from Menning & German Stratigraphic Commission (2002) with some modifications, especially concerning the Carboniferous–Permian boundary, which is set here at 299 Ma, according to Ramezani *et al.* (2003). Linking times between the basins in the Late Carboniferous to early Lower Rotliegend has an estimated error of ± 1 Ma. This relatively small inaccuracy for continental deposits was

Fig. 1. Geographic position of important Permo-Carboniferous basins, Northern and Southern Permian Basin are omitted. AU, Autun Basin; BLG, Blanice Graben; BCG, Boskovice Graben; BU, Bourbon l'Archambault Basin; CA, Carpathian Basin; CR, Carnic Alps; DB, Donetsk Basin; DÖ, Döhlen Basin; EB, Erzgebirge Basin; EBL, Elbe Lineament; FL, Flechting Block; FRL, Franconian Lineament; FR, Franconian Basin; GBFZ, Golf of Biscay Fracture Zone; GMFZ, Gibraltar Minas Fracture Zone; GP, Guardia Pisano Basin; GSH, Grand Sillon Houllier Fracture Zone; HRF, Hunsrück Fracture; IF, Ilfeld Basin; IS, Intra Sudetic Basin; KP, Krkonoše Piedmont Basin; LC, Lu Caparoni Basin; LO, Lodève Basin; MO, Montceau les Mines Basin; NGVC, North German Vulcanite Complex; NS, North Sudetic Basin; PBF, Pays de Bray Fracture; PD, Perdasdefogu Basin; PU, Puertollano Basin; RGL, Rhein Graben Lineament; SB, Saale Basin; SNB, Saar–Nahe Basin; ST, St. Etienne Basin; SV, Salvan-Dorénaz Basin; TF, Thuringian Forest Basin; TTFZ, Tornquist–Teysseyre Fracture Zone; WCB, Western and Central Bohemian Basins; WEI, Weissig Basin; ZÖ, Zöbingen.

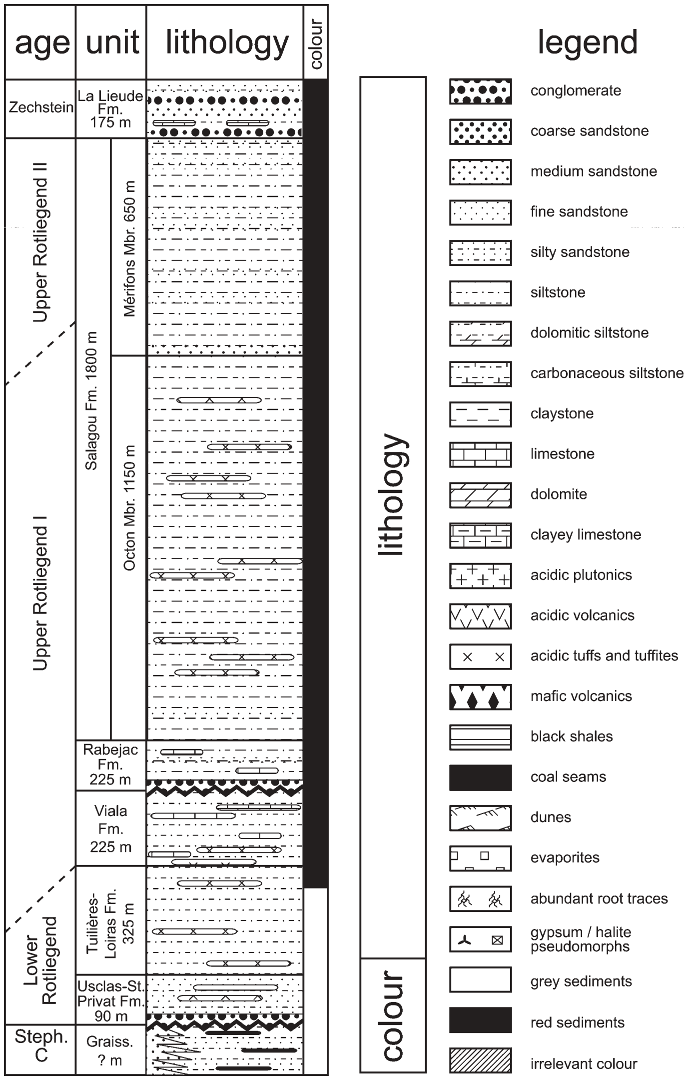


Fig. 3. General succession of the Lodève Basin.

reached by a combination of biostratigraphy (e.g. Schneider 1982; Hampe 1989; Werneburg 1999; Schneider *et al.* 2005*a*), geochronology (e.g. Hess & Lippolt 1989; Lippolt & Hess 1996; Goll & Lippolt 2001; Lützner *et al.* 2003; Lützner *et al.* 2006) and tectonostratigraphy (e.g. Stollhofen & Stanistreet 1994; Schneider *et al.* 1995*b*; Stollhofen *et al.* 1999), as well as lithostratigraphical correlations and comparison of sediment thicknesses (excluding volcanics) and sedimentary facies. For further information see Roscher & Schneider (2005). The connection to marine profiles, as well as to the international time scale, is based on isotopic dating and has a larger tolerance. All the climate indicative features discussed below are displayed in Figure 15a, b. By summarizing all these characteristics through space and time, at least several supra-regional climate events are obvious.

Lodève Basin

The sedimentation (Fig. 3) of the Lodève area (150 km^2) started in the Stephanian, unconformably above Hercynian structures, with grey clastics, several coal seams and lacustrine deposits in a terminal fan system (Gand *et al.* 2001). The overlying fluvial clastics and lacustrine black shales were deposited on an alluvial fan and by a floodplain system with eutrophic lakes in the basin centres. In the Upper Tuilières–Loiras Formation, the facies change from grey to red, as the environment changed from lacustrine to dominantly fluvial. The overlying late Autunian is characterized by a floodplain environment with sheetfloods and a braided river system with adjacent lakes. Above an erosional unconformity, fanglomeratic fan deposits follow. Further deposition took place in an alluvial plain/floodplain environment with periodically water-filled ponds grading into the playa sediments of the overlying Salagou Formation. Cycles of the Octon Member consist of 1-m-thick, massive, structure-less red-brown clayey siltstones (vertisols) and beige-coloured, 1-m-thick calcareous siltstones with characteristic desiccation cracks. Indications of a semi-arid to arid climate, such as vertisols (instead of calcisols) and desiccation crack horizons, became increasingly frequent. The occurrence of fossils is almost completely restricted to ephemeral channels that contain aquatic organisms adapted to dryness of seasonal or longer frequency (conchostracans, triopsids; cf. Gand *et al.* 1997*a*, *b*; Garric 2001).

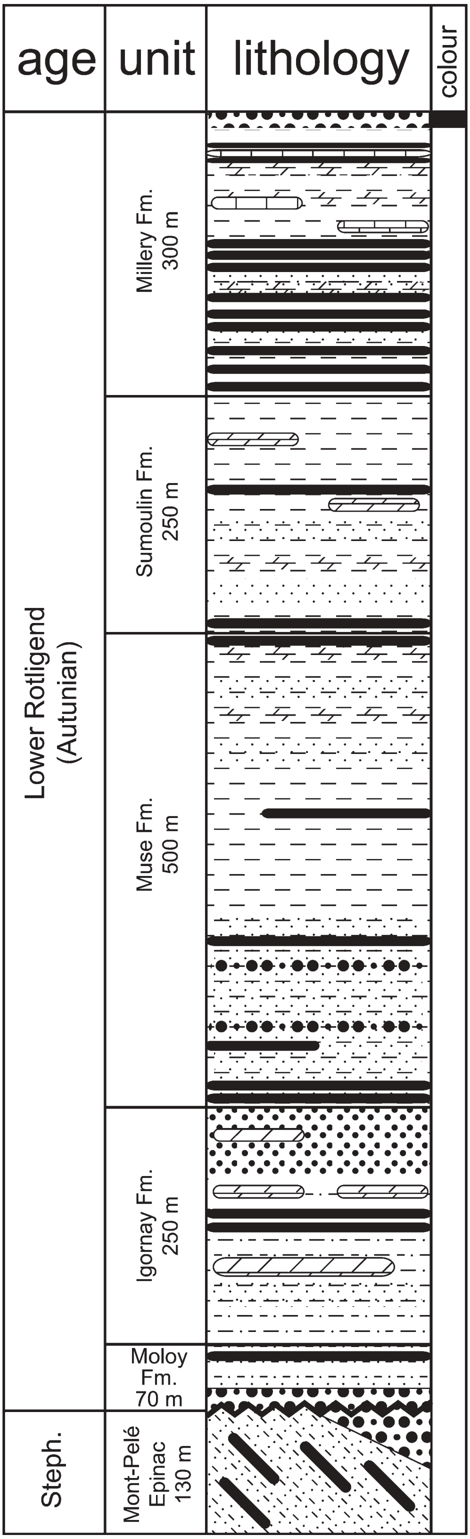
Sedimentary cycles of the Mérifons Member consist of centimetre- to decimetre thick, often laminated, red-brown siltstone and grey-green, 1-cm-thick siltstone with calcareous cements of a distal fan environment. Sand patch fabric and vertisols, as well as dewatering structures, indicate, periodic wetting and drying (cf. Hardie *et al.* 1978). The La Lieude Formation starts quite abruptly with sheetflood and braided river conglomerates, as well as large-scale cross-bedded, pebbly channel sandstones. Frequent greenish-whitish colours of calcareous-cemented sandstone layers and calcic soils indicate a higher ground water level, causing reducing conditions. Calcisols, invertebrate burrows of *Scoyenia*-type, root-penetrated siltstones, tree vegetation and a modern tetrapod track association (Gand 1993; Gand *et al.* 2000) appear within the first conglomeratic levels.

The Graissessac Formation is dated by Bruguier *et al.* (2003) to 295.5 ± 5.1 Ma (U–Pb) and the macroflora is ascribed to the late Stephanian by Doubinger *et al.* (1995). The Sakmarian (*Melanerpeton pusillum–Melanerpeton gracile* zone to *Discosauriscus austriacus* zone see Werneburg 1996, Werneburg & Schneider 2006) Usclas-St Privat Formation and the Tuilières-Loiras Formation were deposited in a semi-humid climate with seasonal rainfall indicated by laminated (varved) lake sediments and an absolutely conifer-dominated macroflora (85%; Galtier, in Lopez *et al.* 2005). The red facies of the Viala Formation (289.3 ± 6.7 Ma (U–Pb) (Schneider *et al.* 2006) were deposited under semi-arid conditions, as indicated by desiccation cracks, xeromorphic calcisols, vertisols and rare pseudomorphs after gypsum and halite crystals. The Rabejac Formation points to seasonal to episodic heavy precipitation events in an overall semi-arid climate. The overlying Octon Member (284 ± 4 Ma U–Pb, Schneider *et al.* 2006) of the Salagou Formation is conspicuous by the absence of plant roots and invertebrate burrows, such as *Scoyenia*, pointing to a lowered ground water level. The maximum of aridity was reached in the Octon Member, which is supported by the geochemical investigations of Körner *et al.* (2003) and Schneider *et al.* (2006). The overlying Mérifons Member is characterized by a semi-arid climate with fluvial and fan deposits. It is dated by insects and conchostracans to a Kungurian to Roadian age (Gand *et al.* 1997*a*, *b*; Bethoux *et al.* 2002). The base of the La Lieude Formation is supposed to be next to the Illawara Reversal (Bachtadse, pers. comm.). These drastic changes in litho- and biofacies patterns from dry playa to wet alluvial plain environments are indications of a rapid increase in the rate of precipitation. This is supposed to be an affect of the Late Permian Bellerophon and Zechstein transgression (Schneider *et al.* 2006).

The overall trend of the climate development within this basin is expressed by an aridization trend from the Stephanian Graissessac Formation above the red Viala and Rabejac formations to the playa environment of the Octon and Mérifons Member (Salagou Formation). After that the trend reverses. This humidization is not strong enough to allow the development of extensive lacustrine or swamp deposits, but the sediments of the La Lieude Formation indicate much more precipitation since the Artinskian than the underlying units do.

Autun Basin

Sediments of this basin (Fig. 4) started on Variscan granitic and metamorphic basement with the deposition of the grey conglomerates and sandstones of Mont Pelé and the intercalated, folded Epinac coal seams. Above the angular unconformity are coarse clastics with intercalated coal seams that, towards the basin centre, grade into varved carbonaceous shales of a palustrine to lacustrine environment. The overlying deposits are characterized by bituminous shales, sandstones containing layers of siderite, and dolomitic horizons of a lacustrine to fluvial environment. The frequency of lacustrine horizons decreases in the Surmoulin Formation, which is built up of a large sequence of grey mudstones with rare sandy, cross-bedded



dolomitic intercalations (Chateauneuf & Pacaud 2001). The subsequent Millery Formation is mainly composed of varved, dolomitic, lacustrine mudstones. This series contains 10 lacustrine bituminous horizons that are terminated by the Margenne boghead coal. Towards the top and the basin border the deposits grade into red coarser clastics, which were formerly described as the Saxonian Curgy Formation.

The sequence from the unconformity at the base of the Moloy Formation to the top of the Millery Formation is defined as the stratotype of the Autunian. The stratotype is of poor quality because of problems with the lower and upper biostratigraphical and lithostratigraphical boundaries (Broutin *et al.* 1999). The Autunian stage is based on macro- and microfloras. Because of the problems with the strong reliance of plants on climate, the correlation presented here is based on fossil fauna. The Muse Formation is dated to the *Melanerpeton sembachense–Apateon dracyiensis* or *Apateon flagrifer flagrifer–Branchierpeton reinholdi* zone, early to middle Asselian/early Lower Rotligend (Werneburg 1996, Werneburg & Schneider 2006) and *Syscioblatta dohrni–Sysciophlebia balteata* zone, early–middle Asselian/early Lower Rotliegend (Roscher & Schneider 2005). The Millery Formation contains teeth of *Bohemiacanthus* 'type Buxières', which is dated there (cf. page 101) to 289 $\pm$ 4 Ma (Pb/Pb) (Schneider *et al.* 2006) and the *Melanerpeton pusillum–M. gracile* zone, latest Lower Rotliegend (Werneburg & Schneider 2006).

This leads to various problems regarding the exact stratigraphical position of the Autunian. In brief, Broutin *et al.* (1999) published that the floras of the upper Goldlauter Formation of the Thuringian Forest Basin (cf. page 104) contain younger elements that do not occur in the Autunian of Autun, but the Goldlauter Formation, dated by amphibians, is older than the late Autunian Millery Formation. So, the dryness-adapted, modern Autunian floras, which do not occur at the stratotype, appear already in the mid-Autunian of the Thuringian Forest Basin. This indicates the sensitivity of plants to climate. The problems with the base of the Autunian are caused by the unconformity that separates the Autunian from the Stephanian as a structural unit (Chateauneuf & Pacaud 2001). This boundary could be used in the sense of a lithostratigraphical marker but not as a biostratigraphical one, which is necessary for stage definition. The base

Fig. 4. General succession of the Autun Basin. For legend see Figure 3.

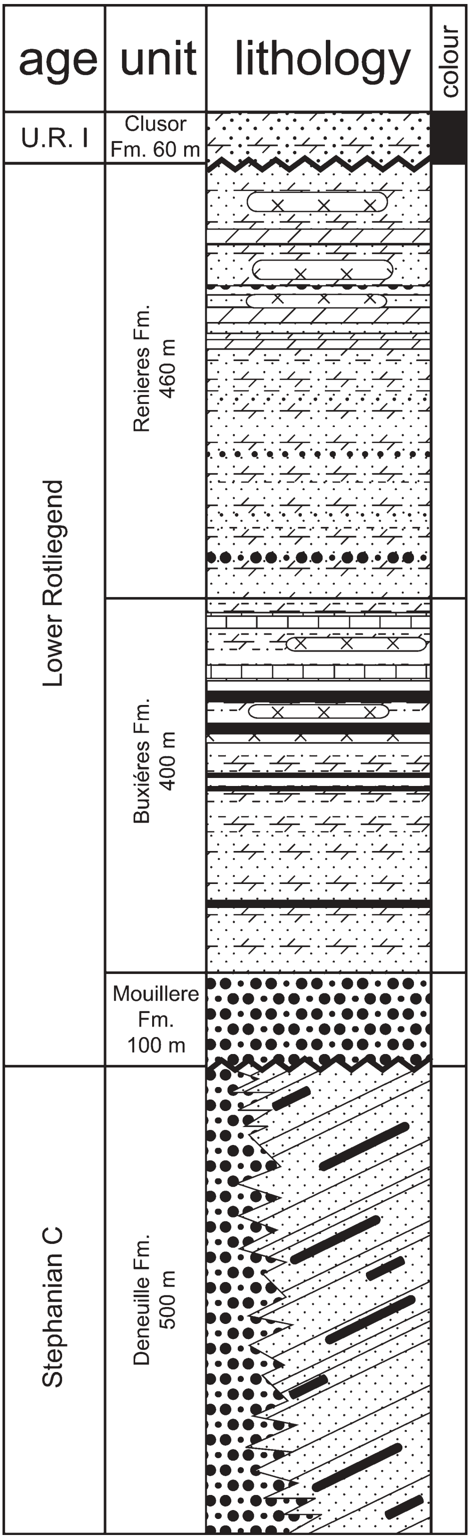


Fig. 5. General succession of the Bourbon l'Archambault Basin. U. R. I, Upper Rotliegend I. For legend see Figure 3.

of the Autunian is not defined by any index fossil (Broutin *et al.* 1999). Therefore, this term should not be used in the sense of a biostratigraphical unit, but can be used as a lithostratigraphical unit comparable, to or even equivalent, to the Lower Rotliegend. For further information see Schneider (2001).

The climate evolution of this depositional area is characterized by an aridization trend from the humid coal seams of Epinac and Moloy to the semi-humid fluvial to lacustrine, seasonally laminated grey clastics of the Muse and Surmoulin formations, which lack larger bituminous and carbonaceous sequences, to the semi-arid coarse red clastics of the upper Millery Formation. This trend is interrupted by the extended humid to semi-humid lacustrine sequence in the lower and middle Millery Formation.

Bourbon l'Archambault Basin

The sedimentation (Fig. 5) in this basin (500 km^2) started on crystalline (granitic) basement of the Massif Central with the alluvial-dominated Deneuille Formation (Steyer *et al.* 2000) with thin coal seams of a fluvial and marginal palustrine environment. The next sedimentary cycle started with alluvial arkoses and conglomerates (Steyer *et al.* 2000). A vertical transition to coarse dolomitic sandstones of fluvial origin is observed. They grade into palustrine coarse sandstones that contain coal seams, lacustrine bituminous clay and siltstones, and dolomites with approximately 30 partial basin-wide correlatable tuff and tuffite horizons (Paquette & Feys 1989). The overlying Renière Formation (Debriette 1997) consists of the fluvial-dominated Renière A Member (conglomerates, arkoses and silt- and claystones; Paquette & Feys 1989) and the lacustrine to palustrine Renière B Member, which is defined by the first occurrence of dolomitic horizons (Debriette 1993). It contains several tuff layers (Paquette 1980). The overlying Clusor Formation (Debriette 1997) is dominated by red sandstones and siltstones. The humid to semi-humid Deneuille Formation is ascribed to the Stephanian (Paquette & Feys 1989). The Buxières Formation is dated by Werneburg (2003) to the *Melanerpeton gracile–Discosauriscus pulcherrimus* zone (late Lower Rotliegend, Sakmarian) and to the *Sysciophlebia alligans* zone or *Sysciophlebia* n. sp. B zone in the sense of Schneider & Werneburg (1993). The coal

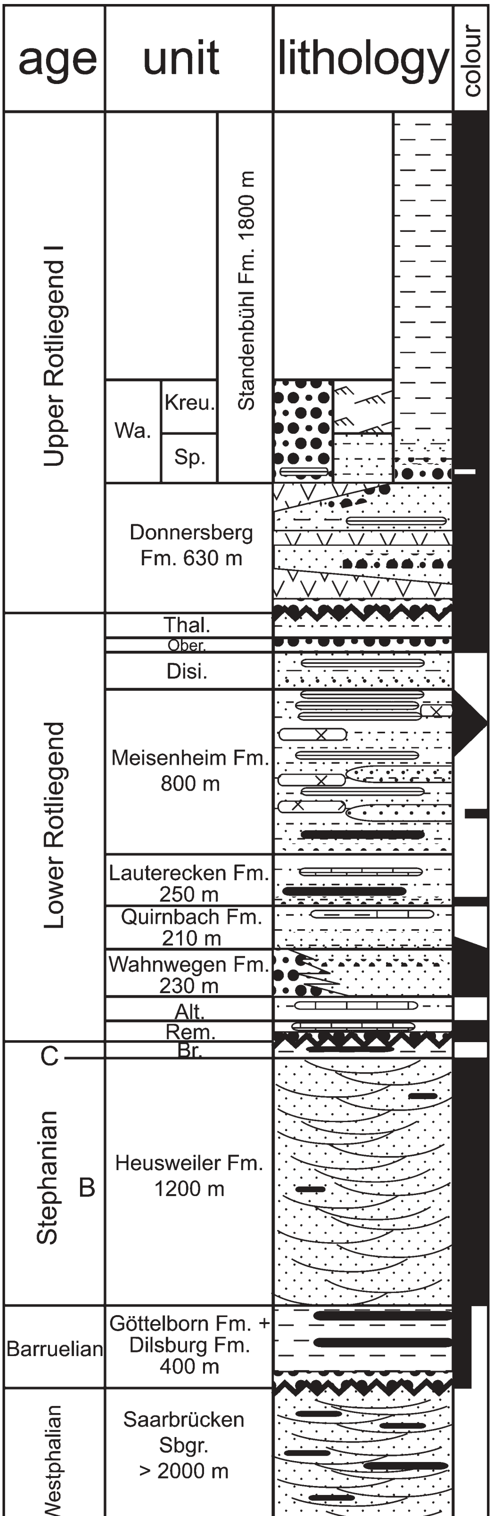


Fig. 6. General succession of the Saar–Nahe Basin. Br., Breitenbach Fm. 80 m; Rem., Remigiusberg Fm. 100 m; Alt., Altenglan Fm. 120 m; Disi., Disibodenberg Fm. 180 m; Ober., Oberkirchen Fm. 70 m; Thal., Thallichtenberg Fm. 120 m; Sp., Sponheim Fm. 240 m; Wa., Wadern Fm. 500 m; Kreu., Kreuznach Fm. 260 m. For legend see Figure 3.

seams of this formation show different palynological associations. The first is dominated by plants of a peat bog environment, whereas the second is dominated by a xerophilous floral association (Steyer *et al.* 2000). This indicates the development of local humid coal swamps within a semi-humid to semi-arid fluvial to lacustrine environment. The Clusor Formation is considered to be Saxonian (Upper Rotliegend) (Brulhet 1982). The climate development within this basin is characterized by an aridization trend from the coal-bearing Stephanian Deneuille Formation to the fluvial Renière Formation and the red beds of the Clusor Formation. This trend is interrupted by a short humidization event in the coal- and black shale-bearing, humid Buxières Formation.

Saar–Nahe Basin

The Westphalian to mid-Permian sediments of the *c.* 300 × 100 km Saar–Nahe Basin (Fig. 6) are subdivided into four subgroups. These are, from oldest to youngest, the Saarbrücken Subgroup, the Ottweiler Subgroup (Göttelborn–Breitenbach Formation), the Glan Subgroup (Remigiusberg–Thallichtenberg Formation) and the Nahe Subgroup (Donnersberg–Standenbühl Formation).

The whole Carboniferous section of this basin contains 140 named coal seams (120 are workable) that are concentrated in the Westphalian. The fluvio-lacustrine Saarbrücken Subgroup (about 3000 m thick: Müller & Konzan 1989) consists of rapid facies changes between grey clay, siltstones and sandstones with intercalated conglomerates and coal seams. Subsequent deposition started with a reddish conglomerate overlain by grey-green to reddish sandstones and claystones with several larger coal seams. This sequence is overlain by red, red-brown and violet-grey claystones, arkoses and conglomerates with various transitions. The environment is dominated by fluvial deposits (Boy 2003) with transitions at the top to lacustrine strata with coal seams.

The Remigiusberg Formation is characterized by alluvial red beds, fan conglomerates and fluvial to fluvio-lacustrine red clastics. Vertically, it grades into lacustrine grey to grey-green fine sediments with several intercalated bituminous limestones to black shales. The next cycle (Wahnwegen Formation) started with a prograding fan indicated by red, medium to coarse sandstones and conglomerates, and an environmental change from lacustrine to fluvial, again grading into grey fine clastics, with rare intercalations of lacustrine black shales and limestone horizons. Just above the basal conglomerate of the fluvio-lacustrine Lauterecken Formation, the up to 15-cm-thick, Odenbach carbonate coal seam spans an area of about 3300 km^2. Within the overlying Meisenheim Formation, approximately 40 tuff horizons occur in grey-brown to grey sandstones with sporadic regional coal seams. The fluvio-lacustrine middle part, with black shales and red sandstones, grade upward to lacustrine grey-brown to grey fine sandstones and pelites with subordinate red clastics (Stapf 1990). The overlying, fluvial-dominated red-grey conglomerates, arkoses and fine sandstones change with a fining-upward trend to red, yellow and grey sandstones and fine clastics. Above that, the Nahe Subgroup follows after a distinct hiatus (Stollhofen *et al.* 1999). It starts with the volcano-sedimentary Donnersberg Formation, consisting of red-grey conglomerates, arkoses and red to grey-green pelites that are strongly influenced by syn-sedimentary tectonics and volcanism (cf. Stollhofen 1994). The overall facies is described as fluvial red beds of a floodplain with meandering features and bituminous biolaminites of very small extent. The upper part of the Nahe Subgroup is represented in the west by playa-like pelites (indicated by the freshwater jellyfish *Medusina limnica*) of the Standenbühl Formation. The Wadern Formation in the northwest is dominated by red, alluvial-fan breccias and conglomerates. In the upper part, these fan deposits are interbedded with the red aeolian sandstone of the Kreuznach Formation. Importantly, in a fine clastic part of the Lower Wadern Formation, a last fish-containing lake horizon is intercalated.

Intensive investigations within this basin lead to a well-sustained stratigraphical correlation. The Westphalian Saarbrücken Subgroup is well dated with macro- and microfloras. Near the Carboniferous–Permian transition, the stratigraphical importance of floras decreases because not all European basins were in the same climatic belt at this time. Therefore, the stratigraphical correlation chart (Fig. 15a) is based on fauna and isotopic ages. The most important facts are summarised here (for further information see Roscher & Schneider 2005).

The Luisenthal Formation of the Saarbrücken Subgroup belongs to the *Archimylacris lubnensis* zone, Lower Westphalian D (Schneider *et al.* 2005*a*). The Dilsburg Formation is dated isotopically to 302.7 ± 0.6 Ma (Burger *et al.* 1997). The Stephanian A Göttelborn Formation belongs in the *Sysciophlebia* sp. A zone and the basal Stephanian B Heusweiler Formation in the *Spiloblattina pygmaea* zone (Schneider 1982; Schneider & Werneburg 1993; Schneider *et al.* 2005*a*). The Breitenbach Formation belongs to the *Branchierpeton saalensis–Apateon intermedius* zone (Werneburg 1996), *Bohemiacanthus* Ug zone (Schneider & Zajic 1994), *Sysciophlebia euglyptica–Syscioblatta dohrni* zone (Schneider & Werneburg 1993) and is geochronologically dated to 300 ± 2.4 Ma (Burger *et al.* 1997), which is Stephanian C. The Meisenheim Formation is dated to 297 ± 3.2 Ma (Königer 2000). The top of the Meisenheim Formation belongs to the *Melanerpeton pusillum–M. gracile* zone (Werneburg & Schneider 2006) and in the *Spiloblattina odernheimensis* zone (Schneider & Werneburg 1993; Schneider *et al.* 2005*a*). The youngest age is documented in the Wadern Formation, Sobernheim horizon, by the *Moravamylacris kukalovae* zone (Roscher & Schneider 2005).

The climatic development of this basin is marked by an obvious aridization, from the humid, coal-bearing grey sediments of the Saarbrücken Subgroup above the semi-humid wet red beds of the Heusweiler Formation, and the semi-humid to semi-arid red beds of the Wadern Formation to the semi-arid to arid playa sediments of the Standenbühl Formation. This overall trend is a multiphase one with several humid episodes in the Breitenbach, Altenglan, Disibodenberg and Wadern formations.

Thuringian Forest Basin

The Thuringian Forest Basin, an approximately 40 × 60 km NW–SE oriented depression, also called the SW Saale Basin, is one of the best, if not the best biostratigraphically investigated and correlated basin in the Variscan area (Schneider 1996; Lützner *et al.* 2003; Andreas *et al.* 2005). The basin is situated on the deeply eroded and peneplained Visean Thuringian Main Granite and the Ruhla Crystalline High of the inverted Mid-German Crystalline High (MGCH) at the outer border of the Variscan Orogen. This basin was at least partially connected to the Saale Basin by river systems (see below).

Sedimentation (Fig. 7) began with red (basin margin) and grey (basin centres) conglomerates and coarse arkosic sandstones followed by fluvial to lacustrine and palustrine fine clastics with fossiliferous lake horizons and thin coal seams of the Gehren Subgroup (Möhrenbach and

Georgenthal formations). Along intense fracture tectonic elements, the effusion of up to 1000 m of intermediate to acidic pyroclatics and lavas took place, and fluvial and muddy red beds that interfinger with laminated black shale lake deposits are intercalated. After a basin-wide erosional disconfomity, pyroclastics as well as rhyolites with minor intercalations of red-brown sediments were deposited. They grade into a fluvial-lacustrine grey facies with root-penetrated fine clastics and carbonaceous to sapropelitic, laminated lacustrine deposits. This is followed by nearly exclusively grey facies of fluvial deposits with a high ground water level. After a relief reactivation, red-brown alluvial fan conglomerates and fluvial sandstones grade into brownish to greenish fluvio-lacustrine and very finely laminated (varved) black shales of the depocentres intercalated by basin-wide tuff marker-beds. This sequence is overlain by interfingering red and grey facies with a high amount (up to 90%) of volcanics (acidic, intermediate and basaltic). Vertically it grades to sometimes playa-like red beds. The last perennial lake horizon of the Thuringian Basin is very widespread and grades laterally from calcareous, bituminous, varved black shales into red, varved, carbonateclay laminites (Schneider & Gebhardt 1993). After extensive erosion the exclusively red facies patterns show an alluvial fan to alluvial plain environment with temporary pools, root-horizons and sometimes playa-like (containing freshwater jellyfish *Medusina limnica*) clay- and siltstone of *Scoyenia*-facies.

With a shift of the depocentres to the north, again after a hiatus, the Tambach Formation follows. Facies patterns range from very coarse, matrix-supported wadi-fill conglomerates to proximal and distal debris-flow dominated alluvial fan clastics with fluvial, reworked aeolian sandstones, primarily accumulated as dunes on the top of the fans in the hinterland. *Scoyenia*-facies and complete bioturbation of *Planolites montanus*-type, indicative for wet red beds, is typical of these alluvial plain deposits. The flora consists of xerophilic walchians and cones of the drought-adapted *Calamites gigas*. Tambach is famous for complete articulated vertebrate skeletons, preserved in mud flows (Martens 1988; Berman & Martens 1993; Berman *et al.* 2000, Eberth *et al.* 2000). The fauna consist of reptiles and terrestrially adapted amphibians; fishes are

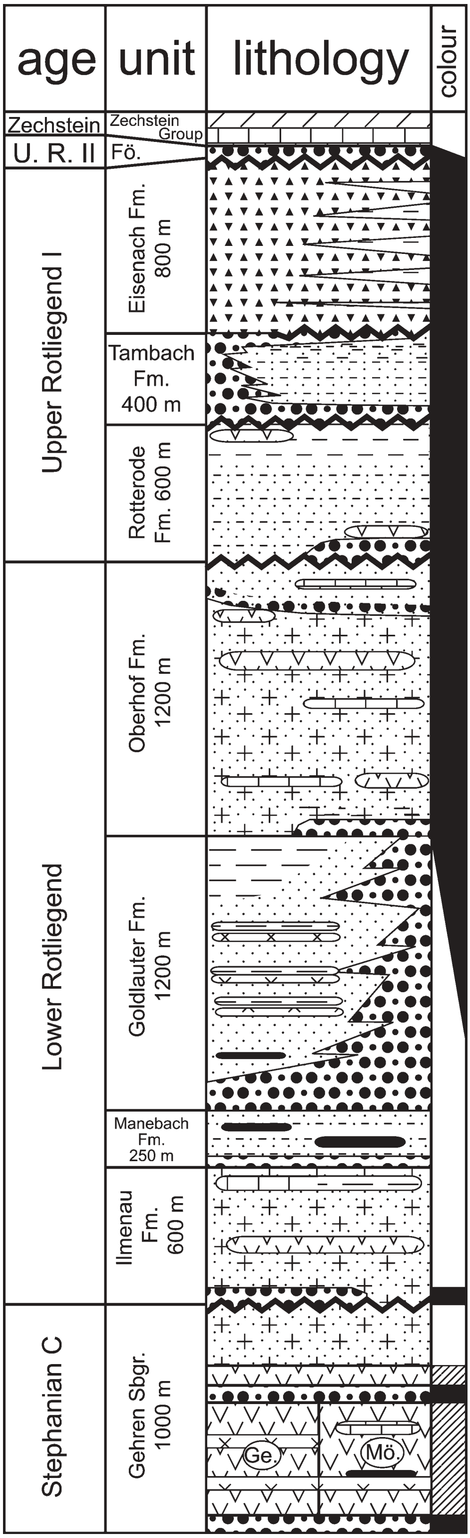


Fig. 7. General succession of the Thuringian Forest Basin. U. R. II, Upper Rotliegend II; Ge., Georgenthal Fm.; Mö., Möhrenbach Fm.; Fö., Förtha Fm. 20 m. For legend see Figure 3.

missing. The conglomerates and monotonous red, silty, sandy to clayey haloturbated siltstones with abundant millimetre-sized gypsum crystal casts in the top of this succession were deposited on an apron of alluvial fans with predominantly sheet flood deposits that, towards the basin centre, interfinger with fine clastics of playa mudflats. Well-rounded coarse sand and granule grains (2–3 mm) in the alluvial fan fine clastics are conspicuous. The playa-jellyfish *Medusina limnica* is common in claystones; ephemeral pond deposits contain conchostracans, and newly discovered are leaves of *Taeniopteris* sp. (Voigt & Rössler 2004). The youngest Permian continental sediments, presumably deposited just before the marine Zechstein transgression, are badly sorted, indistinctly horizontally stratified, matrix-supported debris flow conglomerates. The primary red colour changed to grey some metres below the marine Zechstein conglomerate; even granite pebbles are completely leached to pale grey. The horizontal nodule layers are regarded as ground water calcretes because no plant root structures are present. Schneider (1996) has interpreted this leaching and the ground water calcretes as effects of the marine pre-Zechstein ingressions and the Zechstein transgression (Wuchiapingian) into the Southern Permian Basin (see below, page 108), which caused a maritime print onto the strong arid continental climate. Higher precipitation rates and changing ground water levels triggered this calcrete formation and the leaching, which are observed at many outcrops along the southern coastline of the Zechstein Sea. Marine reworked coarse clastics and the Kupferschiefer (Copper slate) form the base of the Zechstein.

A short summary of the stratigraphy follows (for further information see Roscher & Schneider 2005). The Stephanian C age of the Möhrenbach Formation (in the SE) and the Georgenthal Formation (in the NW) is confirmed by biostratigraphy (Schneider & Werneburg 1993; Werneburg 1996) and radiometric data (Roscher & Schneider 2005; Lützner *et al*. 2006). The environment indicates a humid to semi-humid climate. The basal Rotliegend Ilmenau Formation was deposited after an extensive erosional hiatus and is dated, together with the overlying Manebach Formation, by amphibians to the *Apateon dracyiensis–Melanerpeton sembachense* zone (Werneburg 1996; Werneburg & Schneider 2006). The sedimentary structures, coal seams, lake deposits and the floral remains of both formations indicate a humid climate. The following semi-humid to semi-arid Goldlauter Formation is biostratigraphically dated to the middle Lower Rotliegend by amphibians, insects and sharks and geochronologically as 288 ± 7 Ma (Lützner *et al*. 2003, 2006). Facies architecture of the late Lower Rotliegend Oberhof Formation indicates a semi-humid climate, which gradually changed to a semi-arid climate at the top. The semi-arid wet red-bed and playa sediments of the Rotterode Formation are ascribed to the *Moravamylacris kukalovae* zone (Roscher & Schneider 2005).

The stratigraphical position of the overlying Tambach Formation is defined by the *Lioestheria monticula* zone (Martens 1987) = *L. andreevi* zone (Holub & Kozur 1981); *Seymouria sanjuanensis*, latest Wolfcampian (Berman & Martens 1993); *Dimetrodon teutonis*, early Kungurian (Werneburg & Schneider 2006). However, the tetrapod footprint assemblage of the Tambach Sandstone comprising *Amphisauropus*, *Ichniotherium*, *Dimetropus*, *Varanopus*, and *Tambachichnium* (Voigt 2005) and suggests an age for the Tambach Formation not younger than late Artinskian. A Sakmarian age is more likely, because the typical Artinskian ichnotaxa such as *Erpetopus* (Choza Formation, Texas), *Hyloidichnus* (Hermit Shale, Grand Canyon; Rabejac Formation, Lodève Basin) and *Dromopus palmatus* (Rabejac Formation) are missing in the Tambach Formation (Voigt 2005; Voigt, pers. comm. 2005). Based on the litho- and biofacies, the climate conditions are attributed to semi-humid to semi-arid.

The exact age of the arid to semi-arid playa-like Eisenach Formation remains unclear. The Förtha Formation is of Capitanian to Abadehian age (*Lueckisporites virkkiae*, *Corisaccites*: Kozur 1988). The overlying Zechstein Group is dated by *Merrillina divergens* in the first cycle (Werra) carbonates to early Wuchiapingian (Bender & Stoppel 1965; Kozur 1994) and *Mesogondolella britannica* in the Kupferschiefer, Wuchiapingian (Legler *et al*. 2005). The 257.3 ± 1.6 Ma Re–Os isotopic age of the Kupferschiefer at the base of the Zechstein was published by Brauns *et al*. (2003).

The climatic development of the Thuringian Forest Basin is marked by an aridization from the humid, grey, coal-bearing volcaniclastics of the Gehren Subgroup above the semi-humid Goldlauter Formation to the semi-arid to arid playa sediments of the Eisenach Formation. This trend reversed in the Upper Permian with the influence of the extensive Zechstein transgression. The aridization in the Late Carboniferous and Early to Middle Permian is interrupted by some humidization events, namely in the Oberhof and Tambach formations.

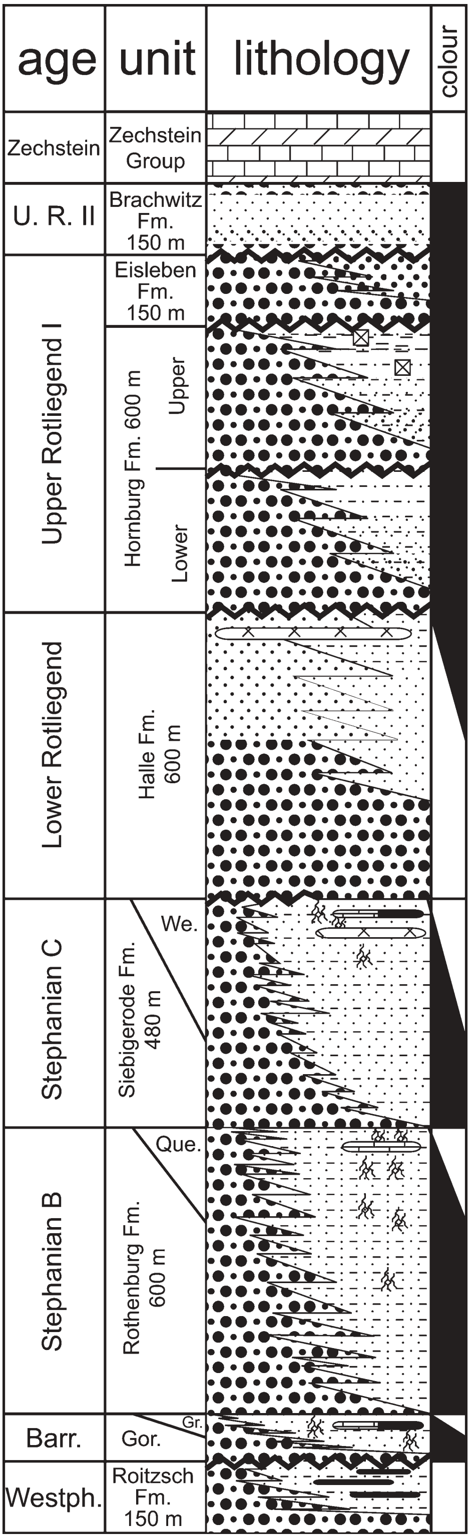


Fig. 8. General succession of the Saale Basin. U. R. II, Upper Rotliegend II; Gor., Gorenzen Fm. 60–100 m; Gr., Grillenberg Subfm. 20 m; Que., Querfurt Subfm. 150 m; We., Wettin Subfm. 300 m. For legend see Figure 3.

Saale Basin

The Saale Basin (*senso stricto*) is a continental basin of 150 × 90 km (Schneider *et al.* 2005*b*). The underlying Visean to Westphalian sediments (Steinbach & Kampe pers. comm. 2005) are known only from drill cores and belong to the southern border of the Variscan fore-deep (Gaitzsch 1998). The sediments of the Roitzsch Formation (Fig. 8), unconformable on Namurian A sediments, were deposited in a drainage system that started at the NE end of the Central Bohemian Basin (Opluštil & Pešek 1998) and ended in the Variscan foredeep (Gaitzsch *et al.* 1999). Above basal coarse clastics, conglomerate/rooted soil cycles with several coal seams, several decimetres to 1 m thick, and fine clastics follow. The Stephanian basin-fill rests disconformably on older sediments and Variscan metamorphics and consists mainly of wet red beds (*Scoyenia*-facies) of proximal to distal alluvial fan and alluvial plain environments of the Mansfeld Subgroup (Gorenzen Formation to Siebigerode Formation). Grey sediments of fluvial, lacustrine and palustrine facies, designated as subformations, are found in depocentres near the top of each megacycle. They contain several impure coal seams and carbonaceous horizons. The widely distributed wet red beds (*Scoyenia*-facies) comprise alluvial fan/sheetflood/braided river associations with immature to mature calcisoils and metre-thick calcretes and decimetre-thick greyish to reddish micritic limestones of ephemeral lakes. At the top, the change of channel geometry indicates a transition from braided to meandering rivers, and the grey facies is vertically and laterally more widespread and includes lacustrine limestones and black shales in the flood basins, as well as palustrine deposits of back-swamp environments that contain several workable coal seams.

The fish faunas of the Querfurt and the Wettin subformations indicate that the Saale Basin was interconnected with an extensive European drainage system that was destroyed at the beginning of the Rotliegend by volcano-tectonic reorganization of the basins (Schneider & Zajic 1994; Schneider *et al.* 2000). This event is indicated in the Saale Basin by the sudden deposition of the 'Kieselschiefer-Quarzit-Konglomerat' (Chert-Quartzite Conglomerate) at the base of the Halle Formation, which grades vertically into reddish sand and siltstones. After a long hiatus, the completely red alluvial fan to alluvial plain conglomerates and sandstones as well as occasional sheet flood sediments with interfingering siltstones and claystones were deposited. These fine clastics represent playa deposits characterized by *Medusina limnica*, millimetre-large halite hopper crystals and metre-deep desiccation cracks. Very noticeable features are aeolian deposits – dune sandstones (flooding of dunes during playa-lake formation resulted in strong deformation of these dune sandstones; Schneider & Gebhardt 1993) as well as dry sandflat sandstones with lag deposits of wind-transported, coarse, well-rounded grains. The latter may also have been re-deposited by flash floods, forming decimetre-thick sequences of bimodal, coarse-grained sandstones with fine-grained, well-sorted matrix of primary aeolian origin. This is capped by a sequence of red conglomerates, sandstones and siltstones of proximal to distal alluvial fans.

After a hiatus, the braided river, sheetflood and wet sandflat deposits of the Eisleben Formation, which is regarded as equivalent to the Dethlingen and Hannover formations of the Southern Permian Basin (Legler *et al.* 2005), were deposited. In the area of the Saale Basin, the Zechstein commonly starts with marine re-worked aeolian sands and above them the Kupferschiefer (Copper slate).

The stratigraphical correlation of this basin is based on the following data: the Wettin Subformation (Siebigerode Formation) belongs to the *Apateon intermedius–Branchierpeton saalensis* zone (Werneburg 1996; Werneburg & Schneider 2006), *Bohemiacanthus* Ug zone (Schneider & Zajic 1994), and *Sysciophlebia euglyptica–Syscioblatta dohrni* zone (Schneider & Werneburg 1993) and is geochronologically dated to 293 ± 2 Ma (Goll & Lippolt 2001). The overlying Halle Formation belongs to the *Apateon intermedius–Branchierpeton saalensis* zone (Werneburg 1996; Werneburg & Schneider 2006) and is dated by Breitkreuz & Kennedy (1999) as $297–301 \pm 3$ Ma. The conchostracans of the Upper Hornburg Formation belong to the *Lioestheria andreevi–Pseudestheria graciliformis–Palaeolimnadiopsis wilhelmsthalensis* assemblage zone (Hoffmann *et al.* 1989; Schneider *et al.* 2005*a*). The Eisleben Formation is lithostratigraphically correlated to the Hannover Formation of the North German Depression (see below).

The climate development of the Saale Basin demonstrates an aridization trend, starting with the humid, grey, coal-bearing sediments of the Gorenzen Formation, through the semi-humid, grey sediments of the Halle Formation, to the semi-arid to arid playa-like red clastics of the Upper Hornburg and Brachwitz formations. This trend is stopped at least by the sedimentation of the epi-continental Zechstein Sea and interrupted by the Stephanian C semi-humid to humid Siebigerode Formation.

North German/Polish Depression

The Central European Basin (CEB) is one of the largest basins on earth. It provides one of the most voluminous data sets for understanding the evolution of similar giant intracontinental accumulations of sediments (Fig. 9). The history of this basin started with formation of the Middle to Late Permian continental megaplaya/megasabkha system of the Southern Permian Basin (SPB). Early to Late Carboniferous sediments in the area of the SPB belong to the Variscan foredeep, which ceased during the Stephanian. Here, sediments from Late Westphalian (comparable to the Ruhr area: Hoth *et al.* 1990) up to the Zechstein base in the eastern part of northern Germany are considered. They comprise a grey, coal-bearing paralic facies (with coal seams up to 1.8 m thick) with a last extensive marine incursion (Aegir or Mansfield marine band) at the Westphalian B/C (Duckmantian/Bolsovian) boundary. After that, the facies changed from grey sediments to wet red beds. The uppermost thin root horizons and a thin coal seam are restricted to the basal Westphalian D.

The red, conglomeratic Mönchgut Formation is overlain by huge volcanic complexes, up to 3000 m thick, with small basins adjacent to and inside them. They are filled with red fan-conglomerates and variegated, red to grey alluvial plain sediments with laminated (varved) grey to reddish bituminous lacustrine limestones as well as tuff and tuffite horizons (Gaitzsch 1995*a*; Schneider *et al.* 1995*a*). Laminated lacustrine black shales, red alluvial plain to ephemeral lacustrine fine clastics, locally replaced by alluvial fan conglomerates, overly this sequence.

The SPB originated by thermal subsidence and the start of post-Variscan rifting processes linked with extrusions of upper mantle basalts (Gebhardt *et al.* 1991). The basin extended from England over the southern North Sea and northern Germany to Poland with a length of 2500 km and a width of 600 km. It was filled by about 2500 m of siliciclastics and evaporites of the Upper Rotliegend II, as well as 2000 m of siliciclastics, carbonates and evaporites during the Zechstein (Ziegler 1990*b*). The Upper Rotliegend II basin fill is dominated by desert sediments affected by an arid to semi-arid climate. Alluvial fans and dunes occur, especially at the southern basin margin, whereas saline

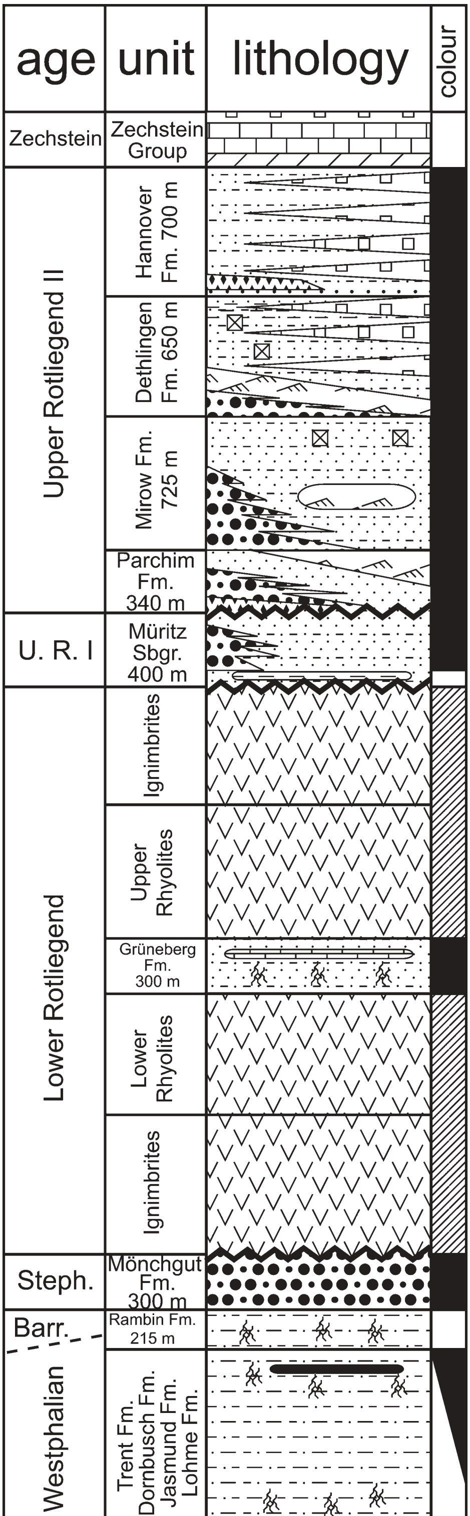
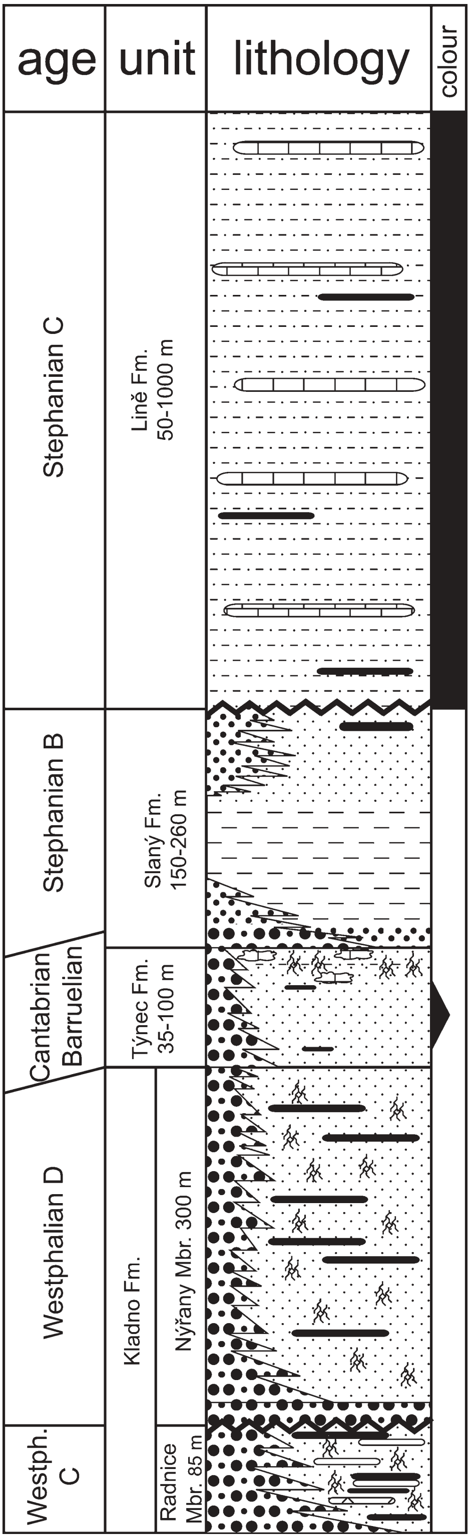


Fig. 9. General succession of the North German/Polish Depression. U. R. I, Upper Rotliegend I. For legend see Figure 3.

lake deposits dominate in the centre. The sedimentation was controlled tectonically as well as climatically, as is reflected by the lithostratigraphical subdivision. Tectonically driven large-scale cycles are interpreted as formations, whereas the members are climatically governed cycles on a smaller scale (Gast 1991, 1993, 1995; Gaupp *et al.* 2000). Basaltic extrusions, which are well known from the Altmark region, the Soltau High and the Horn Graben, are signs of rifting as an important controlling parameter in the evolution of the basin (Gebhardt *et al.* 1991; Gebhardt 1994; Stemmerik *et al.* 2000). Short-term marine ingressions superpose the climatically driven, coarsening-upward cycles of the Hannover Formation (Gebhardt 1994; Legler 2005). The Zechstein Transgression was a sudden flooding event triggered by a sea level highstand and rifting in the Viking Central Graben system (Smith 1979; Glennie 1989). Eustatic sealevel highstands (0.8 to 1 Ma cycles observed in the Proto-Atlantic: Legler 2005) before and together with the Zechstein transgression, as well as the more or less coeval *Bellerophon* transgression, put a maritime print on the otherwise strongly continental climate of northern Pangaea.

The stratigraphy of this basin is based on isotopic data and individual biostratigraphical constraints. The Lower Ignimbrite was dated by Breitkreuz & Kennedy (1999) as 302 ± 3 Ma. The sediment intercalations within the volcanics belong to the *Bohemiacanthus* Om–Ugo zone and the *Pseudestheria paupera* and *Pseudestheria palaeoniscorum* Zones, middle Lower Rotliegend (Gaitzsch 1995*b*, *c*). The Müritz Subgroup belongs to the *Lioestheria andreevi–Pseudestheria graciliformis–Palaeolimnadiopsis wilhelmsthalensis* assemblage zone, Upper Rotliegend I (Hoffmann *et al.* 1989; Schneider *et al.* 2005*a*). The age of the Parchim Formation is fixed by the Illawara Reversal in its basal portion (Menning 1995). In this basin, the conodont *Mesogondolella britannica* was found in the Kupferschiefer (Copper slate) and demonstrates a Wuchiapingian age (Legler *et al.* 2005).

The climatic development of this basin is marked by an overall aridization trend from the grey, coal-bearing Westphalian sediments, via the coaly wet red beds of the Mönchgut Formation, to the wet red beds of the Müritz Subgroup and the Upper Rotliegend II playa/sabkha red beds of the Elbe and Havel Subgroup. With the pre-Zechstein transgressions the environment



became more humid and, finally, marine with the Zechstein transgression.

Central and western Bohemian basins

The formerly connected basins (140 km × 25 km) of western and central Bohemia are (beginning from the west): the Plzeň Basin, Manětín Basin, Radnice Basin, Žihle Basin, Kladno-Rakovník Basin and Mšeno-Roudnice Basin. The basement consists of Late Proterozoic, weakly metamorphosed sediments and volcanic units with subordinate, medium to high grade metamorphic rocks (Pešek 2004). The basal sediments (Fig. 10) are characterized by fine- to medium-grained, grey, green or red-brown unbedded sediments, with matrix-supported conglomerates of an alluvial fan, to alluvial plain environment on uneven bedrock. These environments grade vertically to extensive peat bogs in a broad alluvial plain with many coal seams, often separated by numerous tuffitic layers. A conglomerate complex was deposited after a hiatus. After that the facies changed slowly, as indicated by the decrease of coal seams and sediment colour change to red-brown. Overlying are fluvio-lacustrine grey conglomerates to sandstones with rare, intercalated greenish or brownish red mudstones. The following biogenic laminated and varvite-like claystones are ascribed to a lacustrine environment of an extensive freshwater lake with a depth of several tens to some hundred metres. The stronger fluvial influence and shallowing of the environment is indicated by an increase of coarser sediments (grey to reddish violet sandstones) and culminates with the reappearance of coal seams, up to 1 m thick, that formed at the lakeside. After a depositional break, mostly red to deep red siltstones with a low coal content were deposited. The facies with several calcareous horizons is ascribed to an alluvial plain with a system of small, interconnected lakes.

The stratigraphy of these deposits is determined by various methods. The Radnice Member is dated by macroflora (Pešek 2004) and insects (*Archimylacris lubnensis* zone, Schneider *et al.* 2005*a*) to late Bolsovian to early Westphalian D, and has a high amount of coal, indicating a warm and humid climate. For the Nyřany Member, Werneburg (1989, 1996, 2001) indicated, by amphibians, the *Branchiosaurus salamandroides–Limnogyrinus elegans* zone (Westphalian D to Stephanian B). Climatically,

Fig. 10. General succession of the Western and Central Bohemian basins. For legend see Figure 3.

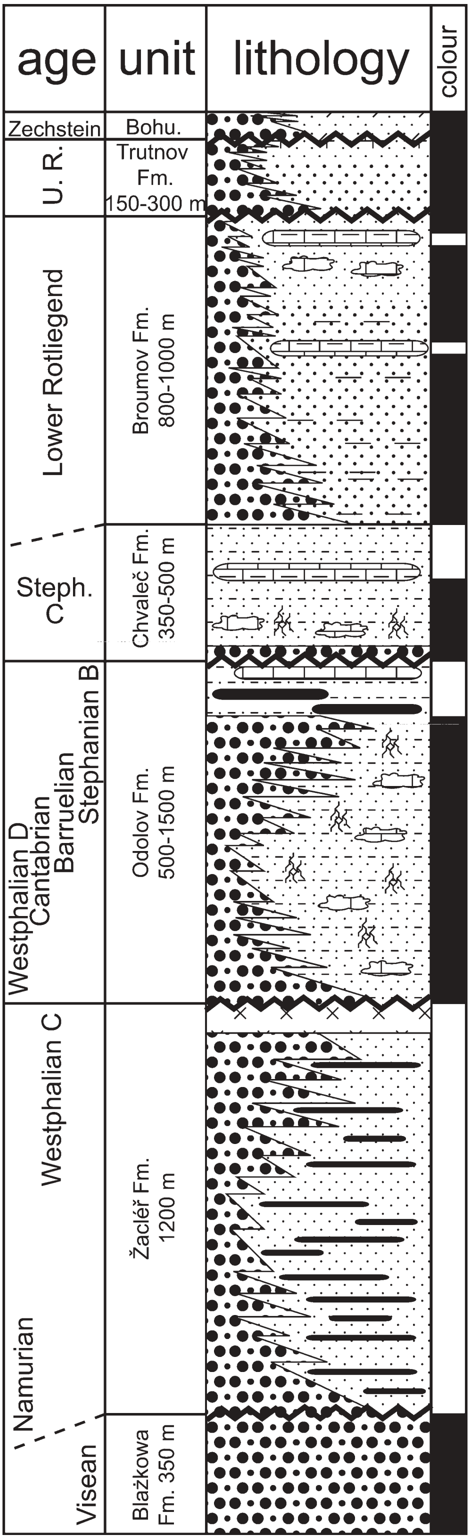


Fig. 11. General succession of the Intra-Sudetic Basin. U. R., Upper Rotliegend; Bohu., Bohuslavice Fm. 30–120 m. For legend see Figure 3.

it is in distinguishable from its precursor. The Týnec Formation is dated by macroplants to a Barruelian age (Pešek 2004), and the climate conditions were drier and possibly warmer. After a climate change to more humid conditions, the Slaný Formation was deposited during the early Stephanian B *Sysciophlebia grata* zone (Schneider 1982). During this time, finely laminated (varved) claystones formed within a freshwater lake of about 55 000 km² (Pešek 2004). The Líně Formation was dated to Stephanian C age by amphibians: *Apateon intermedius–Branchierpeton saalensis* to *Apateon dracyiensis–Melanerpeton sembachense* zone (Werneburg & Schneider 2006) and macroplants (Pešek 2004). The climate was drier than before, but cannot be regarded as pronouncedly arid (Pešek 2004) because of the interconnected lakes.

The climate development of all these basins shows an overall aridization starting from the warm-humid, coal-forming environments of the Kladno Formation (Westphalian D–?Cantabrian) to the alluvial plain red beds of the Líně Formation (Stephanian C). This trend was interrupted by the more humid climate, starting in the uppermost part of the Týnec Formation and culminating in the large lake development of the lower Slaný Formation (earliest Stephanian B). Within the upper Slaný Formation the aridization strengthened again.

Krkonoše Piedmont and Intra-Sudetic basins

The volcano-sedimentary filling of these basins covers the crystalline basement of the Krkonoše–Jizera complex and partially the Late Proterozoic and Ordovician/Silurian low-grade metamorphic rocks. The basal purple to brown-red conglomerates of the Intra-Sudetic Basin (ISB, 1800 km²; Fig. 11) fills up the palaeo-relief. The following sequence is characterized by cyclic fluvial and alluvial plain sediments, dominated by brownish to variegated conglomerates and sandstones with rare intercalations of pelites. Within these upward-fining cycles more than 60 coal seams were deposited. The final phase of the Žacléř Formation (ISB) deposition was accompanied by strong volcanic activity that produced the up to 100 m thick and laterally widespread Křenov rhyolithic tuff (309.0 ± 3.7 Ma: Lippolt *et al.* 1986). After a short break in sedimentation, the Odolov (ISB) and the Kumburk Formation of the Krkonoše Piedmont Basin (KPB, 1100 km²,

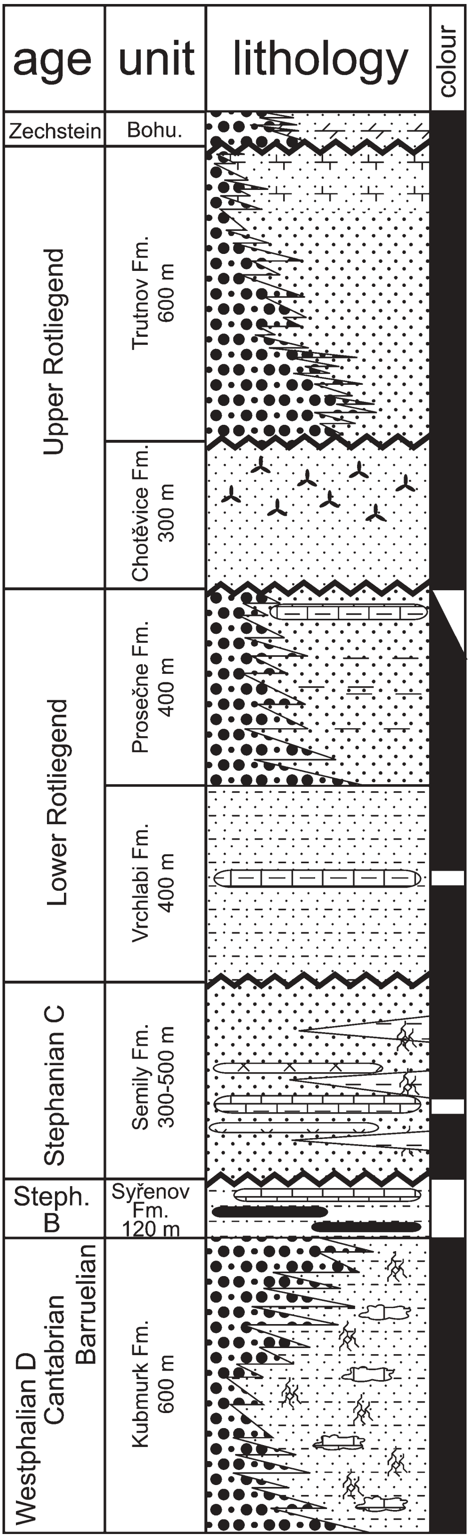


Fig. 12. General succession of the Krkonoše Piedmont Basin. Bohu., Bohuslavice Fm. 30–120 m. For legend see Figure 3.

Fig. 12) began with mostly reddish pelites arkoses and conglomerates of an alluvial plain (Opluštil & Pešek 1998). Next to the top of the Odolov Formation (ISB) (= Syřenov Formation, KPB) the facies changes to grey alluvial plain sediments containing several coal seams and some lacustrine deposits, which were connected to the large lake of the western and central Bohemian basins (Slaný Formation: Pešek 2004). In the KPB, it is overlain by mostly red-coloured, coarse-grained sediments with occasional mudstone intercalations deposited on a broad braid plain with common pedogenic horizons and a widespread horizon of bituminous shales with thin lacustrine carbonate and volcaniclastic beds (Opluštil & Pešek 1998).

The Vrchlabi Formation (KPB) as well as the Chvaleč Formation (ISB) are separated from the underlying units by an unconformity. The basal conglomerate of the latter has no equivalent in the KPB. The overlying sediments are dominated by red-brown silty and clayey sediments with common palaeosols. Some layers of lacustrine-laminated (varved), bituminous, grey limestones with cherts are intercalated in this sequence of alluvial plain wet red beds. The overlying deposits are an assemblage of red-brown arkoses with intercalated conglomerates, thin pelites and limestones of ephemeral lakes. The environment is marked by the development from an alluvial plain with ephemeral ponds to lacustrine conditions and black-shale deposition. The upper part of the Broumov Formation (ISB) is characterized by brownish, playa-like alluvial plain fine clastics with common carbonate nodules. Within the erosively overlying Chotěvice Formation (KPB) the lithology changed from red-brown sandstones to brown-red claystones with abundant gypsum pseudomorphs as the environment changed from alluvial plain to a playa-like system. The trend continues with monotonous red-brown conglomerates, sand-, silt- and claystones, containing redeposited windblown grains of the overlying formations. The latest Permian Bohuslavice Formation consists of partially dolomitic conglomerates and sandstones, and was deposited by periodic streams and sheet floods.

The age of the units, shown in Figure 15b, is based of the following. The Blażkowa Formation is of late Viséan and/or early Namurian age (Pešek 2004), as determined by macroplants in this and the Late Namurian–Bolsovian Žacléř

Formation. These units suggest a semi-humid warm climate. The earliest Westphalian D part of the Odolov Formation represents a warm-humid climate, indicated by the coal seams. The Kumburk Formation was deposited during late Westphalian D to Barruelian and is comparable to the semi-humid, wet red-bed environment of the lower Odolov Formation. The more humid conditions of the early Stephanian B (*Sooblatta stephanensis* Schneider 1983) Syřenov Formation correspond to the upper Odolov Formation. The Semily Formation indicates semi-arid conditions for the Stephanian C. The base of the Chvaleč Formation is characterized by wet red beds of a semi-arid climate. The appearance of lacustrine black-shale sediments in the upper part indicates a humid to semi-humid climate, as in the Lower Rotliegend Vrchlabi Formation. At the top of this formation a transition to more arid conditions is indicated. The Prosečné Formation is lithostratigraphically correlated to the Broumov Formation (ISB), which is determined by Werneburg (1996, 2001) to be in the *Melanerpeton pusillum–Melanerpeton gracile* zone.

The aridization trend, starting in the upper Vrchlabi Formation, continues in the lower part of the Prosečné and middle part of the Broumov Formation, but is reversed in the development of the Kalna and Ruprechtice Limestone Horizon. The re-occurrence of alluvial plain deposits in the upper Broumov Formation is consistent with an ongoing aridization to the playa system of the Chotěvice Formation. Similar to the sediments of other basins, with its arid features as well as its aeolian content, the Trutnov Formation indicates an Upper Rotliegend I to II age. The Bohuslavice Formation, which is, by analogy to the North Sudetic Basin, assigned to a Zechstein age, indicates a semi-arid climate.

The sedimentary successions of these basins show a well-established aridization trend from the humid to semi-humid Blażkowa Formation and Žacléř Formation, to the wet red beds of the Kumburk/Odolov Formation, through the semi-humid to semi-arid red beds of the Prosečné/Broumov Formation to the arid evaporitic playa environment of the Chotěvice Formation. Within this long geological record, several short-lived humidization events (reversals) are observable in the coal-bearing lower Odolov Formation, the coal-bearing Syřenov/upper Odolov Formation (humid) and the prominent lacustrine horizons in the lower Vrchlabi/upper Chvaleč Formation (semi-humid) and the upper Prosečné/middle Broumov Formation (semi-humid to semi-arid). Generally, each reversal of the aridization trend is weaker than its precursor. The aridization maximum is reached within the Chotěvice and Trutnov formations. After that a climatic reversal is observable, characterized by humidization up to the semi-arid climate of the Bohuslavice Formation.

Boskovice Graben

The N–S-striking elongated Boskovice Graben is about 100 km long and 3–10 km wide. Sedimentation (Fig. 13) began with the red-brown, clast-supported Balinka Conglomerate and cyclically intercalated sand- and siltstones on a crystalline basement. It grades to fluvio-lacustrine, grey sand-, silt- and claystones that are interrupted by several coal seams and subordinate lacustrine black shales. The overlying reddish alluvial plain deposits of the *Scoyenia*-facies are capped by the decimetre-thick Oslavany Conglomerate. Sedimentation continued, with cyclic red-brown and yellow-brown silt- and sandstones with intercalated arkoses to conglomerates and some thin grey successions. The environment is characterized by an alluvial plain to lacustrine facies that is strongly influenced by alluvial fan deposits. A facies change within the following deposits is only traceable by the intercalations of some larger complexes of bituminous, clay-rich limestones to carbonaceous black shales. They developed on an alluvial plain with several ephemeral pond and lakes and are concentrated in the lower and upper parts of the Letovice Formation.

The humid Rosice–Oslavany Formation is dated by macroplants to Stephanian C. The Padochov Formation was deposited subsequently. A major hiatus is marked by the Oslavany Conglomerate, which is thought to be the lithostratigraphical Rotliegend base. This is well supported by fossil insects in the Říčany Horizon (*Spiloblattina homigtalensis*, early Lower Rotliegend: Schneider 1980, 1982; Schneider & Werneburg 1993). The environment of this unit indicates a semi-humid climate in the lower part, developing to semi-arid conditions in the upper part. The Letovice Formation spans the Lower Rotliegend to the Upper Rotliegend I. The Zboněk–Svitavka Horizon belongs to the *Sysciophlebia alligans* zone (late Lower Rotliegend: Schneider & Werneburg 1993). The Bačov Horizon with the localities of Bačov, Obora and Sudice belongs to the *Moravamylacris kukalovae* zone (Schneider 1980) and the *Discosauriscus austriacus* zone (Werneburg 1996), which is early Upper Rotliegend I. Both horizons, the Zboněk–Svitavka Horizon in the lower part of the Letovice Formation, as well as the Míchov and Bačov Horizons in the uppermost part, are represented by a large sequence of grey clastics with metre-thick lacustrine, bituminous, calcareous and extensively laminated black shales. Therefore, they indicate a semi-humid climate, whereas the red clastics in between show semi-arid conditions.

The palaeoclimate reconstruction of this basin shows a well-established aridization trend through the whole profile, from the humid Rosice–Oslavany Formation, through the semi-arid red beds of the Veverská Bítýška Formation, to the semi-arid alluvial plain sediments of the Letovice Formation. This trend is interrupted by two humid to semi-humid phases that produced large lake horizons in the lower and upper Letovice Formation.

Moroccan basins

In the Moroccan Meseta and High Atlas Mountains, late Palaeozoic sediments and volcanics crop out, which document the development of this area during the formation of the Mauretanide part of the Hercynian orogeny. In the Western Meseta, late Visean to Early Westphalian marine turbidite sequences mark the early stages of foreland basin development, which is interpreted as the southern extension of the European Variscan belt. During Stephanian and Permian times, pure continental intramontane basins developed (Saber *et al.* 1995; for details, see Hmich *et al.* 2006). Here only a short synthesis is presented.

Biostratigraphically well-dated grey sediments of earliest Stephanian B age in the Souss Basin (Fig. 14) are transitional between the Early Stephanian wet phase and the Middle Stephanian dry phase. In a first preliminary approximation, the Moroccan Souss Basin could be interpreted as being situated during this time in the southern subtropical summer-wet belt (biome 2 of Ziegler 1990*a*). The wet red beds (*Scoyenia* facies) of the Khenifra Basin (Fig. 14), based on macrofloras, are regarded as transitional Autunian to Saxonian (Broutin *et al.* 1998). This fits well into the Artinskian wet phase and is perhaps transitional into the Kungurian. The wet red beds (*Scoyenia* facies) of the Tourbihine Member, Ikakern Formation, in the Argana Basin (Fig. 14) are dated by pareiasaur remains, which are closely related to those from the Late

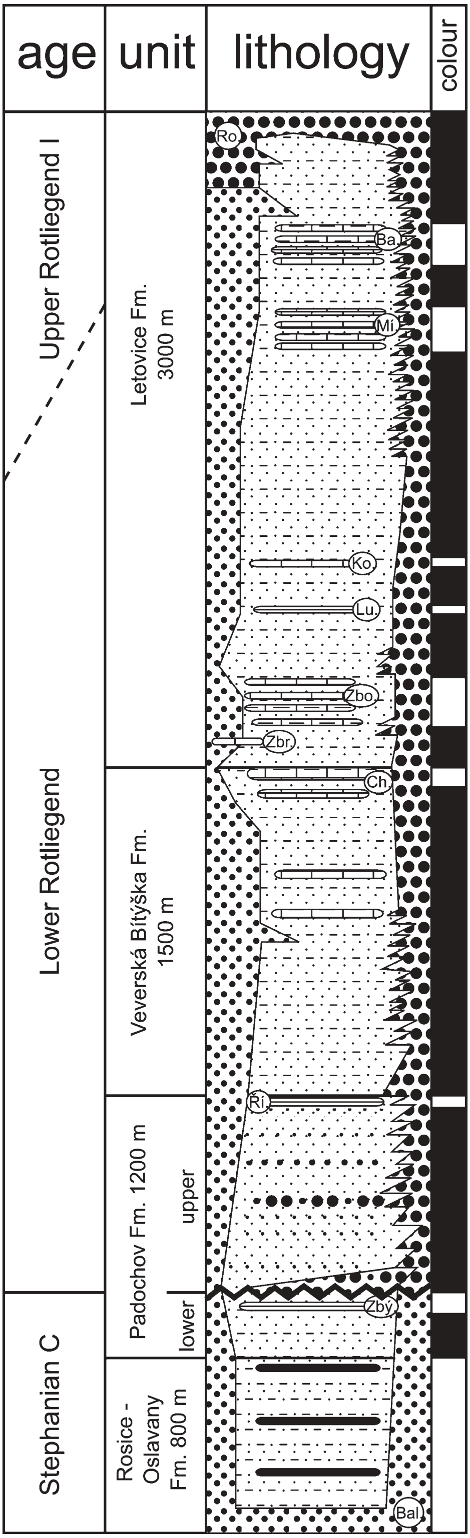


Fig. 13. General succession of the Boskovice Graben. Bal., Balinka Conglomerate; Zbý., Zbýšov Horizon; Ří., Říčany Horizon; Ch., Chudčice Horizon; Zbr., Zbraslavec Horizon; Zbo., Zboněk-Zvitávka Horizon; Lu., Lubě Horizon; Ko., Kochov Horizon; Mí., Míchov Horizon; Ba., Bačov Horizon; Ro., Rokytná Conglomerate. For legend see Figure 3.

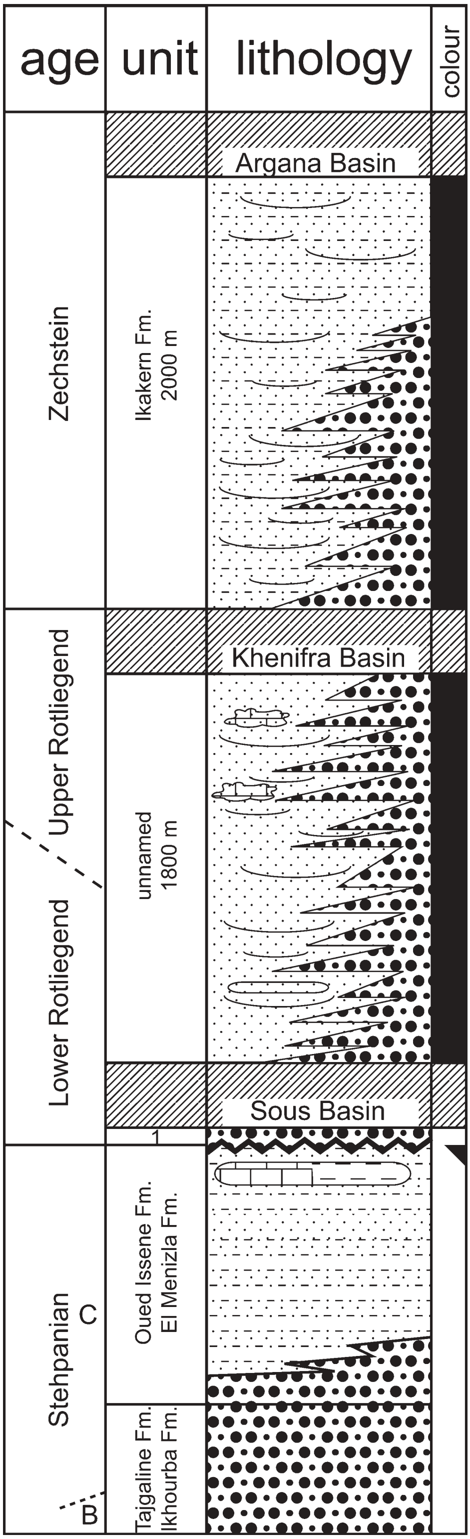


Fig. 14. General succession of the Moroccan basins. For legend see Figure 3.

Permian of Elgin, Scotland, and the Zechstein of Richelsdorf, Germany, as Middle Wuchiapingian (Jalil per. comm. 2002). Wet red beds similar to those in the Argana Basin are known in Europe, such as those from the La Lieude Formation, Late Permian of the Lodève Basin. They originated during the Wuchiapingian wet phase.

Climatic development of northern Pangaea

The most significant climate indicators mentioned above are shown in Figure 15a, b. Based on a well-justified stratigraphy it is obvious that the climate development of each basin shows more than local climate variations. Nevertheless, the climate of every basin was influenced by local features, such as subsidence, orographic barriers, slope gradient and exposition and vegetative cover. All these meso- to microclimates were local modifications of the inter-regional Euramerican climate. In addition to the undisputed aridization trend in northern Pangaea, several wet phases are correlatable between the basins. These phases are characterized by more humid conditions during a short period (2–3 Ma) in contrast to the over- and underlying units. Because of the general aridization trend, every single wet phase is weaker than its precursor. The wet phases are situated in the Barruelian, Stephanian C to early Lower Rotliegend, late Lower Rotliegend and Early Upper Rotliegend I. A further one phase the Westphalian C/D cannot be demonstrated by the dataset used here, but its occurrence is suggested because of the mechanism discussed below.

The major problem for the reconstruction of continental climate is the lack of complete successions. Therefore, a well-justified stratigraphy for many basins with incomplete successions is indispensable. All boundaries of the wet phases are fixed within continuous successions, but within different basins. For example, the lower boundary of the Stephanian C to early Lower Rotliegend wet phase can be defined only in the Saar–Nahe Basin, the Thuringian Forest Basin, the Saale Basin and the Intra-Sudetic Basin, whereas in the other sections (Fig. 15a, b) this boundary cannot be documented because the wet phase starts just above a discontinuity. These wet phases can be seen also in many basins with a smaller extent, for example the Ilfeld Basin and the Döhlen Basin (Fig. 15a).

The separation of climatic versus tectonic imprints in the sediment facies patterns is difficult, but within complete successions a drastic influence of tectonics can be ruled out. Within the late Lower Rotliegend and early Upper Rotliegend major lake horizons of wet phases are summarized. The combination of tectonic artifacts (closed system lakes) is ruled out by the distribution and migration of aquatic faunas (Schneider & Zajic 1994; Boy & Schindler 2000). The maximum of aridity was reached in the late Upper Rotliegend I to early Upper Rotliegend II. This is supported by geochemical and sedimentological investigations by Schneider *et al.* (2006). They place the maximum of aridity within the Octon Member of the Salagou Formation (Lodève Basin). This is also based on the sedimentary features of the Eisenach Formation of the Thuringian Forest Basin and the Chotěvice and Trutnov formations of the Krkonoše Piedmont Basin and Intra-Sudetic Basin. The last but most dramatic climate change in the central European area was caused by the Zechstein and *Bellerophon* transgression. The filling of the Northern and Southern Permian Basin, as well as the weaker pre-Zechstein transgressions, represent the source area of the moisture that influenced the arid climate. The effect of this large epi-continental sea is well recognized in the rapid change of facies patterns within the Late Permian of the Lodève Basin (La Lieude Formation).

The correlation of the wet phases to the global standard is difficult, because at the moment no profile with interfingering marine/continental sediments has been sufficiently studied. Therefore, no biostratigraphical link between the Permian continental sediments and the marine standard is possible, so all age determinations of the wet phases are given in 'Central European continental terms'. The correlation to the international scale presented here is based on various isotopic ages of different basins (for further information, see Roscher & Schneider 2005). The Barruelian wet phase is correlated to the middle to late Kasimovian, and the late Stephanian C to early Lower Rotliegend wet phase to the early Asselian. The late Lower Rotliegend wet phase is in the early Sakmarian and the early Upper Rotliegend I wet phase is in the early Artinskian. For the latter, the correlation is the weakest because its absolute age is not clearly defined by isotopic ages. To understand the mechanism responsible for the climatic variations we must look at the southern hemisphere.

Climate development of southern Pangaea (Karoo Basin)

When these more humid phases (wet phases) are plotted on a stratigraphical correlation chart, it is striking that they correlate with the deglaciation cycles of the Dwyka Group of the Karoo Basin (Fig. 15b & 16). The stratigraphy of the Dwyka Group is defined by isotopic ages published by Bangert *et al.* (1994, 1999), Stollhofen (1999), Stollhofen *et al.* (1999, 2000), and Rohn & Stollhofen (2000). Visser (1995, 1996, 1997) and Scheffler *et al.* (2003) dealt with the relative sea level during these times. The positioning of the transgression and regression cycles in relation to the lithological profile by both authors is the same. Differences occur with reference to absolute ages, potentially an artefact of different dating techniques and time scales.

The humidity increases towards the top of every deglaciation cycle. It is widely accepted that these cycles are of a deepening-upward (transgressive) nature. The sealevel fluctuations have a frequency of 5–7 Ma (Scheffler *et al.* 2003). This duration is not known from any orbital frequency. Therefore, the cause of these fluctuations cannot be explained at present. Nonetheless, the impact of the waxing and waning of the Gondwanan glaciation is traceable globally. The higher frequency glacio-eustatic fluctuations are represented in the marine-paralic Kansas cyclothems (Heckel 1999, 2002*a*, *b*). These high-frequency cycles are not traceable within continental sections because the age resolution for this kind of sediments is not high enough, but the lower frequency of 5–7 Ma is reflected on the continent by humid-arid alternations.

The Gondwana glaciation

The Gondwana Glaciation began around the Mid-Carboniferous boundary at 320 Ma (Bruckschen *et al.* 1999; Saltzman 2003) or just before it at 330–335 Ma (Holkerian–Asbian boundary) (Dickins 1996; Wright & Vanstone 2001; Wright 2003). Saltzman (2003) ascribed the onset of the large-scale Gondwana Glaciation to the closure of an equatorial seaway. The simultaneous reorganization of the global ocean currents supported the transportation of moisture to the southern continent. This equatorial seaway can be seen, for example, in the Western Meseta (Morocco), where the late Viséan to Early Westphalian marine turbidite sequences mark the early stages of foreland basin development (Abbou *et al.* 2001). Rapid isostatic uplift, after subduction under Laurussia (maximum of high pressure metamorphism at 340 Ma), of the continental Gondwanan crust caused the closure of the strait between the Palaeo-Tethys and the Rheic Ocean (Fig. 17). This supports the older Viséan Age for the beginning of the glaciation.

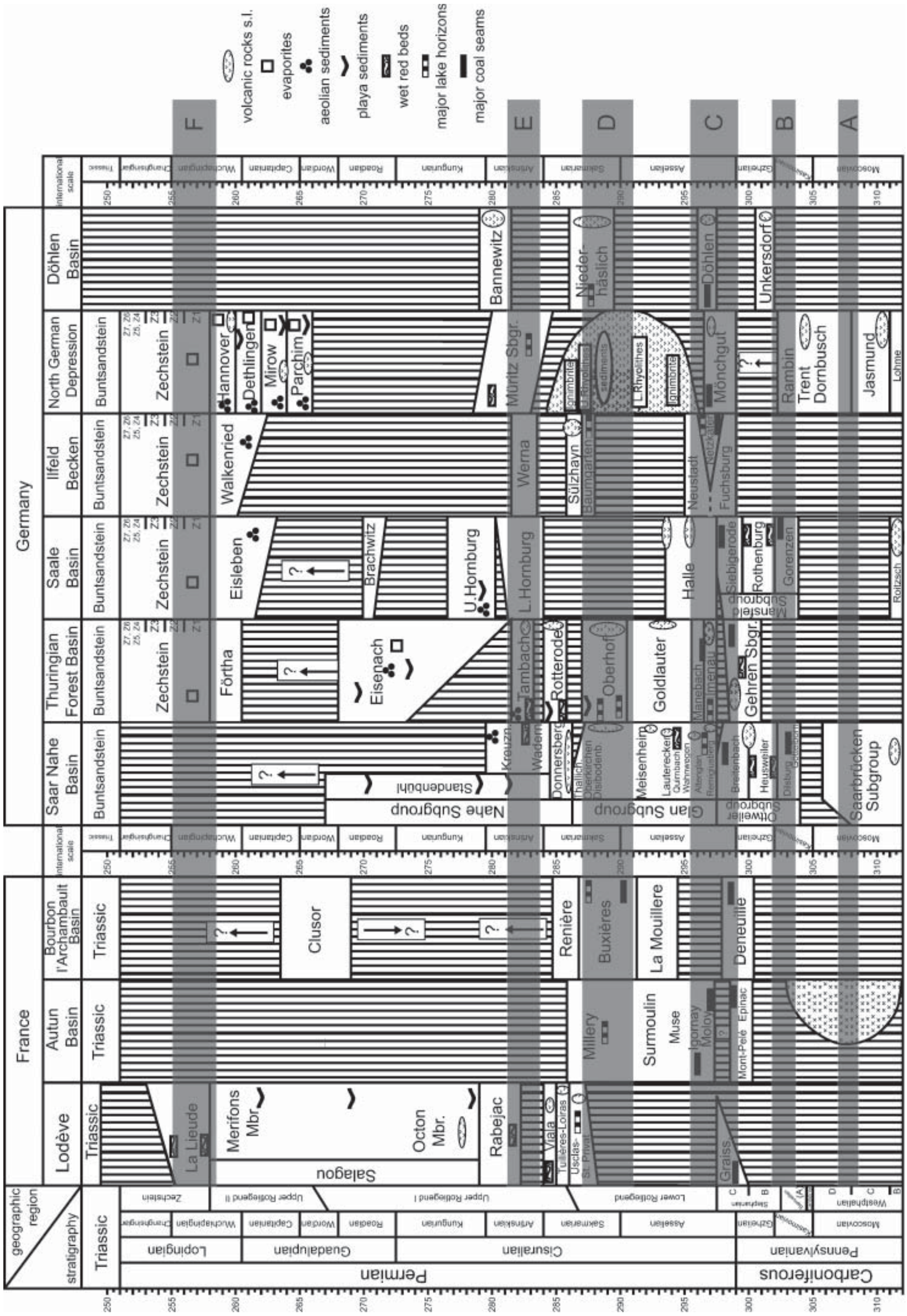
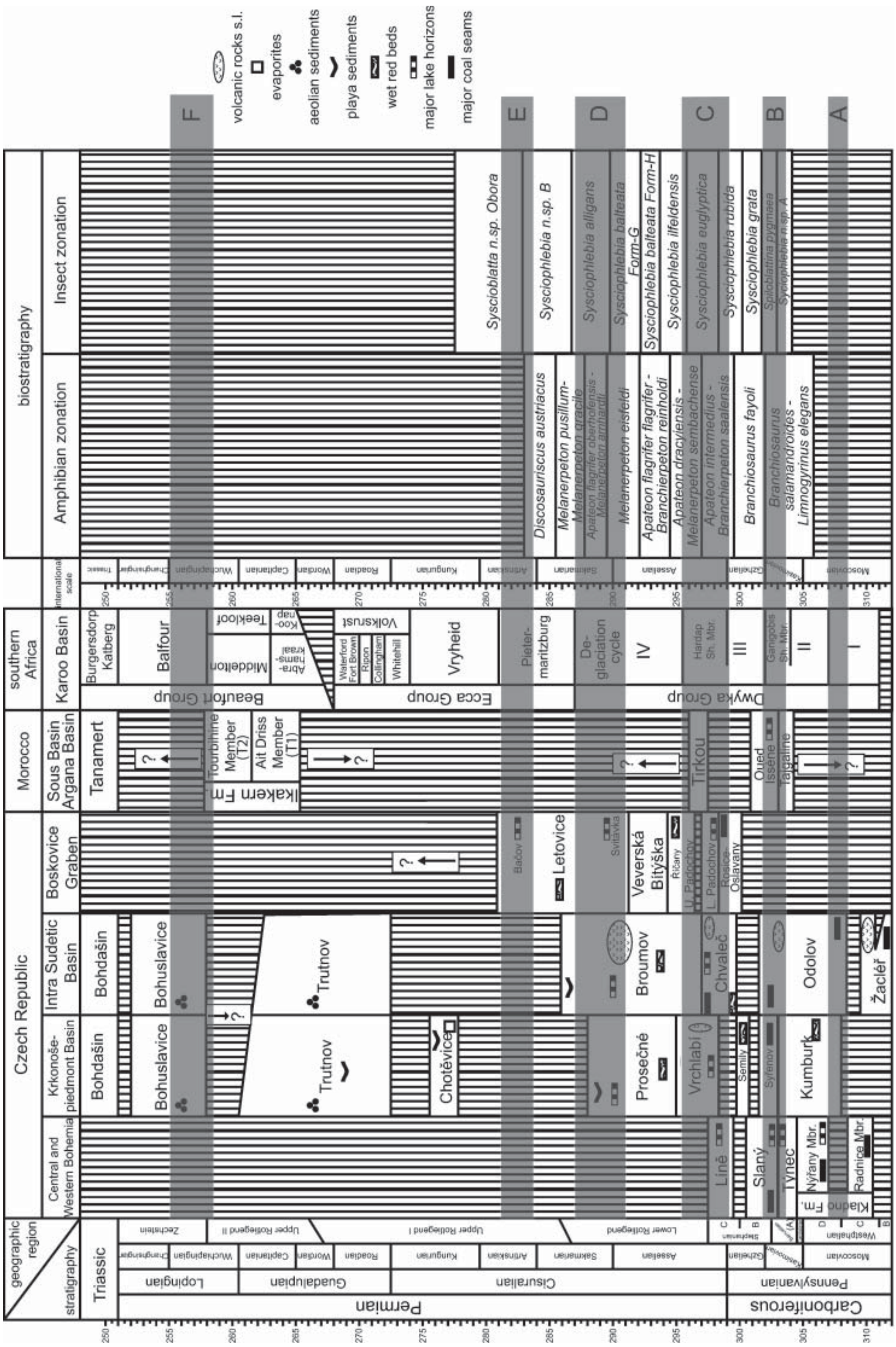


Fig. 15(a). Correlation chart of European Basins. A, assumed Westphalian C/D wet phase; B, Barruelian wet phase; C, Stephanian C to early Lower Rotliegend wet phase; D, late Lower Rotliegend wet phase; E, Early Upper Rotliegend I wet phase; F, Zechstein/Bellerophon Transgression.



(**b**) Correlation chart of European Basins, the Karoo Basin and biostratigraphy. A, assumed Westphalian C/D wet phase; B, Barruelian wet phase; C, Stephanian C to early Lower Rotliegend wet phase; D, late Lower Rotliegend wet phase; E, Early Upper Rotliegend I wet phase; F, Zechstein/Bellerophon Transgression.

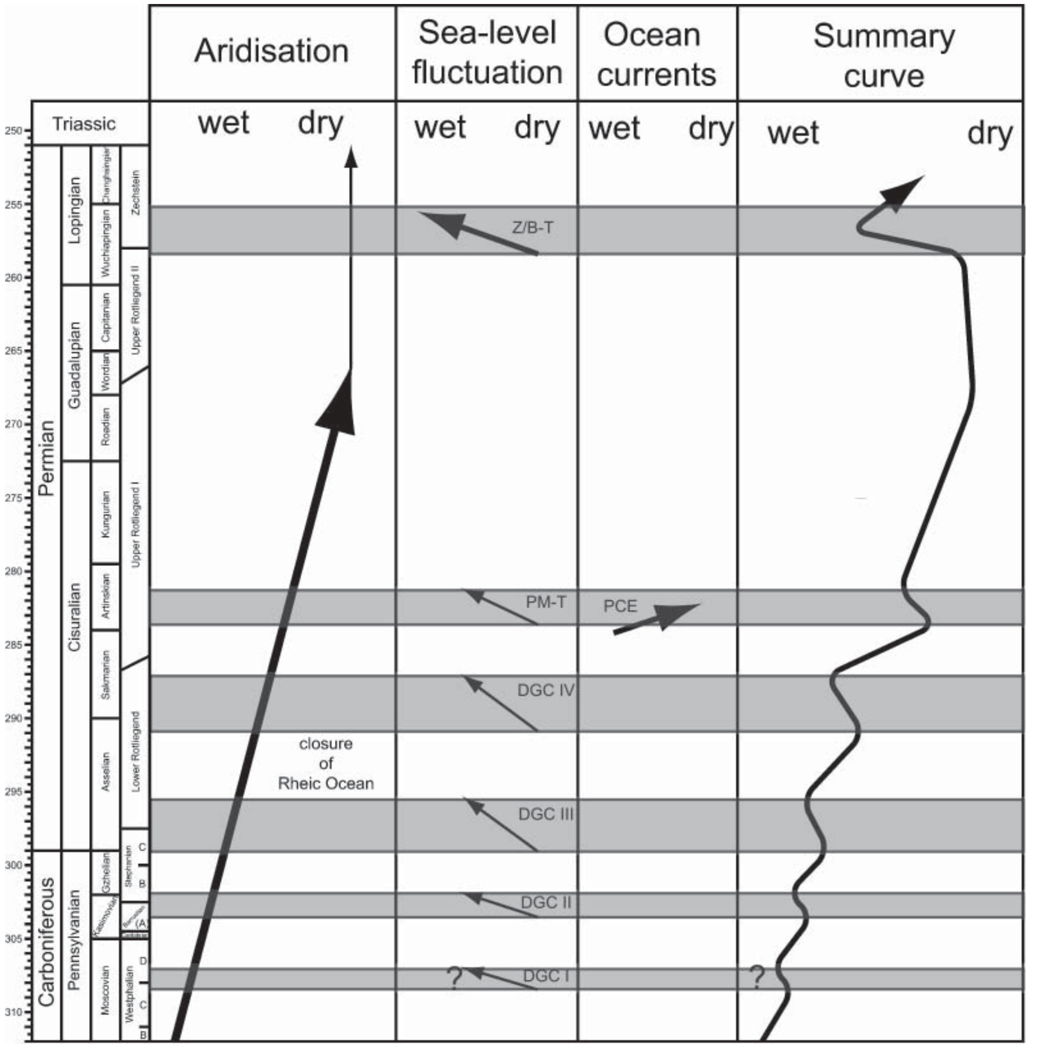


Fig. 16. Climate curve. Thickness of lines indicates the strength of the single process: DGC I–IV, Deglaciation Cycle 1–4 of the Gondwana Icecap; PM-T, Pietermaritzburg Transgression; Z/B-T, Zechstein/Bellerophon Transgression; PCE, Permian Chert Event.

The Gondwana Glaciation is recorded by glacial sediments in South Africa, South America, Australia, India and Antarctica. As in the Pleistocene, this icecap supported the expression of Milankovich cycles. This high cyclic sedimentation is described in Canada (Falcon-Lang 2003, 2004), as well as in the Mid-Continent Basin (e.g. Heckel 1999, 2002*a*, *b*). Within these fourth to fifth order cyclothems, glacio-eustatic sea-level changes of 60 to 200 m are documented (e.g. Crowell 1978, 1999; Veevers & Powell 1987; Heckel 1994; Isbell *et al.* 2003). An ice-volume difference of 35–115 million km^3 was calculated by Crowell & Baum (1991) between glacial and interglacial periods to produce eustatic sealevel fluctuations of this scale. This implies a huge icecap located over Antarctica, Australia, South America, India and South Africa, as proposed by Ziegler *et al.* (1997), Golonka (2000) and Scotese (2001).

In contradiction, Isbell *et al.* (2003) reported ice-free Trans-Antarctic Mountains during Carboniferous times and an ice marginal position for it in the Early Permian. Broutin *et al.* (1995) pointed out that the early Westphalian Al Khlata Formation of Oman was influenced only by alpine glaciers. Similar deposits of the same age are described from Queensland, Australia (Dickins 1996). The global maximum of glaciation is described by DiMichele *et al.* (1996) and Sano *et al.* (2003) for the late Westphalian. For the Paraná Basin in South America, dos Santos *et al.* (1996) published the ice maximum adjacent to the Permian–Carboniferous boundary, as Broutin *et al.* (1995) did for Oman. For the Early Permian, the glaciation is described as alpine-type for the Oman region (Angiolini *et al.* 2003*a*, *b*); in Australia glaciation occurred in five phases (Dickins 1996) and sediments of alpine glaciers in Queensland were deposited between non-glacial fluvial, lacustrine and shallow marine strata (Jones & Fielding 2004). For South Africa, Catuneanu (2004) published a general direction of ice movement from north to south.

To summarize, it remains unclear where and when this huge, continent-wide polar cap was established, which produced the estimated 60–200 m eustatic sealevel changes. A lower eustatic amplitude seems more realistic, so the ice-covered area was probably not as large as originally thought. Sedimentological research has shown that the amplitude of the glacio-eustatic fluctuations are on a scale of a maximum of 100 m for the major cyclothems, with incised valleys up to 30 m deep (Soreghan & Giles 1999; Feldman *et al.* 2005).

One of the most familiar and best-investigated profiles within the outcrop area of the glacial sediments is the Dwyka Group in the Karoo Basin, South Africa. The sedimentology and palaeontology of this unit have been published by Visser (1995, 1996, 1997). Isotopic ages were measured by Turner (1999) and Stollhofen *et al.* (2000), and geochemical investigations were carried out by Scheffler *et al.* (2003). The main features are the four deglaciation cycles with fining-upward sequences (Figs 15b, 16). Correlating this profile by isotopic ages (Fig. 15b) to the international time scale, the four eustatic maxima are in late Moscovian (Westphalian C/D, ~308 Ma), late Kasimovian (Barruelian, ~304 Ma), early Asselian (Stephanian C, ~298 Ma) and Asselian–Sakmarian transition (late Lower Rotliegend, ~289 Ma). The Gondwana glaciation ends at the beginning of the Sakmarian. This is concurrent with a worldwide classical *Eurydesma* transgression (Visser 1997). Sakmarian transgressive systems are also described in Australia (Dickins 1996) and Oman (Angiolini *et al.* 2003*a*, *b*). With the disappearance of the polar cap, the highly cyclical transgression-regression sequences within coastal and shallow marine sedimentary basins also stop (Beauchamp & Baud 2002).

The next eustatic transgression in southern Gondwana is in the late Artinskian (early Upper Rotliegend I, ~282 Ma) Pietermaritzburg transgression (Visser 1997). The cause for this flooding is not found in the icecap. After the Asselian, only seasonal and/or local glaciations are reported (e.g. Dickins 1996; Henderson 2003). All these Late Carboniferous and Early Permian transgressions occurred with a constant rhythm of about 5–7 Ma, ascribed to the Gondwana Glaciation. The Pietermaritzburg transgression, however, does not fit this pattern. Perhaps eustatic sealevel fluctuations with a low frequency of 5–7 Ma drove the glaciation, and only the higher order cycles are glacio-eustatic. The cause for the 5–7 Ma cyclicity is not known, and no processes with a comparable frequency have been described.

Palaeo-ocean currents and salinity

Oceanic and atmospheric circulations control the distribution of climatic belts. They are responsible for the transport of heat from the equator to the poles and vice versa. Near the equator (~30°S to ~30°N) the oceans carry more than 50% of the heat, and in higher latitudes the atmosphere plays the more important role (Von der Haar & Oort 1973).

The position, direction and relative temperature of the ocean currents with respect to their surroundings significantly influence the modern continental environment. Two types of large currents that impact continental climate should be mentioned as recent examples. The cold and upwelling currents of the Humboldt and Benguela Stream cause zones of high bio-production in the marine environment and coastal deserts on the land. These effects occur because the moist air, coming from the oceans with westerly winds, deposits rain over the cold ocean and does not reach the continent. The second example is the Gulf Stream in the northern Atlantic. This brings heat and moisture from the Gulf of Mexico to the northern high latitudes of Europe and is responsible for the mild climate in the United Kingdom and Norway. All these currents are part of a conveyor-belt-like heat pump that stretches around the whole world. The mechanism that keeps these circulations in motion changed during the past. Today, thermal and thermohaline circulation patterns produce cool oxygen- and nutrient-rich bottom water that generates areas of high organic productivity when upwelling. The opposite mode, haline circulation, is driven by the sinking of warm, high salinity water due to intense evaporation. The Mediterranean Sea is a small recent analogue (Talley 1996). The high salinity trail that leaves the strait of Gibraltar can be observed across the entire Atlantic (Worthington & Wright 1970). The warm saline bottom waters are dis- to anoxic and poor in nutrients. Their upwelling generates areas of lowered organic productivity and areas that are hostile to life (Kidder & Worsley 2004).

The reconstruction of palaeo-ocean current systems is mostly made by numerical models (Winguth *et al*. 2002). These calculations for an atmosphere/ocean simulation are based on many parameters, such as palaeogeography, seafloor spreading, palaeotopography, salinity, carbon dioxide level in the atmosphere, vegetation cover and position of ice. Given these variably known parameters, the models can only be tested by looking at the effects of the circulation on shallow marine and continental systems.

The model of Winguth *et al*. (2002) for the Middle Permian (Wordian) shows a cost-parallel, cold current at the western border of Euramerica. This is supported by the investigations of Beauchamp & Baud (2002), who described a cold stream along the west coast of North America that caused the Permian Chert Event (PCE). This event is marked by shallow-water chert deposition, which implies a cooler environment on the shelf than could be assumed by latitude. This model is supported by LePage *et al*. (2003), who discovered warm-temperate macrofloras within Kungurian cold water shelf carbonates in the Canadian High Arctic, which requires warm-temperate continental climates in the neighbourhood of cold oceanic conditions. For this part of the model (Winguth *et al*. 2002), the reconstructed ocean currents match the geologic record.

On the other side of the Panthalassic Ocean, there is a discrepancy between the model and biogeography. Shi & Grunt (2000) and Weldon & Shi (2003) showed that, during Roadian and Wordian time, cool water currents from Australia to Siberia, Mongolia and the Pamir Mountains are necessary to explain the distribution of coldwater adapted brachiopods (*Terrakea* Booker 1930). This is not explained by the numerical model of Winguth *et al*. (2002), although these authors used different values for the atmospheric carbon dioxide concentration. However, the variations of this parameter do not deliver satisfying reconstructions. Perhaps changing other parameters, such as salinity, topography, palaeogeography, could provide models to explain this geological situation.

Palaeogeography

For most recent numeric palaeoclimate models, there has been the question of using a Pangaea A or B reconstruction for calculation (Klein 1994; Kutzbach & Ziegler 1994; Fluteau *et al*. 2001; Gibbs *et al*. 2002; Berthelin *et al*. 2003). Pangaea A, the Wegnerian one, is not under discussion for the Triassic (Muttoni *et al*. 2003). But, for the Carboniferous and Early Permian, the Pangaea B reconstruction introduced by Irving (1977) better fits the palaeomagnetic data as well as the numeric climate models versus geological record (Fluteau *et al*. 2001). Nevertheless, the Pangaea B model has one master problem: how was it transformed to the younger Wegnerian Pangaea A?

Major strike-slip systems are proposed by many authors (e.g. Doblas *et al*. 1998; Muttoni *et al*. 2003; Youbi *et al*. 2003). But, these continent-wide mega-shear systems have not been demonstrated in the field. Also, the geometry of a strike-slip fault zone through the whole of Pangaea is difficult to imagine, because a stress field that produces only one major fault along the compressive Gondwana–Eurasia suture is not plausible. In the older literature (e.g. Arthaud & Matte 1977), there are descriptions of some large dextral transverse fault systems within Europe. The Tornquist–Teysseyre, Bay of Biscay, Gibraltar and South Atlas fracture zones are mentioned in many publications, such as Ziegler (1990*b*), Doblas *et al*. (1998) and Golonka (2000).

Kroner (in Schneider *et al*. 2006) plotted a Pangaea A (Wegnerian) reconstruction (255 Ma: map after Scotese 2001) on a true sphere with a recent Earth diameter (Fig. 2). The grid and Laurussia were fixed in position, and everything else except Gondwana and Laurussia was omitted. Now, the Ouachita, Alleghenian, Appalachian, Mauretanid and Variscan Orogen were stripped back along their major fault systems. During this, the four European fracture zones mentioned above became most important. The relative motion of Gondwana with respect to Laurussia was reconstructed by the combination of structural elements as fault zones, fold and thrust belts and magmatic activity. Because of the geometric constraints on the sphere, the complete closure of the Rheic Ocean requires two consecutive plate rotation processes, as already demonstrated by Wise (2004) for the deformation style of the Pennsylvania salient of the North American Appalachians. Therefore, the reconstruction of Pangaea 300 Ma ago (Fig. 2) looks somewhat different from that in use by most authors (Ziegler 1988, 1990*b*; Ziegler *et al.* 1997; Golonka 2000; Scotese 2001).

The post-Caledonian orogens along the Gondwana–Laurussia suture testify to collision processes taking place from the Early Devonian to the Late Permian. The existence of a continuous Trans-Pangaean Mountain Belt, with high altitudes reaching from the Variscides to the Ouachita Mountains (Keller & Hatcher 1999), as used for climate modelling by Kutzbach & Ziegler (1994), Fluteau *et al.* (2001) and Gibbs *et al.* (2002), requires contemporaneous orogenic processes along the whole plate boundary. Actually, the individual orogens give evidence of diachronous tectono-thermal events. http://dict.tu-chemnitz.de/dings.cgi?o=3001;count=50;service=de-en;query=auf. Particularly with regard to uplift and exhumation processes taking place in the Early to Late Carboniferous and Permian, large differences between the European Variscides, the North American Appalachian Orogen and the Ouachita Mountains exist.

The climax of high-pressure metamorphism during the central European Variscan Orogen occurred in the Early Carboniferous. Strong erosion and differential uplift accompanied the subsequent rapid exhumation and cooling of the high-grade metamorphic complexes. At the end of the Carboniferous, the orogenic crust of Europe equilibrated (Kroner *et al.* 2004). Finally, dextral transpression during the Permo-Carboniferous took place in local, small-scale tectono-thermal events (Gaitzsch 1998; Kroner & Hahn 2003; Capuzzo & Wetzel 2004). Late Carboniferous arc-continent collision led to the very low-grade fold and thrust belt of the Ouachita Mountains at the southern edge of Laurentia (Thomas 1989). There is no hint of large-scale uplift and erosion at a subsequent stage. The final closure of the Rheic Ocean during the Permian, caused by the clockwise rotation of Gondwana and contemporaneous shearing along the South Atlas and Gibraltar Fracture Zone (Fig. 2), created the Alleghenian–Mauretanides Orogen of North America and western Africa (Arthaud & Matte 1977; Ziegler 1982). Convergent tectonics, producing large-scale fold and thrust belts, started in the Late Carboniferous and ceased in the Late Permian. Tectonothermal events, uplift and exhumation processes continued during the whole time span (Hatcher *et al.* 1989). Large-scale uplift and erosion lasted until the Triassic.

As a consequence of the processes delineated above, a Pangaea-wide mountain chain during the Middle Permian is very doubtful. High altitudes should occur in Appalachian and west African mountain belts at this time. In comparison, a strong topography is ruled out in the region of the Ouachita Mountains and the European Variscides. Furthermore, the reconstruction of the late Palaeozoic plate movement of Gondwana relative to Laurussia (Figs 2 & 17) shows that there is no need for a strong collision between South America and Laurentia. Instead of this collision and orogen as published by many authors (e.g. Kutzbach & Ziegler 1994; Keller & Hatcher 1999; Fluteau *et al.* 2001), the remnant of the Rheic Ocean formed an embayment from the Panthalassia Ocean to the mid-European areas that persisted at least from Early to Middle Permian time. This oceanic embayment was situated in the low latitudes of Pangaea and would explain the marine Permian sediments described by Benison & Goldstein (2000) in the southwest of the Mid-Continent Basin (Utah, Colorado, west Wyoming), as well as the basin facies in the southwest of the Mid-Continent Basin. This deep shelf facies is not situated next to an orogen, as shown by Heckel (1999, 2002*a*, *b*).

Palaeotopography of Pangaea in European surroundings

The elevation of the Variscides, Mauretanides and Appalachians is of major importance for European and global climate during the Permo-Carboniferous. Different estimates exist for the altitude range of the Variscan mountain chain, some in excess of 5000 m (Kutzbach & Ziegler 1994; Becq-Giraudon *et al.* 1996; Fluteau *et al.* 2001; Gibbs *et al.* 2002). All numerical climate models that include topography generalize an equatorial mountain chain, as published by, for example, Keller & Hatcher (1999), with an average elevation of about 2000–2500 m, contrary to Gibbs *et al.* (2002), who assumed a maximum height of 1800 m. Based on our knowledge of the European realm (section 6) these hypotheses do not seem plausible.

The culmination of pressure-emphasized metamorphism during the Variscan continent–continent collision was during the Early Carboniferous (Mississippian) (Gaitzsch 1998; Gaitzsch *et al.* 1999; Bosse *et al.* 2000; Matte 2001; Medaris *et al.* 2003; Rodriguez *et al.* 2003; Kroner *et al.* 2004), at about 340–350 Ma. The following main uplift/exhumation processes documented a rapid cooling of the high-pressure domains by isothermal decompression (Kröner & Willner 1998, and references therein), so the main elevation should have been reached just after that. The uplift and tectonic exhumation of the high-grade

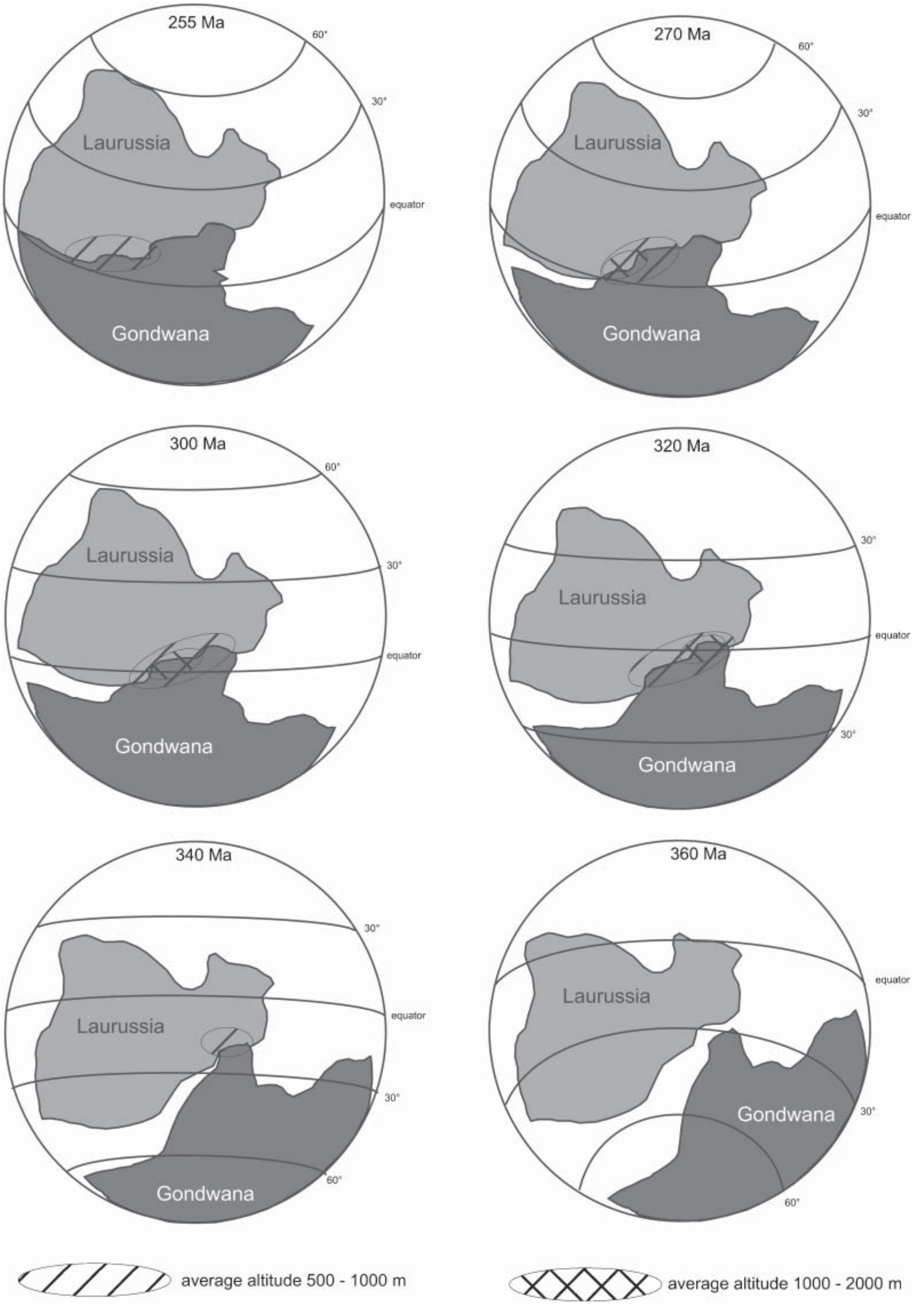


Fig. 17. Sketch map of the Pangaea formation and estimated palaeotopography.

metamorphics was simultaneous with the erosion, as is observed today in New Zealand (Simpson *et al.* 1994). When the orogen reached its maximum elevation, most of its roof was already eroded. The first undeformed sediments within the Variscan realm were described by Ahrendt *et al.* (2001) from the Erzgebirge Basin, which rests on deformed turbiditic flysch sediments. These sediments were dated by macroflora to the Early Carboniferous (e.g. Gothan 1949; Daber 1959). Palynological investigations by Bek (1997 pers. comm.) delivered a younger, Asbian age (early Late Viséan, V3b). Isotopic dating of a tuff yielded an age of 330 ± 4 Ma (Gehmlich *et al.* 2000), and Ahrendt *et al.* (2001) demonstrated the existence of Variscan detritus by dating white mica with a temperature of closure to ~330 Ma. Capuzzo & Wetzel (2004) reported a rapid removal of more than 5 km of upper Variscan crust in the intramontane Salvan-Dorénaz Basin during the Late Carboniferous. They justified their assumption by $^{39}Ar/^{40}Ar$ chronometry of white mica from subaerially cooled volcanics in contrast to ages derived from the sedimentary detritus above the volcanics. Intramontane basins, which may have existed, would have been cannibalized during the Late Carboniferous. Also, the Bohemian basins and the French basins of the Massif Central (cf. pages 98–102), which rest on Variscan roots, granites and high metamorphic complexes, started their evolution by at least Stephanian time. For example, the basement of the Blanice Graben consists of metamorphic rocks derived from a depth of 30–35 km (Opluštil 2004). Therefore, an attempt to explain high altitudes versus sedimentary basins is presented here.

Recent denudation rates of 3.0 ± 1.3 to 4.5 ± 1.7 mm/a (3.5 mm/a in average) for the western Himalayan mountain range, which is in a totally different climatic belt than the Variscides, were published by Garzanti *et al.* (2005). This would imply an erosion of 10 km within a time span of less than 3 Ma. This does not seem to be very realistic. To bring it down to earth, using an erosion rate of 0.5 mm/a, an orogen summit of 10 km would be eroded during 20 Ma. The only factor to account for the lowering of the erosion rate is that the linear calculation does not represent nature.

Thus, during most of the Carboniferous, the Variscan mountain range was situated next to the equator in a warm-humid tropical climate. This implies a high precipitation rate and consequently a high weathering and erosion rate. This allows a comparison to recent New Zealand, where extremely high exhumation rates are observed. The uplift rate for this region over the last 16 ka has been calculated at 13.7 mm/a (Simpson *et al.* 1994). However, the topography does not reflect this uplift because only some individual mountains exceed 3000 m. Comparing palaeo-erosion rates with recent ones is difficult because of the differences in the evolutionary level of the flora, but it gives a hint of truth. Also, few Palaeozoic plants had root systems comparable to modern plants.

The Thuringian Main Granite, which forms the basement of the Thuringian Forest Basin, underwent, by dextral transpression, an exhumation from 310–300 Ma by a denudation of 8–10 km of overlying rocks (Zeh *et al.* 1998). This underlines the hypothesis of fast erosion and levelling of the Variscan mountain chain. For the Pyrenees, Maurel *et al.* (2004) published ages of granite intrusions of 305 Ma. Using various isotopic systems, these authors estimated cooling rates of 30 °C/Ma. These rates decrease at 290 Ma to 1 °C/Ma. In comparison to Central Europe, it is the same duration of the orogen but the phase is displaced by about 15–20 Ma. This displacement can also be seen by the later maximum of anatexis in Central Iberia between 335–305 Ma (Montero *et al.* 2004).

However, a palaeotopographical reconstruction of the whole Hercynian orogen has not been established, and the juxtaposition of contemporaneous sedimentary basins and deeper crustal units excludes a Hercynian-wide, Himalayan-type mountain belt at any time of the orogeny. At least by the beginning of the Stephanian (~305 Ma), the Central European Variscan orogen was levelled to a hilly landscape. However, it cannot be ruled out that local transpressional horst systems produced alpine-type mountains of limited extent. Recent analogues are the Tatra Mountains in the Czech Republic (Sperner *et al.* 2002).

Generally, the clockwise rotation of Gondwana (Fig. 17) caused the migration of the maximum elevation through the continent. The first uplift occurred in the central east of Pangaea. This was continued in the Early to Middle Permian by the Appalachians and Mauretanides.

Seasonality

As shown above, laminated (varved) sediments of perennial lakes are known since at least the late Stephanian (Stephanian B of Bohemian basins: Pešek 2004) and can be traced up to the youngest perennial lakes (Oberhof Formation,

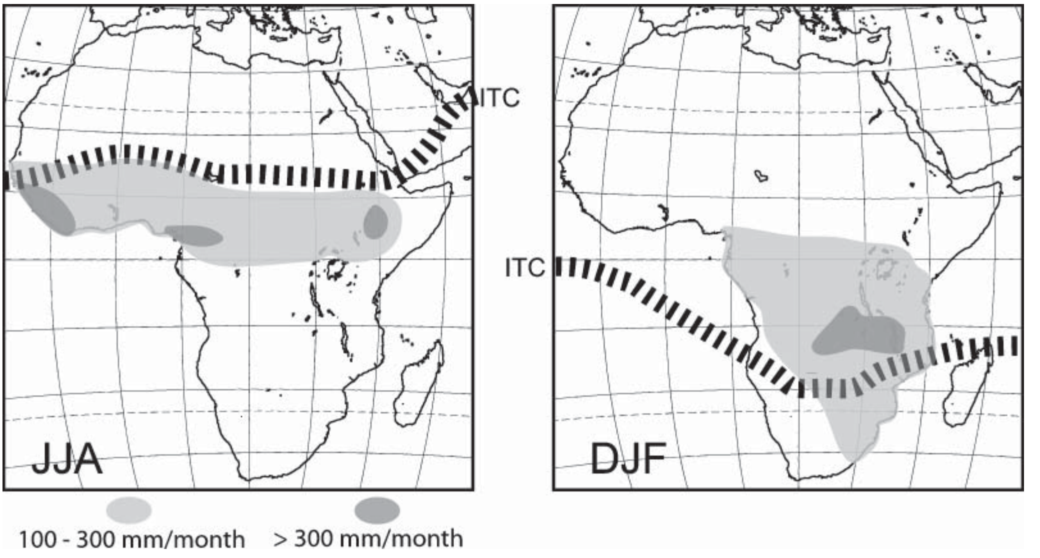


Fig. 18. Recent precipitation over Africa. JJA, northern summer; DJF, northern winter.

Buxières Formation, etc., see Fig. 15a, b) in the topmost Lower Rotliegend (Schneider 1996). They are interpreted as the result of annual changes of wet and dry seasons of strong monsoonal climate (Schäfer & Stamm 1989; Stapf 1989; Schneider & Gebhardt 1993; Clausing & Boy 2000). Further, Ziegler (1990*a*) showed by investigations on macroplants that, during the Permian, the biome of the ever-wet tropics disappeared in central and eastern Pangaea next to the equator, but not on the Cimmerian continents. Exceptions are reported only from Morocco, Spain and Texas, but, without references. The model proposed here is based on an increasing seasonality starting in the Stephanian and gaining strength in the Permian, which inhibits ever-wet biomes. The huge landmasses in the northern and southern mid-latitudes generated a strong monsoonal climate, as has already been modelled, for example by Parrish (1993) and Kutzbach & Ziegler (1994). During Permian time the zone of Inter-Tropical Convergence (ITC) was much more displaced over the continents, like today (recent values for Africa ~20°N/S: Zahn 1991; see Figure 18). In the Permian it reached latitudes of about 30°N and 30°S (Kutzbach & Ziegler 1994). The associated precipitation belt was not displaced as much as the ITC (Fig. 19).

The broad shift of the rainfall area is comparable to that of eastern Africa, but on a larger scale. Seleshi & Zanke (2004), for example, published data for Ethiopia; in the south at 5°N, the wet seasons are in the spring and autumn. This seasonal distribution is comparable to the equatorial climate in central Pangaea. Additionally, the biomes compare well. The recent arid to semi-arid savannahs of Ethiopia (Zahn 1991) may be a modern analogue for the Permian summer-wet tropical deciduous forests or savannahs (biome 2 *sensu* Ziegler 1990*a*). Northern summer (JJA) and winter (DJF) dry seasons are expected for the whole of central Pangaea (Fig. 19). This is caused by the relatively large distance to the ITC and its precipitation belt and the influence of trade winds. Spring and autumn wet seasons occur while the ITC is placed directly above the basins, which are discussed herein. All Permian climate models assume moisture for the precipitation over the CEBs associated with the trade winds from the Tethys. These passat winds rained out in front of the Trans-Pangaean Mountain Chain (Keller & Hatcher 1999). But, as shown above, these mountains were not as high as would be necessary to induce heavy rainfall accompanied with a complete moisture loss at the southern flank of this supposed mountain chain.

In the modern world, only a small part of the central African rain is achieved from the east. For ancient times, by analogy, the moisture of the ITC must have been obtained from the west, as it is today over Africa. The large distance to the Panthalassic Ocean is in contradiction to this

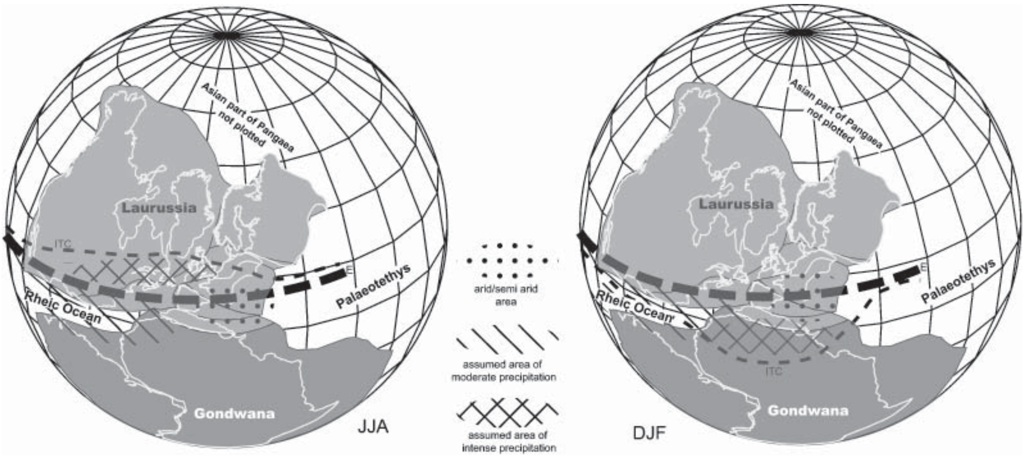


Fig. 19. Precipitation over Gondwana 300 Ma ago, grid not arranged to the palaeo-equator. E, equator; ITC, Inter-Tropical Convergence Zone; JJA, northern summer; DJF, northern winter.

hypothesis, if existing palaeogeographic reconstructions are utilized (e.g. Ziegler 1988, 1990*b*; Ziegler *et al.* 1997; Golonka 2000; Scotese 2001). The palaeogeographic reconstruction of Kroner (in Schneider *et al.* 2006), used here, does not create this problem. The supercontinent Pangaea receives its precipitation from the mega-ocean Panthalassia, with its embayment of the remnant Rheic Ocean. Compared to most previous climate models, the topography is assumed not to have been high enough to produce a continent-wide rain shadow (Fig. 17). Why did the first seasonal sediments occur during Stephanian B? The Variscan Orogen started to rise as an elevation near the mid-Carboniferous boundary. The emerging mountain chain was high enough to induce a significant effect on the distribution of rainfall. By the beginning of the Stephanian B, the morphology of the Variscan Mountains was no longer important for the climate evolution of central Pangaea. This coincides with the estimates made earlier for erosion versus exhumation rates.

Aridization

The pronounced seasonality was strengthened by development of more arid conditions in the continental interior. This aridization began in the Westphalian (Chevalier *et al.* 2003) and is interrupted by multiple wet phases. Gibbs *et al.* (2002) attributed the aridization of the northern part of Pangaea to the northward movement of the whole continent over about 10°–15°N away from the equator. But, this explains only the younger half (Middle to Late Permian) of a trend that began in the Carboniferous. If this continental movement was the only cause of the northern Gondwanan aridization, there should have been a humidization trend in the Carboniferous while moving from the south towards the equator. But, there is no indication of this.

Also, this movement should but did not leave a traceable record of a southward shifting of the ever-wet tropical belt. The appearance of more arid/continental conditions can also be interpreted as being due to a barrier to maritime influences of the Rheic Ocean remnant. The slow closure of the Rheic Ocean between North and South America during the Pennsylvanian and Permian had strong effects on the climate. It changed step by step from maritime humid/semi-humid climates to more continental arid/semi-arid conditions. Gradually, the large Carboniferous coal swamps disappeared towards the Permian. The aridization trend in the European basins is marked by the spreading of red-bed facies. Near the Sakmarian–Artinskian boundary, this effect becomes stronger (Fig. 16). For example, Hembree *et al.* (2004) described episodic, perhaps seasonal droughts in the late Wolfcampian (late Sakmarian–early Asselian) coastal plain of Kansas. They based their hypothesis on amphibian burrows and ephemeral ponds (Speiser Shale, Kansas). Similar phenomena are observed in the Lodève Basin (Körner *et al.* 2003; Schneider *et al.* 2006). The increasing pace of the aridization cannot be explained by a faster closure of the Rheic Ocean.

The breakdown of the Carboniferous coldhouse and the associated Gondwana glaciation by the Sakmarian caused a reorganization of the major ocean currents. This is expressed by the onset of the Permian Chert Event (Beauchamp & Baud 2002). The sedimentation of cherty sequences in a shallow marine environment is interpreted as a cold-water environment. Because this depositional style can be found along the northwestern Pangaea shelf, they attributed this to the presence of a cold longshore current. The model of Winguth *et al.* (2002) showed that the cold conditions could be explained simply by upwelling. This could have also been induced by the offshore transport of surface water due to the wind-driven Ekman transport (perpendicular to the right of the wind direction in the northern hemisphere) (Gibbs *et al.* 2002).

This current had strong effects on continental climate. This cold barrier to the west of the north-drifting continent inhibited the moisture, which came with the westerly winds to rain over the continent. Recent similarities can be found in SW Africa (Namib Desert) and the SW of South America (Atacama Desert), due to the Benguela and Humboldt Currents. These two phenomena, the slow closure of the Rheic Ocean since the Viséan, and the rapid change of the oceanic circulation patterns, are assumed to be the cause of the central Pangaea aridization. Studies of very young sediments in the marine successions next to the Congo River mouth (Schefuss *et al.* 2005) revealed correlations between monsoon strength and equatorial precipitation. The stronger the monsoon, the less it rains. They attributed this relationship to the effect of the monsoon winds on cloud distribution. A stronger wind with an east–west component transports more clouds away to the ocean. This causes less precipitation on the continent.

The gradual closure of the Rheic Ocean had two effects on the continental climate in Central Europe. First, the distance to the ocean in the west increased and, second, the monsoon strengthened. This effect of accelerated aridization can be observed in many basins in Central Europe (Figs 15a, b & 16). The post-Artinskian climate development in the Euramerican continental environment is not reconstructable as accurately as in the Late Carboniferous and Early Permian.

In most of the European basins, with the exception of the Saar–Nahe Basin and the Lodève Basin, sedimentation from the middle Cisuralian up to the late Guadalupian is interrupted by several hiatuses (Fig 15a, b). Only scattered deposits without sustainable biostratigraphy occur in most basins. The Standenbühl Formation of the Saar–Nahe Basin (Fig. 15a) is not well enough investigated. The only section with essentially continuous sedimentation is the Salagou Formation of the Lodève Basin. These red beds were well investigated by Körner *et al.* (2003) and Schneider *et al.* (2006) in a sedimentological and geochemical sense and for its climatic implications. Within this succession, the ongoing aridization is obvious, and the maximum is reached in the upper Octon Member (Fig. 15a). This unit is estimated to be of Roadian age. The following Mérifons Member seems to have been deposited under more humid conditions.

A process that could be responsible for this maximum of aridization is published by Shi & Grunt (2000) and Weldon & Shi (2003). They showed the necessity for a cool current along the Cimmerian continents to explain the distribution of cold-water-adapted brachiopods. The brachiopod *Terrakea* Booker 1930 from Australia is distributed in Siberia, Mongolia and the Pamirs during Roadian and Wordian time. This cold current could prevent moisture from the east reaching the continent. But, this current was next to the Cimmerian continents, and the Tethys was located to the west of it. So, this current is a hint of a global reorganization or modification of the oceanic currents. Until more sections of continental Middle Permian sediments are thoroughly investigated, the climatic development of the Central European region for this time span is based on estimates only. Probably, the climax of the Permian aridization produced conditions that did not allow normal sedimentation, due to the lack of the transport medium. Alternatively, nearly the entire European region was elevated above base level by doming (due to a superplume: Doblas *et al.* 1998; or crustal underplating: Bachmann & Hoffmann 1995). Most of the scattered Upper Rotliegend I deposits of Central Europe are characterized by angular to subangular components and a high content of primary aeolian sediments. This is interpreted to indicate a more or less exclusive physical weathering with a deficit of debris removal over large areas of Permian Europe.

Aeolian sediments

The appearance of aeolian sediments is connected to the previously mentioned aridization trend. The first occurrence of important amounts of wind-transported grains is observable in the early Upper Rotliegend I (Fig. 15a, b). The first dunes appear in the Upper Rotliegend I of the Hornburg and Kreuznach formations, but are more prominent in the Upper Rotliegend II

(Walkenried Formation, Ilfeld Basin; Northern and Southern Permian Basin: Glennie 1982).

Two features controlled the occurrence of aeolian sediments. The first was the aridity within northern Pangaea. As was shown earlier, the process of ongoing aridization produced large areas of arid to semi-arid environments. Therefore, a large amount of sediment was available. The second controlling fact was the velocity of the wind. In order to produce large dune systems, a strong and directionally constant wind system is necessary. Glennie (1982) showed that, in the Upper Rotliegend II of the Northern Permian Basin, mega-dunes occur. For the formation of such giant structures, comparable to the Pleistocene ones, high wind velocities are necessary. The high atmospheric velocities in the Pleistocene were caused by the compression of the climatic belts by the spreading out of the icecaps. In the Late Permian (late Guadalupian and Lopingian) no large glacial deposits are known. The climatic belts in the Late Permian were not compressed but spread out. This is supported by the investigations of Cúneo (1996), who showed that the Antarctic Continent was forested with cold temperate and humid *Glossopteris* vegetation. Ziegler (1990*a*) ascribed it to biome 6 (nemoral broadleaf–deciduous forests), which, in recent times, are typical in eastern and central Europe as well as in the eastern USA.

The discrepancy of the high velocity of the winds over northern Pangaea and the spreading out of the climatic belts is difficult to explain. Two facts should be considered. On the one hand, the Pangaean continent was a much larger connected land area than today. On the other hand, the ever-wet tropical biome disappeared and the equator was surrounded by desert to semi-desert regions. The development of this huge wasteland was caused by closure of the Rheic Ocean. The marine influence on the continental climate was retreated step by step until the final oceanic embayment, the Rheic Ocean remnant, was closed in the Late Permian (Fig. 17). This retreat of the ocean due to plate tectonics caused, in the Permian, a slow shift from marine to continental climate in northern Pangaea. Starting from the Lower Permian, the monsoon system strengthened so that larger dune systems could be established. With the further development of the exceptionally large and dry central Pangaean desert, the wind velocities were raised to values comparable to those of the Pleistocene

Conclusions

The Permo-Carboniferous climate was marked by an aridization trend (Chumakov & Zharkov 2002). This trend was interrupted by several wet phases in the Barruelian, Stephanian C to early Lower Rotliegend, late Lower Rotliegend and early Upper Rotliegend I. These more humid phases can be seen in all European and North African sedimentary basins (compare Figs 1 & 15a, b). All these wet phases are linked to the waxing and waning of the Gondwana icecap. This is supported by correlations with the South African Karoo Basin. The melting of the icecap caused a eustatic sealevel rise, which led to a more humid climate in North Pangaea. A further wet phase in the Westphalian C/D cannot be proven by the dataset used here, but its existence is presumed because of its origin due the waxing and waning of the Gondwanan icecap.

The wet phases themselves, as a part of the aridization trend, weakened over time. The aridization and the weakening of the wet phases were caused not only by the northward drift of the supercontinent and the shifting of North Pangaea to the arid climatic belt (which was modelled by Gibbs *et al.* 2002). A spreading-out of the arid/semi-arid belt in the Lower to Middle Permian is traceable by the disappearance of ever-wet tropical associations (Ziegler 1990*a*). This spreading-out can only be explained by using a new reconstruction of the configuration of Pangaea (cf. Figs 2 & 17). During the Late Carboniferous and the Early Permian, an oceanic embayment of the Panthalassic Ocean, a large remnant of the Rheic Ocean, existed between North and South America. This ocean was successively closed during the Permian. The retreat of this water body displaced the source area of moisture for central North Pangaea stepwise to the west. With increasing distance from the ocean, the aridization and the monsoon system strengthened. The stronger monsoon system transported more clouds to the offshore area (Schefuss *et al.* 2005). The result was less precipitation on the continent. Later, at the end of the Gondwana Glaciation in the latest Sakmarian, oceanic circulation was rearranged. This led to a cold coast parallel ocean current that induced chert sedimentation on the shallow shelf along the west coast of North Pangaea. The cold-water surface temperatures blocked moisture coming with the westerly winds.

The maximum of aridity was reached during Roadian and Wordian time. A chief premise of all these hypotheses is that the major part of the precipitation was sourced to the west. The Trans-Pangaean Mountain Belt was non-existent in the new palaeogeographic reconstruction (Figs 2 & 17). The altitude of the Hercynian system never exceeded an average of 2000 m, and the maximum elevation of the Trans-Pangaean orogeny shifted during time from the east to the west with the ongoing closure of the Rheic Ocean. The Hercynian Orogen never acted as a large orographic barrier blocking moisture and precipitation from the west. The Variscian Orogen was levelled down at least by the middle Stephanian (Stephanian B). The lack of an orographic east–west barrier led to a strong seasonality. The Inter-Tropical Zone of Convergence (ITC) was displaced over larger distances than today. This huge displacement of the associated precipitation area caused four seasons at the equator. The summer and winter were dry because the ITC was further north or south. Only the spring and autumn were marked by heavy precipitation. This environment was not suitable for ever-wet tropical ecosystems. These can only persist in areas where onshore winds bring moisture from the oceans to the continents. This was the case for Cathaysia and the regions around the Rheic Ocean (Morocco, Spain and Texas: Ziegler 1990*a*). The Permo-Carboniferous climate of North Pangaea is a result of many processes that originated in palaeogeography, palaeotopography, glacio-eustatic sealevel changes and ocean current organization (Fig. 16), which can only be separated by investigations based on well-justified stratigraphic correlation of the sedimentary basins (Fig. 15a, b).

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