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ORIGINAL PAPER

Brunovistulian terrane (Bohemian Massif, Central Europe) from late Proterozoic to late Paleozoic: a review

J. Kalvoda · O. Babek · O. Fatka · J. Leichmann · R. Melichar · S. Nehyba · P. Spacek

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Abstract The Brunovistulian terrane represents a microcontinent of enigmatic Proterozoic provenance that was located at the southern margin of Baltica in the early Paleozoic. During the Variscan orogeny, it represented the lower plate at the southern margin of Laurussia, involved in the collision with the Armorican terrane assemblage. In this respect, it resembles the Avalonian terrane in the west and the Istanbul Zone in the east. There is a growing evidence about the presence of a Devonian back-arc at the margin of the Brunovistulian terrane. The early Variscan phase was characterized by the formation of Devonian extensional basins with the within-plate volcanic activity and formation of narrow segments of oceanic crust. The oldest Viséan flysch of the Rheic/Rhenohercynian remnant basin (Protivanov, Andelska Hora and Horní Benesov formations) forms the highest allochthonous units and contains, together with slices of Silurian Bohemian facies, clastic micas from early Paleozoic crystalline rocks that are presumably derived from terranes of Armorican affinity although provenance from an active Brunovistulian margin cannot be fully excluded either. The development of the Moravo–Silesian late Paleozoic basin was terminated by coal-bearing paralic and limnic sediments. The progressive Carboniferous stacking of nappes and their impingement on the Laurussian foreland led to crustal thickening and shortening and a number of distinct deformational and folding events. The postorogenic extension led to the formation of the terminal Carboniferous-early Permian Boskovice Graben located in the eastern part of the Brunovistulian terrane, in front of the crystalline nappes. The highest, allochthonous westernmost flysch units, locally with the basal slices of the Devonian and Silurian rocks thrusted over the Silesicum in the NW part of the Brunovistulian terrane, may share a similar tectonic position with the Giessen–Harz nappes. The Silesicum represents the outermost margin of the Brunovistulian terrane with many features in common with the Northern Phyllite Zone at the Avalonia–Armorica interface in Germany.

Keywords Bohemian Massif · Brunovistulian terrane · Paleogeography · Tectonic structure · Variscan orogeny

Introduction

The Brunovistulian terrane represents a Cadomian unit of a problematic provenance located on the eastern flank of the Bohemian Massif (Fig. 1; Dudek 1980; Matte et al. 1990; Schulmann et al. 1991; Lobkowitz et al. 1998; Kalvoda 1995; Zelazniewicz et al. 1997, 2001; Finger et al. 1995, 2000a; Belka et al. 2000, 2002; Kalvoda et al. 2003;

J. Kalvoda (✉) · O. Babek · J. Leichmann ·
R. Melichar · S. Nehyba
Institute of Geological Sciences,
Masaryk University, Kotlarská 2,
611 37 Brno, Czech Republic
e-mail: dino@sci.muni.cz

O. Babek
Department of Geology,
Palacky University, Tr. Svobody 26,
77146 Olomouc, Czech Republic

O. Fatka
Institute of Geology and Paleontology,
Charles University, Albertov 6,
128 43 Prague 2, Czech Republic

P. Spacek
Institute of Physics of the Earth,
Masaryk University, Tvrdeho 12,
612 00 Brno, Czech Republic

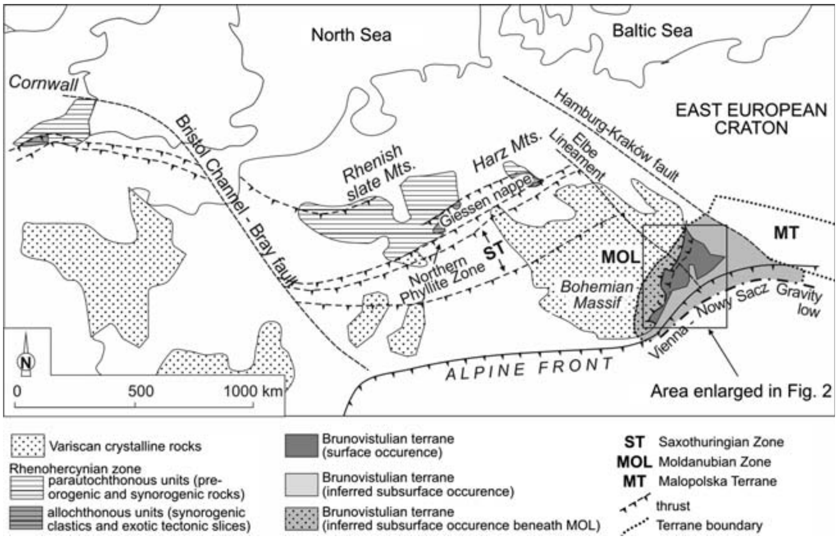


Fig. 1 Geological position of the Brunovistulian terrane in the European Variscides

Schulmann and Gayer 2000; Edel et al. 2003). This terrane was formed during the Pan-African (Cadomian) orogeny and was strongly reworked and then incorporated into the Variscan Bohemian Massif (van Breemen et al. 1982; Franke 1989; Jelínek and Dudek 1993; Finger and Steyrer 1995; Kalvoda 1995; Schulmann et al. 1994; Lobkowitz et al. 1998; Finger et al. 2000a). During the Variscan orogeny, in contrast to the remaining part of the Bohemian Massif, the Brunovistulian terrane acted as the southern margin of Laurussia (Finger and Steyrer 1995; Kalvoda 1995; Finger et al. 1998; Kalvoda 1998). In this respect, it shared a similar position with the eastern Avalonia and both terranes correspond to the Rhenohercynian Zone and, in a more distal foreland, to the Subvariscan Zone in the classical concept of the Central European Variscides (Kossmat 1927; Engel and Franke 1983; Franke 1989).

In the east, it is covered by the nappes of the Outer Western Carpathians and below these units it extends as far as the Vienna–Hodonín–Nowy Sacz axis of gravity low (Stranik et al. 1993). The Brunovistulian terrane faces the Malopolska unit along the Hamburg–Krakow Fault Zone in the north, the Alpine crystalline rocks in the south (Frisch and Neubauer 1989; Raumer and Neubauer 1993; Frisch et al. 1993; Neubauer and Frisch 1993; Finger et al. 1993; Neubauer and Handler 2000) and the Lugodanubian units (terranes of the Armorican terrane assemblage in the Bohemian Massif, i.e Moldanubian, Lugian and Central Bohemian unit—Chlupac and Vrana 1994) of the Bohemian Massif in the west (Figs. 1, 2).

The term Brunovistulicum was introduced by Havlena (1976) and later redefined by Dudek (1980) as a late Proterozoic unit comprising the Brno and the Vistulian (Visla) areas, covered by the basal Devonian siliciclastics. The term Brunovistulian unit is preferred here to other terms such as the Brno, Upper Silesian or Moravo–Silesian units because the first two refer solely to the southern and the northern parts of the Brunovistulian terrane, respectively, whereas the Moravian and Silesian units (the terms Moravicum and Silesicum of Suess 1903, 1912 will be preferred here) constitute narrow zones composed mostly of metamorphosed rocks with a Brunovistulian affinity at the contact with the Lugodanubian unit (Mísar and Dudek 1993; Lobkowitz et al. 1998; Schulmann and Gayer 2000; Kröner et al. 2000). In the regional geological classification, the Brunovistulian terrane includes the Brunovistulicum as well as the eastern part of the Silesicum (Desná dome; Dudek 1980; Mísar and Dudek 1993) and most probably the Moravicum (Lobkowitz et al. 1998; Friedl et al. 2004) and the western part of the Silesicum unit (Keprník dome; Schulmann and Gayer 2000) and at least a part of Velke Vrbno high-grade rocks forming a narrow belt along the contact of the Silesicum and Lugicum (Fig. 2; Kröner et al. 2000). For the most part, both the Moravicum and the Silesicum thus represent metamorphosed equivalents of the Brunovistulicum, which were incorporated in the Moravo–Silesian shear zone (Fig. 2), most probably including in the Silesicum also slices of Lugodanubian units (Hladil et al. 1999; Kröner et al. 2000).

In the present paper, we will examine the geologic evolution (Pan-African and Variscan) and paleogeographic position of the Brunovistulian terrane from its late Proterozoic birth to the late Paleozoic amalgamation with the remaining parts of the Bohemian Massif.

Pan-African (Cadomian) tectonosedimentary record

The Pan-African (Cadomian) Brunovistulicum is exposed in the large Thaya (Finger et al. 1995) and Brno batholiths

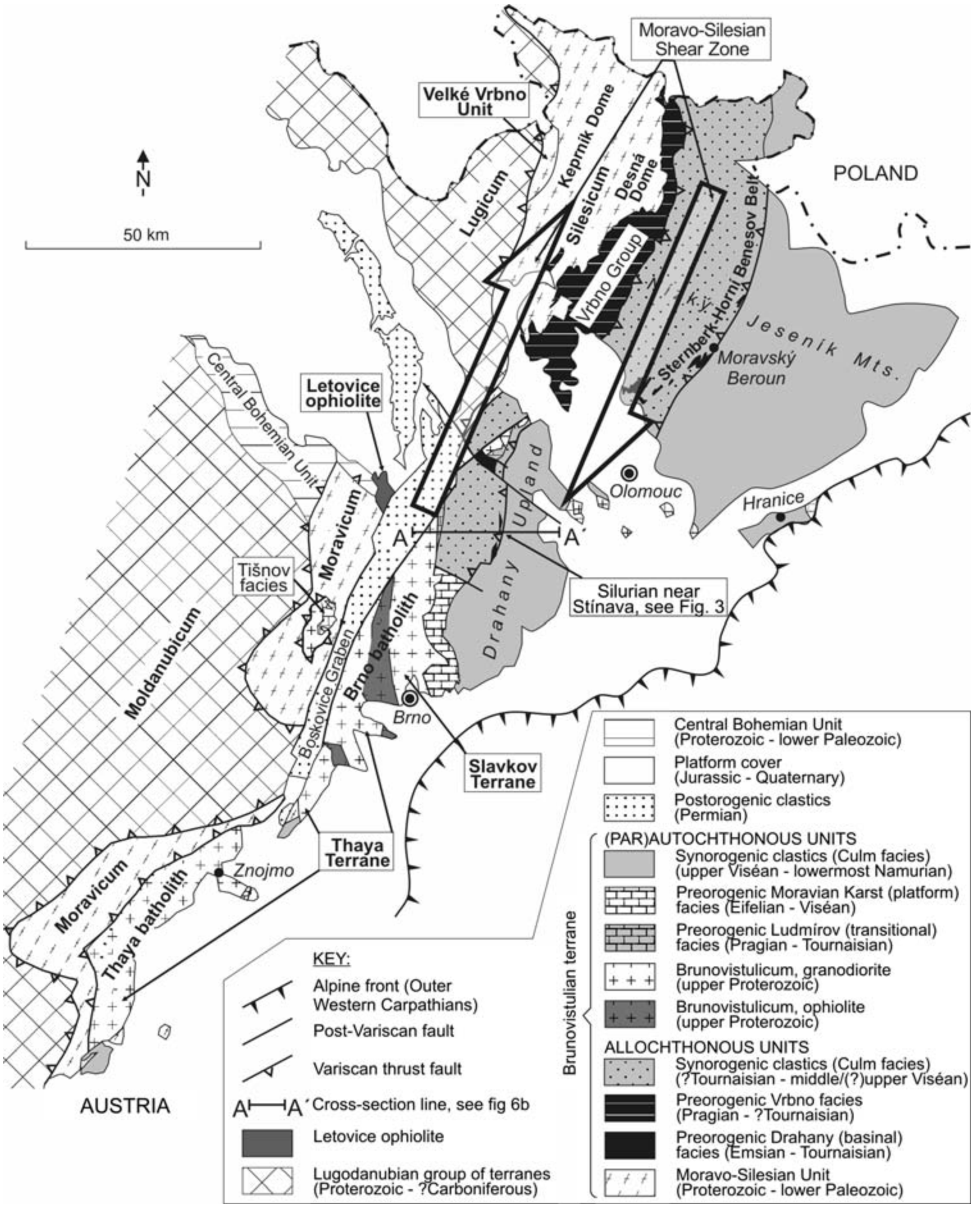


Fig. 2 Simplified geological sketch and subdivision of the Brunovistulian terrane and its position at the easten margin of the Bohemian Massif

(Leichmann 1996), each of which of about 600 km^2 in size (Fig. 2). On the surface, the Brunovistulicum and its equivalents are also known from the Jeseníky Mts. (the Desna and Keprnik domes; Fisera and Patocka 1989; Schulmann and Gayer 2000; Kröner et al. 2000), a small tectonic window beneath the Variscan Nappes near Tisnov and several isolated outcrops in the Carpathian Foredeep near Olomouc. The Brunovistulicum is known from deep boreholes penetrating the Alpine Carpathian foredeep in the Czech Republic (Jelínek and Dudek 1993), Austria (Dirnhofer 1996; Finger and Riegler 1999; Riegler 2000) and Poland (Moczydlowska 1995a, 1997). The reach of the Brunovistulian positive magnetic anomaly to the S beneath the Alps and to the NE beneath the Outer Western Carpathians suggests that the total area of the Brunovistulicum is an order of magnitude higher than that indicated by its surface exposure.

A structure of crucial importance for the whole Brunovistulicum is the Brno batholith. It consists of three major units, each of distinct evolution, which were amalgamated at the end of the Pan-African (Cadomian) orogeny (Finger et al. 2000a). A narrow, N–S trending almost complete ophiolite belt exposed in the central part of the Brno batholith divides the whole Brunovistulicum into two different units, the Thaya terrane in the SE and the Slavkov terrane in the NE (Fig. 2; Finger et al. 2000a). The available petrological data (Leichmann 1996; Hanzl and Melichar 1997; Finger et al. 2000b) indicate that the ophiolite complex was derived most probably from a supra-subduction environment. 580–600 Ma granitoids (van Breemen et al. 1982, Dallmeyer et al. 1994) intrude into the ophiolite belt on both sides.

The granitoids of the Thaya terrane are not uniform; both S-types as well as I-types predominate over small granitic plutons with an A-type affinity. The areal extent of biotite–amphibole bearing diorites and tonalites is limited, they are spatially associated with S-type granites. However, the gravity data (Skacelová and Weiss 1978) indicate that the S-type granites are underlain by the diorites. The granitoids and the diorites intruded into high-grade metapelites. A high $^{87}Sr/^{86}Sr$ initial ratio (0.708–0.710) and low εNd from –4 to –7 are typical for the granitoids (Finger et al. 2000a). Therefore, the S-type granitoids of the Thaya terrane originated as a consequence of basalt underplating. Such geological structure indicates the presence of older crust in the Thaya terrane.

The granitoids of the Slavkov terrane are petrographically more homogenous, consisting of I-type, amphibole- to amphibole–biotite bearing quartz diorites to granodiorites. Typically, they have low (0.704 to 0.705) Sr initial ratios and high (0 to +3) εNd (Finger et al. 2000a), high Sr/Rb ratio and low concentration of HFS elements. All these data suggest that the rocks can be interpreted as primitive volcanic-arc granitoids (Chappel and White 1992; Pearce et al. 1984). Metamorphic rocks of the Slavkov terrane are known mainly from boreholes in its NE part. They include flyschoid greywackes, silstones and arenites with intercalations of metabasalts and metaandesites (Dudek 1980). The initial $^{87}Sr/^{86}Sr$ ratios (0.704–0.706) as well as the εNd values (–1 to +2) are similar to those of the granitoids. Consequently, Finger et al. (2000a) interpreted them as arc-derived metamorphosed volcanoclastic sediments.

For the most part the Brunovistulian crystalline rocks are directly overlain by Paleozoic deposits but in the N and NE parts of the Brunovistulicum in southern Poland, a Vendian flysch sequence of phyllites, metapelites, metarenites and metaconglomerates is present. These anchimetamorphic rocks are regarded as deposits of a Cadomian foreland basin (Bula and Jachowicz 1996). The Cambrian Pan-African (Cadomian) molasse occurs in two areas, one, which can be correlated with the Slavkov terrane, is located in the N and NE part of the Brunovistulicum in Upper Silesia and the other corresponding to the Thaya terrane, is located in its SE part. The two occurrences were not necessarily deposited in the same geotectonic position, reflecting the as yet unclear course of the Cadomian orogeny (Finger et al. 2000a). The lithology of these rocks ranges from conglomerates to shales, with a predominance of sandstones (quartzose sandstones to subarcoses). Sedimentological and ichnological data (Mikulas and Nehyba 2001; Vavrdova et al. 2003) reflect terrestrial (fluvial) and marine (shoreface, shallow marine) depositional environments. The source area of the clastics was an active continental margin with predominance of intermediate and acid igneous rocks (Gilikova et al. 2003). Micropaleontological and palynological data (acritarchs, vendotaenids, prasinophytes) suggest an Early Cambrian age for a majority of these deposits (Jachowicz and Prichystal 1997; Fatka and Vavrdova 1998; Vavrdova et al. 2003) and reveal strong paleobiogeographic similarity with the East European Platform.

Variscan tectonosedimentary record

Ordovician and Silurian rocks

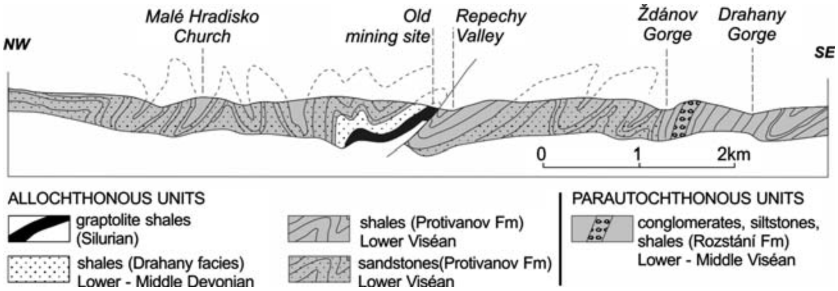
The early Paleozoic sedimentary record represented by Ordovician and Silurian rocks is rather scanty, restricted only to very small areas. The only occurrence of the Ordovician sediments includes light green clay-rich siliceous rocks, interbedded with fine-grained quartz sandstones with a variable degree of silicification. It has only been encountered in a restricted area in the northern part of Upper Silesia (Gladysz et al. 1990; Bula and Jachowicz 1996).

The only record of Silurian sedimentation is preserved in a tectonic slice at the base of a Carboniferous flysch overthrust near Stínava in the W allochthonous part of the Brunovistulian terrane (Figs. 2, 3; Kettner and Remes 1936; Chadima and Melichar 1998; Babek et al. 2006). The provenance of the tectonic slice is difficult to identify more precisely. The Silurian succession comprises graptolite shales and calcareous shales with subordinate lenses of impure limestone and contains graptolites and nautiloid cephalopods, which indicate Telychian, Sheinwoodian, Gorstian and Ludfordian ages (Kettner and Remes 1936; Kraft and Marek 1999). Silurian (438 ± 16 Ma) magmatic activity has been documented through K–Ar dating of a basalt vein intruding into the Brunovistulian basement (Prichystal 1999). The cooling ages of detrital white micas and monazites present in the Devonian and Carboniferous sediments ranging from 487 to 420 Ma (Schneider 2002) suggest an important magmatic or thermal activity in their source area during the Ordovician–Silurian period.

Devonian and Carboniferous rocks

We can subdivide the Devonian to Carboniferous interval into two evolutionary phases, an extensional phase and a compressional to transpressional phase. The extensional phase is documented by volcano-sedimentary facies deposited in a passive continental margin setting. During the compressional phase, deep-marine siliciclastics (Variscan synorogenic flysch) were deposited in a trench to a deep-marine foreland basin setting (Kumpera and Martinec

Fig. 3 Geological profile in the central part of the Drahany Upland showing the tectonic position of the Silurian sediments. Modified according to Kettner (1966)



1995). The compression gradually turned to a dextral transpressive tectonic regime, during which marine to terrestrial sediments accumulated in a foreland basin setting.

Extensional phase followed by thermal subsidence

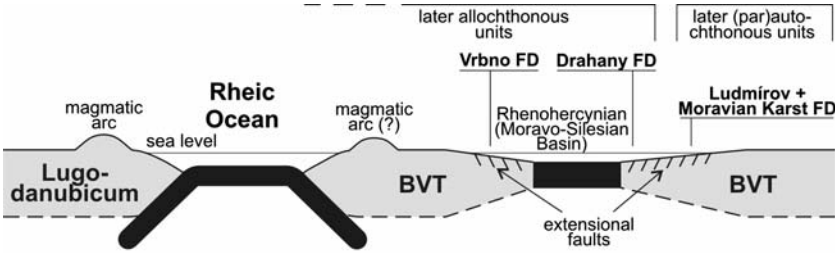
In the Devonian, tectonic extension predominated in most of the area. In the western marginal part of the Brunovistulian terrane, facing the Rheic Ocean (Fig. 4; Cocks and Fortey 1982), a back-arc extension is assumed. Patocka and Valenta (1996) and Patocka and Hladil (1997) outlined a model in which the volcanic rocks of the NW part of the Brunovistulian terrane originated in a volcanic arc geotectonic setting with a transition to a back-arc spreading. In the eastern interior part of the terrane, the extension is thought to have resulted from a slab pull with subsequent rifting connected with the collision and thrusting of the Lugodanubian group of terranes (Kalvoda 1995, 1998; Kalvoda and Melichar 1999).

During the rifting, old NW–SE trending basement faults parallel to the Teisseyre–Tornquist Zone and the Krakow–Lubliniec Zone were reactivated perpendicular to the Drahany (Rhenohercynian) aulacogen (Fig. 5; Hladil 1994; Hladil et al. 1999; Kalvoda 1998; Kalvoda and Melichar 1999). Relicts of the tectonically undisturbed sedimentary record deposited in extensional zones can be distinguished in the Nesvacilka, Rataje, Jablunka, Jablunkov and other grabens in the eastern part of the Brunovistulian terrane (Fig. 5). More to the W, tectonically detached segments of Devonian to Carboniferous sediments and their basement were incorporated in the complicated mosaic of the Moravo–Silesian shear zone (Fig. 2; Hladil 1994; Hladil et al. 1999; Kalvoda 1998; Kalvoda and Melichar 1999).

The extensional Moravo–Silesian basin was recently reconstructed by Kalvoda (1998) and Kalvoda and Melichar (1999) as a complicated structure that comprised NW–SE trending halfgraben subbasins, where more pronounced passive rifting took place in the west (Fig. 5) and where a marked W–E polarity can be distinguished. Five principal Devonian facies domains (facies developments following the terminology of Chlupac 1964) running parallel with the present day NNE–SSW and NW–SE tectonic strike were distinguished reflecting different geotectonic regimes. From the W to the E, they include the Vrbno and Drahany, Ludmírov (transitional), Tišnov and Moravian Karst (platform) facies domains (Figs. 4, 5; Chlupac 1965; Zukalova and Chlupac 1982; Hladil 1994; Hladil et al. 1999). The allochthonous Vrbno and Drahany facies and the parautochthonous Ludmirov facies (see Fig. 4) are preserved only in tectonic slices, the former incorporated in the metamorphic nappes of the Silesicum and the latter two in the Carboniferous flysch nappes (Chab et al. 1990; Kumpera and Martinec 1995; Chadima and Melichar 1998; Schulmann and Gayer 2000; Babek et al. 2006). The Tisnov facies restricted only to a small area (Fig. 2) and the western part of the Moravian Karst facies are incorporated in the nappe structure of the Moravicum and the Carboniferous flysch nappes. The eastern part of the Moravian Karst facies is in a (par)autochthonous position.

In all the facies domains except the Vrbno one (Figs. 4, 5), sedimentation started with Devonian basal clastics (Fig. 6) which have been recently interpreted by Kalvoda (1995) and Nehyba et al. (2001) as representing the initial phase of the passive rifting. In the Drahany facies domain, the basal clastics are represented by quartz sandstones, greywacke sandstones and oligomict to polymict con-

Fig. 4 Plate tectonic scheme showing the closure of the Rheic Ocean between the Brunovistulian terrane and Lugodanubian terranes and the facies of the Brunovistulian foreland basin in Devonian



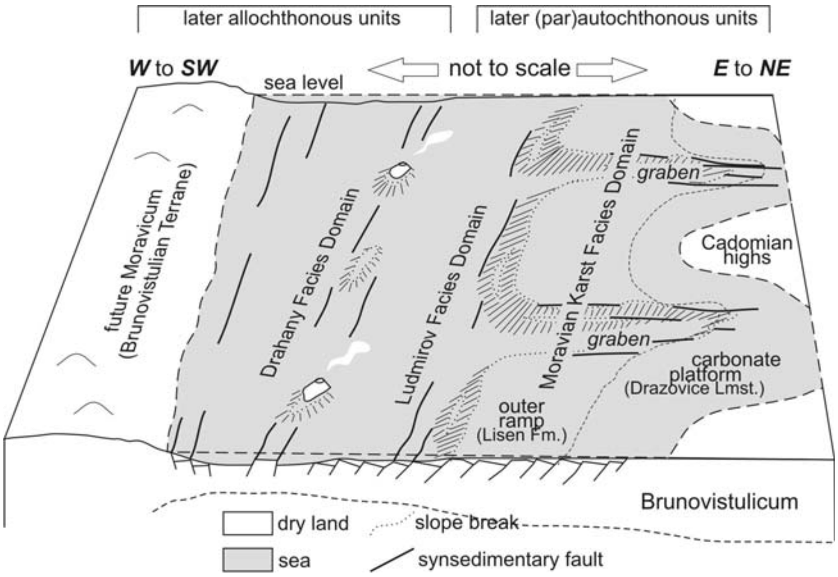


Fig. 5 Paleogeographic reconstruction of late Devonian rift-related sedimentary basins developing on the Brunovistulicum in the southern part of the Moravo–Silesian basin

glomerates with greywacke matrix (Chlupac and Svoboda 1963; Dvorak 1973), which yielded in their upper part early Devonian marine fauna (Chlupac and Svoboda 1963). Similar age is assumed for quartz conglomerates and sandstones and locally arcose sandstones in the Ludmirov facies domain (Chlupac and Svoboda 1963). In the Moravian Karst facies domain, two basic types of basal clastics, quartzose siliciclastics and polymict siliciclastics, have been recognized (Skocek 1980). The Devonian basal clastics are very similar to the Cambrian molasse (see above) and the available petrological, mineralogical and geochemical data show a similar or even identical provenance (Gilikova et al. 2003). Paleontological data from limestone intercalations (Zukalova 1976; Skocek 1980) in the upper part of ‘‘basal clastics’’ indicate early to middle Devonian ages. Facies and plant relics indicate deposition in alluvial to fluvial depositional environment. Towards the top of the succession, the depositional setting gradually transformed to a shallow marine environment (Nehyba et al. 2001).

In the early to late Devonian, deep-marine fossiliferous shales (Stínava–Chabicov Fm.), calciturbidites (Jesenec Limestone) and submarine bimodal within-plate volcanics with transition to oceanic crust (Prichystal 1990; 1993; Hanzl 1999) were deposited in the basinal Drahany (Rhenohercynian) facies domain located in the western part of the Moravo–Silesian basin (Figs. 4, 5, 6). A succession including Lower Devonian shallow marine quartzites (Drakov Quartzite), Emsian to Famennian black phyllites, metavolcanites–amphibolites and greenschists interfingering with Givetian to Famennian carbonates and radiolarian cherts (Fig. 6) are preserved in the Vrbno facies domain (Rhenohercynian basin; Fig. 4) in the NW part of the Brunovistulian terrane (Chlupac 1989; Hladil et al. 1987). Petrological study of metavolcanic suite indicates different tectonic settings in different locations of the Vrbno facies ranging from arc and back-arc within-plate volcanites (Patocka and Valenta 1990, 1996; Patocka and Hladil 1997; Janousek et al. 2006) to oceanic tholeites with transition to the continental tholeites associated with rift zones (Soucek 1981; Jedlicka and Pecina 1990).

During the same time interval, terrestrial siliciclastics accumulated in the Moravian Karst (platform) facies domain, followed by platform carbonates with coral-stromatoporoid reefs (Macocha Fm.). In contrast, at the transition between the platform and basin (Ludmírov facies domain) fossiliferous shales (Stínava–Chabicov Fm.), periplatform carbonates and carbonate turbidites (so-called equivalents of Macocha Fm.; Fig. 6) were deposited. These deposits are interpreted as indicating incipient crustal extension, platform drowning and basement subsidence (Babek 1996).

In the late Frasnian to late Tournaisian time, carbonate-free pelagic shales and radiolarian cherts with rare limestone intercalations (Ponikev Fm.), which were deposited in the both Drahany and Ludmírov facies domains, indicate further deepening, retreat of carbonate platforms source and/or local submersion under CCD, most probably due to advanced crustal extension and subsequent subsidence (Babek 1996). In the Frasnian to late Famennian interval, the carbonate platform in the Moravian Karst facies domain began to be progressively destroyed in an W–E direction and differentiated along NW–SE oriented half-grabens with the deposition of hemipelagic nodular limestones and carbonate turbidites (Lisen Fm) as the result of advanced crustal extension and block subsidence (Kalvoda 1998; Kalvoda and Melichar 1999). In the eastern part of the Moravian Karst facies domain on the slope of the Bohemian massif shallow water carbonate platform sedi-

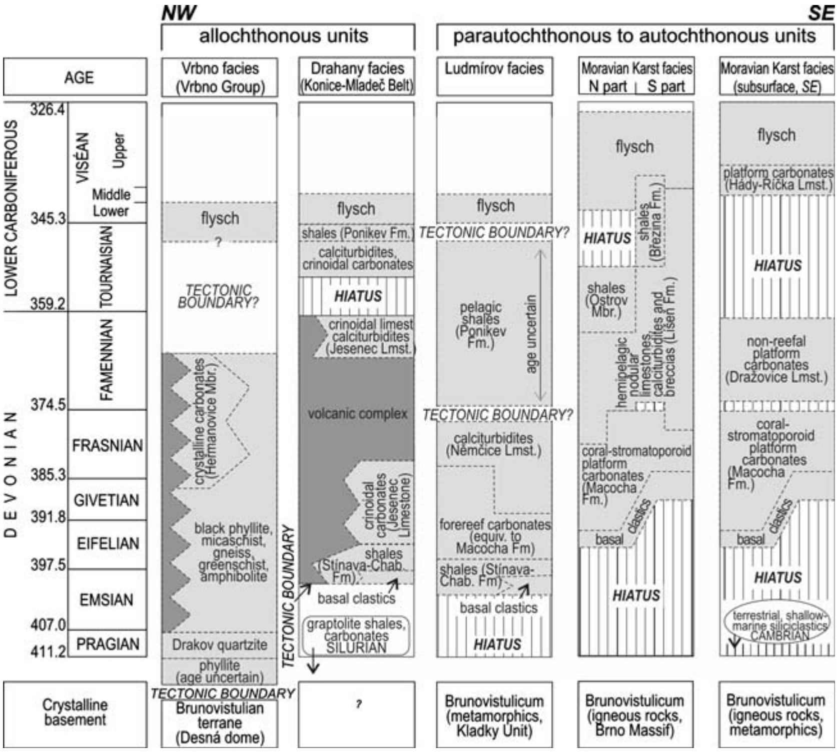


Fig. 6 Representative lithostratigraphic logs of the principal preorogenic facies domains and synorogenic clastics of the Moravo–Silesian basin

mentation (Drazovice Limestones) continued also in the Famennian (Figs. 5, 6). In the western part of the Moravian Karst facies domain, hemipelagic limestones and calciturbidites accumulated till the middle Tournaisian.

In the late Tournaisian to Viséan time, the deep-water carbonates of the Moravian Karst facies domain were replaced by deposition of sandy limestones, microbrecciated and brecciated limestones and flyschoid Brezina Formation (Fig. 6). Even more to the east, this interval corresponds to a prominent unconformity (Fig. 6; Kalvoda 1982a). At the same time, several carbonate platforms persisted in the northeastern portion of the basin in Poland (Narkiewicz 2005). In the Tournaisian to late Viséan interval, these platforms became progressively drowned in a W–E direction and carbonate platform sedimentation was replaced by basinal spiculitic wackestones and carbonate turbidites preceding the onset of siliciclastic flysch sedimentation (Belka 1987).

Compressional (flysch) phase

The final closure of the Rheic Ocean and the marginal Rhenohercynian within-plate to ocean basin (Prichystal 1990, 1993; Hanzl 1999; Fig. 4) was a compressive to transpressive event related to the accretion of the Armorican microcontinents to the Brunovistulian terrane. This convergence took place along an N–S trending dextral fault zone that marks the boundary between elements of the northern and southern shores of the ocean (Kalvoda 1995, 1998; Finger et al. 1998). In many places, the transition from an extensional to a compressional phase is documented by long-term hiatuses followed by the deposition of breccias and fossiliferous shales (Moravský Beroun breccia, Lisen Fm., Brezina Fm.; Figs. 6, 7). The breccias contain reworked platform carbonate clasts, reworked Upper Devonian and Tournaisian conodonts, phosphorite fragments and red-coloured quartz grains resembling those of the Lower Paleozoic ''basal clastics'' of the Brunovistulicum. The first occurrence of the breccias shows a strong diachroneity in the W/SW to E/NE direction (Kalvoda et al. 1999, Dvorak et al. 1987, Dvorak and Friakova 1978). The clast composition indicates a relative lowstand conditions related to basement uplift and erosion (Kalvoda et al. 1999). Based on their composition and diachroneity in the first occurrence, we consider the breccias to record the W–E and probably S–N propagation of a wave of tectonic uplift that preceded the major flux of siliciclastic flysch.

The onset of plate convergence and a compressional tectonic regime is indicated by the deposition of synorogenic, deep-marine siliciclastics (Variscan flysch or ''Culm facies''). The age of the synorogenic siliciclastics (Fig. 7) is early Viséan to earliest Namurian (Dvorak 1973; Kalvoda et al. 1995; Kumpera 1983; Spacek and Kalvoda 2000; Zapletal et al. 1989). In the past, Famennian and Tournaisian ages were claimed for the oldest flysch formation (Andelska Hora Fm.) by Dvorak (1994), but the evidence presented by the authors is problematic and does not correspond to biostratigraphic and geochronologic data

(Otava et al. 1994, Schneider 2002). The synorogenic siliciclastics are exposed in the Drahany Culm and the Nizky Jesenik Culm basins (Fig. 2). Two major tectonic units are distinguished, allochthonous unit deposited in the Rheic/Rhenohercynian remnant basin (Andelska Hora, Horni Benesov, Protivanov and partly Rozstani formations) and parautochthonous unit deposited in the Variscan foreland basin (Moravice, Hradec–Kyjovice, Myslejovice and partly Rozstani formations). In the foreland basin phase, the Drahany subbasin represented a proximal and the Jeseniky subbasin a distal section of the formerly united Moravo–Silesian Culm basin (Kumpera and Martinec 1995; Hartley and Otava 2001). The synorogenic sediments include polymict conglomerates, greywackes, quartzolithic and quartzofeldspathic sandstones, siltstones and mudstones deposited from turbidity currents, sandy and cohesive debris flows and hypopycnal flows. The parautochthonous flysch was deposited in elongated turbidite systems parallel to the NNE–SSW oriented basin axis. The sediments show cyclic stratigraphic arrangement, which resulted from pulsating tectonic activity in the hinterland and variations in the sediment supply from multiple point sources (Babek et al. 2004). Kumpera and Martinec (1995) interpreted the filling of both the Culm basins as a multiphase tectonic event resulting from plate convergence between the Lugodanubian group of terranes and the Brunovistulian terrane. Sediments of the first, remnant basin phase (Protivanov, Andelská Hora and Horní Benesov Fm.) are preserved in the western internal part, while deposits of the second, deep-marine foreland basin phase (Rozstani, Myslejovice, Moravice and Hradec–Kyjovice Fm.) are preserved in the eastern external part of the Culm basin (Fig. 7). Sandstone and conglomerate composition data from the foreland basin phase indicate increasing sediment supply from high-grade metamorphic and magmatic sources and decreasing supply from sedimentary, volcanic and low-grade metamorphic sources in upward and W–E directions. This trend is related to the uplift of the source area and the progressive unroofing of its structurally deeper crustal regions (Hartley and Otava 2001; Babek et al. 2004). In the upper parts of the flysch succession (Myslejovice, Moravice, Hradec–Kyjovice Fm.), detrital material with a typical Moldanubian provenance is abundant indicating the late stages of collision and underplating of the Brunovistulian terrane under the Lugodanubian terranes (Hartley and Otava 2001; Schulmann and Gayer 2000). Granulite pebbles in the Upper Viséan conglomerates of the Myslejovice formation witness very rapid exhumation and cooling of the Moldanubian orogenic root during Middle to Late Viséan, with minimum exhumation rates reaching 2.8–4.3 mm/year (Kotkova et al. 2003, 2007; Tajcmanova et al. 2006). In late Viséan times, the basin started to become overfilled (Hartley and Otava 2001; Babek et al. 2004), which eventually led to cessation of the deep-marine deposition and beginning of shallow-marine and continental sedimentation in the early Namurian time.

The synorogenic flysch facies pass gradually upward to Namurian to Westfalian paralic and continental coal-bearing molasse of the Upper Silesia basin (Ostrava and Karviná Formations) reaching up to 3,800 m of thickness.

The onset of the oldest synorogenic flysch in the Viséan in Moravia contrasts with the middle to late Devonian ages of its equivalents in the Rhenohercynian zone of SW England and Germany (Franke and Oncken 1995; Dörr et al. 1999) and may reflect the W–E diachronism in the closure of the Rheic/Rhenohercynian Ocean.

Permian record of the Variscan gravitational collapse

The formation of narrow grabens with Stephanian to Autunian sediments at the western margin of the outcropping Brunovistulicum and in the Lugodanubian hinterland

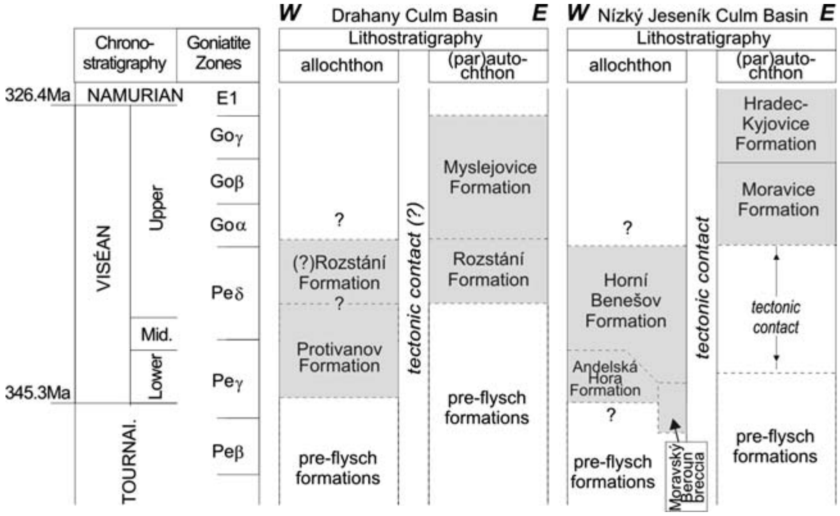


Fig. 7 Geologic sketch of Lower Carboniferous synorogenic clastics of the Moravo–Silesian basin

indicates a gravitational collapse and extension of thickened Variscan orogenic crust (Grygar and Vavro 1995). The possible post-collisional extension may be indicated also by the intrusions of basaltic dykes into the Permian sediments (Maly 1993; Prichystal 1994).

The youngest Paleozoic sedimentation of the Brunovistulian cover has been recorded in the Boskovice graben. The Boskovice graben is an SSW to NNE trending, elongated asymmetrical basin filled with Permo–Carboniferous terrestrial deposits (Fig. 8). The present-day dimensions of the basin are 5–12 km in width by 90 km in length, but the original area of the basin was larger (especially in the E–W direction). The maximum thickness of the basin fill is about 2,000 m and the post-Autunian denudation is estimated at a maximum of 500 m. The formation, development and deposition of the Boskovice graben were controlled by a major NNE-trending marginal dextral strike-slip fault. This basin was classified as a half graben with several stages of development (Mastalerz and Nehyba 1997). The first, extensional stage created space to accommodate Permo–Carboniferous deposits. This stage was followed by compressional deformation of the basin fill and westward thrusting of the Brno batholith (and locally its Devonian and Lower Carboniferous cover) over the eastern basin margin; however, the post-Paleozoic age (Alpine) of this deformation cannot be excluded. The interior part of the basin was overthrusted leading to the formation of duplexes. Transversal segmentation of the basin by NW-trending faults produced several ‘‘sub-basins’’ with a slightly different sedimentation history. Deposition started in the southern part of the Boskovice graben (the Rosice–Oslavany area) during the Stephanian C and spread towards the N and NE, where is also the deepest point of the basin. The termination of the basin filling was diachronous; sedimentation ended in the Early Autunian in the south, in the Early to Middle Autunian in the centre and in the Middle Autunian in the NE part of the basin (Jaros 1961; Maly 1993; Melichar 1995).

The basin has typically strongly asymmetric distribution of sedimentary facies and depositional settings (Fig. 8). Deposits of alluvial fans (breccias and conglomerates) represent the initial period of deposition. Afterwards, two different facies successions developed in the opposite (E–W) parts/limbs of the basin. In the eastern part, deposition of alluvial conglomerates continued up to the Autunian, whereas in the western part the conglomerates passed upward into a heterogeneous, generally more fine-grained fluvial, deltaic and lake deposits. Several coal seams developed in the Rosice–Oslavany depression. Repeated cyclic alternation of red and grey beds reflects the important role of climate in the deposition.

Lateral transport, indicated through provenance from the opposite basin margins (Moldanubian unit, Moravian unit in the W vs. Devonian to Lower Carboniferous sedimentary cover of the Brno Batholith in the E), turned to axial transport. The position and importance of the axial fluvial transport and lake evolution were strongly influenced by tectonic processes (Mastalerz and Nehyba 1997).

Tectonic structure

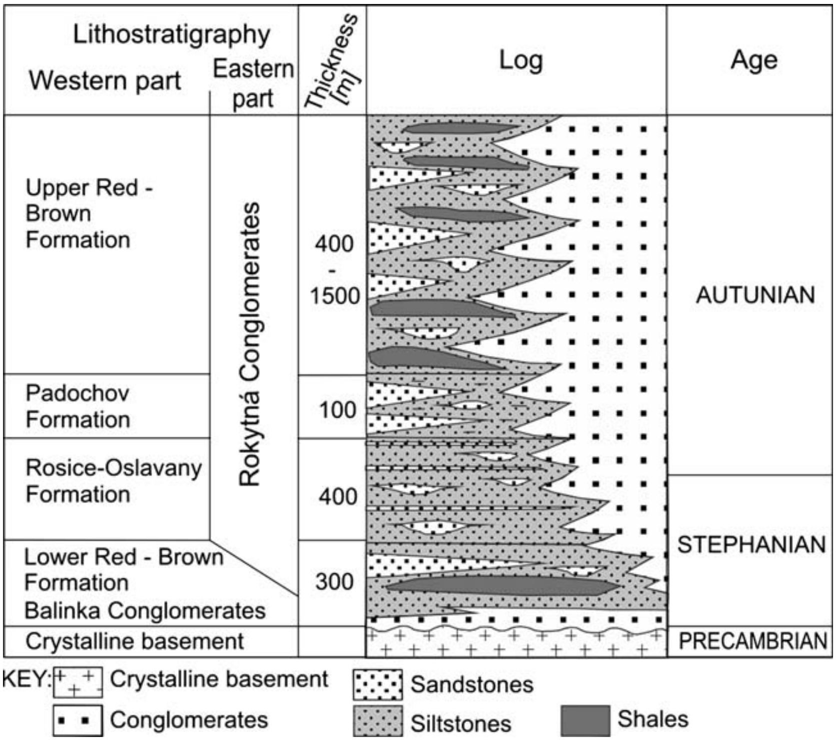
The Silurian to Westfalian rocks of the Brunovistulian terrane are preserved in a strongly imbricated stack of both E-vergent and W-vergent tectonic slices of the two major units—allochthonous and parautochthpnous ones (Figs. 2, 9). The progressive stacking of the nappes and their impingement on the Laurussian foreland led to crustal thickening and shortening, regional metamorphism and a number of distinct deformation and folding events (Schulmann et al. 1991; Grygar and Vavro 1995; Chadima and Melichar 1998; Schulmann and Gayer 2000; Havir 2000). A certain E–W polarity in the tectonic and thermal regime can be observed. The general trend is one of the increasing grades of strain and thermal overprint from late diagenetic conditions (<200°C) in the distal foreland in the SE to the high-temperature metamorphic conditions and ductile deformation in the westernmost part. In most of the area, however, temperatures were generally lower than 300°C (anchizone) and primary features of the rocks were preserved (Francu et al. 2002).

The western part of the Brunovistulian terrane, including the preorogenic Drahany facies domain, the Silurian exotics and the synorogenic Protivanov, the Andelská Hora and Horni Benesov Formations (Fig. 7) are clearly allochthonous (Chab et al. 1990; Hladil et al. 1999; Schulmann and Gayer 2000; Babek et al. 2006). These units represent a relic of a lower to middle Viséan remnant basin incorporated into an accretionary wedge (Kumpera and Martinec 1995, Grygar and Vavro 1995; Francu et al. 2002; Babek et al. 2006).

The eastern part of the Brunovistulian terrane including the preorogenic Ludmirov and Moravian Karst facies domains, the synorogenic Myslejovice, Rozstani, Moravice and Hradec–Kyjovice formations and the sediments of the Upper Silesia basin represents a parautochthonous thin-skinned stack and, for a small section, possibly a true autochthon (Chab et al. 1990; Cizek and Tomek 1991; Babek et al. 2006). In this part, the Variscan thrusting ceased in the Late Westfalian to Early Stephanian interval (Grygar and Vavro 1995).

This tectonic style is generally considered to be the result of Variscan oblique plate convergence between the overriding Lugodanubian group of terranes and the subducted Brunovistulian terrane (Dallmeyer et al. 1992; Fritz and Neubauer 1995; Kalvoda et al. 2002, 2003).

Fig. 8 Lithologic and lithostratigraphic log of Permian sediments of the Boskovice graben



The NNE structural trend of the Brunovistulian terrane shows approximately 90° deflection from the E trend of the central Rhenohercynian zone. It was explained by Devonian orogenic bending (clockwise rotation) of the Brunovistulian terrane with respect to the Rhenohercynian Belt in Germany by Hladil (1995), Krs et al. (1995) and Tait et al. (1996). More recently, the orogenic bending was modelled by Franke and Zelazniewicz (2002). In the model of Kalvoda (1995) and Kalvoda and Melichar (1999), the deflection resulted from transpressional rotation of individual tectonic slices during nappe stacking in the W part of the Brunovistulian terrane while in the E, the original Devonian and Carboniferous trends of facies zones remained intact. Finger and Steyrer (1995) related the southward bending of the Variscan fold belt in Moravia to considerable indentation of the Moldanubian terrane. Hladil et al. (1999) reinterpreted the paleomagnetic data as indicating strong clockwise tectonic rotation and wedging of individual massifs. Edel et al. (2003) stressed that the magnetizations in the Devonian units are likely to be Carboniferous overprints and contradicted the views of Devonian rotation of the rigid Brunovistulian promontory published by previous authors. Grabowski et al. (2004a, b) reported a major syn- and postfolding, late Variscan remagnetization and a minor, residual, pre-325 Ma magnetic component from the BVT; the latter accounting for about 60° clockwise rotation of the BVT with respect to the Old Red continent. The same authors stressed a need for reevaluation of the orogenic bending hypothesis.

During the late phases of the collision (330–310 Ma), the ongoing plate convergence resulted in a transition from compressive to transpressive tectonic style and a major zone of dextral shearing, the Moravo–Silesian Shear Zone (Fig. 2) formed between the colliding terranes (Rajlich 1990; Schulmann and Gayer 2000). The transpression led to imbrication and uplift of the western parts of the Brunovistulian terrane (Fig. 9) under medium- to low-temperature conditions (Schulmann and Gayer 2000; Stipska and Schulmann 1995) and it was associated with the development of NNE-trending stretching lineations and asymmetric structures indicating top-to-NNE shearing.

Paleobiogeography of the Brunovistulian terrane

Different and often conflicting paleogeographical positions of the Brunovistulian terrane have been suggested in the literature (Fig. 10; Moczydlowska 1997, 1998; Belka et al. 2000, 2002; Cocks 2002; Nawrocki et al. 2004a, b). Moczydlowska (1995a; b, 1997, 1998) analysed acritarch assemblages from three boreholes that reached the Lower–Middle Cambrian sequence in northern part of the Brunovistulian terrane in Poland and defined the Upper Silesian terrane as a distal segment of the eastern Avalonia (cf. Fig. 10). On the other hand, Belka et al. (2000, p. 94; 2002, p. 30) considered the Lower Cambrian trilobite genera described by Orlowski (1975, 1985) from the northern part of the Upper Silesian Block as indicating a peri-Baltic affinity

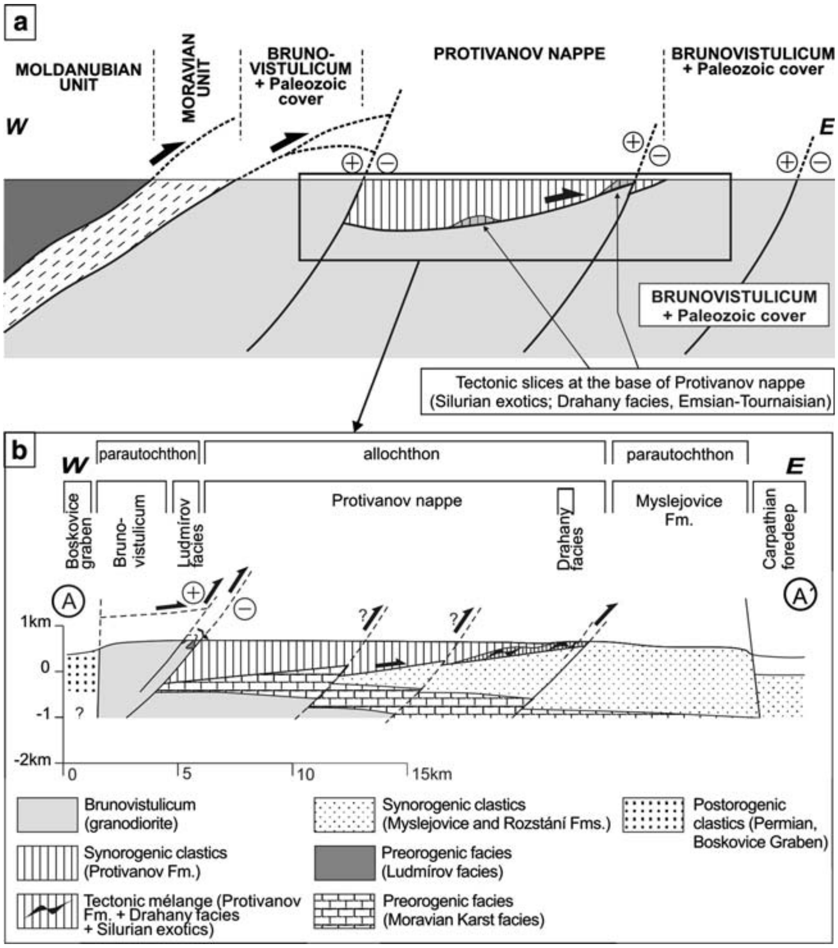


Fig. 9 **a** Simplified tectonic cross-section showing tectonic relationships between autochthonous and allochthonous units in the souhern part of the Brunovistulian terrane; **b** Geologic cross-section of the Drahany Upland. For location refer to Fig. 2

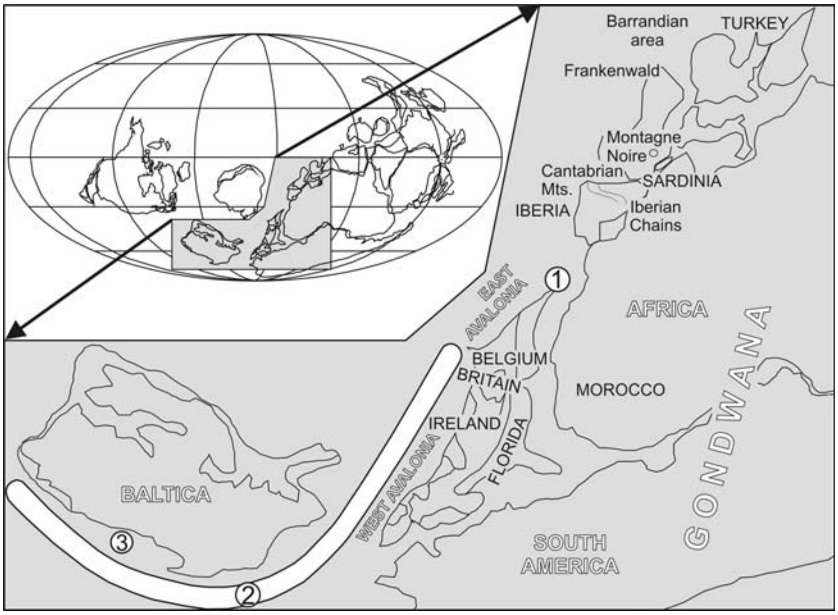


Fig. 10 Paleogeographical map showing possible position of the Brunovistulian terrane in Cambrian after various authors: *1* Moczydlowska (1995a, 1995b, 1997, 1998); *2* Fatka and Vavrdova (1998); *3* Winchester et al. (2002), Vavrdova et al. (2003), Nawrocki et al. (2004a, b). Map modified after McKerrow et al. (1992) and Courjault-Radé et al. (1992). *White band* illustrates the range of possible location of the Brunovistulian terrane

for this terrane. Based on the study of early Cambrian microflora the paleogeographic position was also discussed in the southern part of the Brunovistulian terrane. Fatka and Vavrdova (1998), Vavrdova et al. (2003) and Vavrdova (2006) mentioned relations to the Baltoscandinavian Early Cambrian microflora. Similarly, Mikulas and Nehyba

(2001) correlated the intensity of bioturbation and the ichnofabric pattern of lower Cambrian sediments in the Moravian part of the Brunovistulian terrane and East European Platform. In each of their seven paleogeographical maps covering the Late Proterozoic to Early Carboniferous period, Winchester et al. (2002) first placed the Brunovistulian terrane (indicated as BM = Bruno–Silesia–Moesia) in a position between the Baltica and peri-Gondwanan terranes (end of Proterozoic to Early Cambrian). During the Middle Cambrian to the Ordovician, the Brunovistulian terrane separated from the Gondwana but it was still attached to Baltica (Nawrocki et al. 2004b) SE of its recent position, while during the Silurian it started moving more or less along the southern margin of the Baltica. In this model, the Brunovistulicum acted as a bridge between the Baltica and peri-Gondwana, never moving far from Baltica. Cocks (2002) provided a short but comprehensive evaluation of the Cambrian trilobites and inarticulate brachiopods established in the Malopolska and Lysogory blocks. He summarized his critical paleontological data in his conclusions: ‘‘There seems little doubt that neither the Malopolska nor the Lysogory blocks formed part of the same terrane in the Lower Paleozoic as the Bruno–Silesian Block. However, the faunal evidence from the latter is currently inconclusive as to whether the latter formed part of Baltica or Gondwana or was a separate and independent terrane’’ (Cocks, 2002, p. 44). In their recent discussion on the trilobites described by Orlowski (1975) from the Goczalkowice IG-1 borehole, Nawrocki et al. (2004a) confirmed the presence of the Baltic endemic genus *Schmidtiellus*, which is associated with the paleogeographically more widely distributed genera *Ornamentaspis* and *Strenuaeva*.

Silurian fauna was described from graptolite shales at Stinava (Figs. 2, 3) by Kettner and Remes (1936) and Kraft and Marek (1999), which form a tectonic slice at the base of the Protivanov Formation thrust (Fig. 9; Kettner 1966; Chadima and Melichar 1998; Babek et al. 2006). Similar facies are widespread in the Barrandian area (Chlupac et al. 1992) and other Armorican terranes. However, graptolite shales also occur in the SW part of Laurussia in Poland (Malopolska terrane) and their graptolite associations show a resemblance to both Avalonia and Armorica (Masiak et al. 2003).

According to Hladil and Bek (1999) and Hladil et al. (1999), the Early to Middle Devonian shallow water coral fauna of the Brunovistulian terrane is quite dissimilar to the associations of the Barrandian area – a representative of the Armorican Terrane Assemblage, in contrast with the plant spores, although the Barandian area was already located fairly close to the southern Laurussian margin. In their interpretation, the Ibermaghian faunas of the Barrandian area show close links to the other southern terranes located at the southern Rheic margin and NW Africa. They hypothesized that a land barrier producing populations of Laurussian plant spores may have prevented the migration of shallow water marine fauna directly across the Rheic remnant ocean.

The Late Devonian and Early Carboniferous paleobiogeographic data are based mostly on calcareous foraminifers. The Brunovistulian terrane contains highly diversified foraminiferal fauna, which is typical for the tropical/subtropical Paleotethyan Realm (Mamet and Belford 1968). Its clear affinity to the East European Platform was already recognized by Kalvoda (1982b). Kalvoda (2001) defined the Fennosarmatian Province to include the East European Platform, the Urals and the accreted pre-Variscan Brunovistulian group of terranes, which included the Malopolska, Brunovistulian, Moesian and Istanbul terranes (Fig. 1, 11) and, with some reservation, also the eastern Avalonia. In SW and W Europe, he defined the Armorican Province to include the Variscan peri-Gondwana terranes. This province is characterized by incomplete foraminiferal phylogenies and foraminiferal assemblages with lower diversity than those of the Fennosarmatian Province. The differences, however, decreased progressively during the Viséan. In this respect, the Brunovistulian terrane was in the same geotectonic position as the eastern Avalonia, i.e. on the northern margin of the closing Rheic Ocean in the Devonian. Although paleomagnetic data demonstrate that the final consolidation of the Armorican Terrane Assemblage with the Laurussia occurred in the Late Devonian (Tait et al. 1997), the terranes located originally on the southern and northern margins of the Rheic Ocean interestingly show different paleobiogeographic affinities to the Armorican and Fennosarmatian provinces even at the beginning of the Carboniferous.

Discussion

Proterozoic to early Paleozoic record

There exists a broad similarity in the geological structure, lithology and geochronology between the Cadomian crystalline basement of the Brunovistulian terrane and the Istanbul Zone, supported by their similar Devonian and Carboniferous tectonosedimentary record and paleobiogeography (Kalvoda 2001, 2002, Kalvoda et al. 2002, 2003). The Proterozoic affinity and evolution of the Brunovistulian terrane remains open. In the view of Finger et al. (2000a) and Friedl et al. (2000, 2004), petrological data seem to fix its position within the Amazonian part of the Pan-African orogenic belt. On the basis of the lithological and structural evidence, Leichmann (1996) advocated its similarity with the Eastern Desert of Egypt. On the other hand, there is also a good fit with the geochronological data

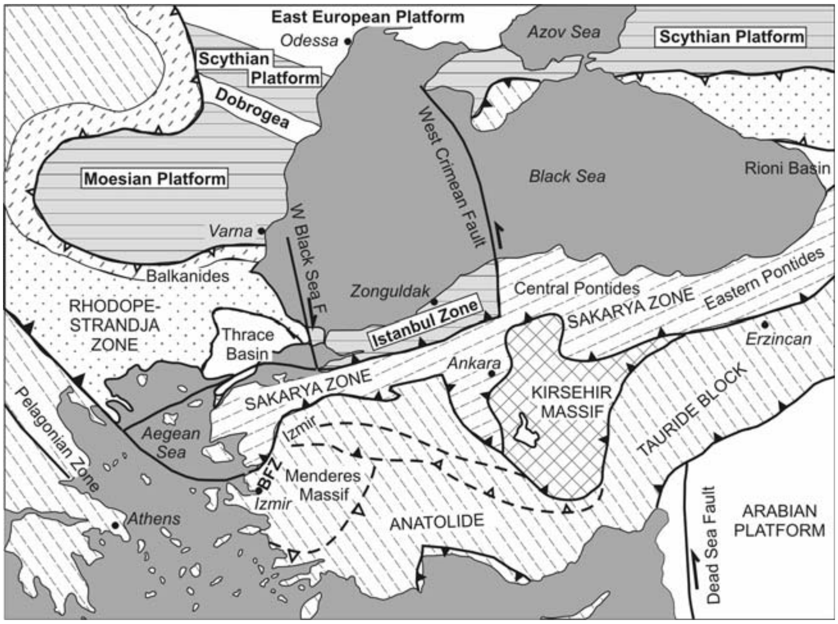


Fig. 11 Tectonic map showing the location of the Moesian terrane and Istanbul Zone. Before the opening of the Black Sea the Istanbul zone was situated originally along the Odessa shelf of the East European platform. Modified after Okay and Tüzsüz (1999)

from the Urals (Scarrow et al. 2001; Gee 2001; Glassmacher et al. 1999) and, accordingly, some authors prefer the Baltic affinity of the Brunovistulian terrane (Pharaoh 1999; Zelazniewicz et al. 1997, 2001). According to Pharaoh (1999), terranes with the Cadomian basement originating in the vicinity of the Urals may have dispersed dextrally along the Tornquist margin of Baltica in the Early Paleozoic.

Unfortunately, the available paleomagnetic studies offer diverse Late Proterozoic to Cambrian paleogeographic reconstructions (Hartz and Torsvik 2002; Torsvik and Rehnström 2001; Nawrocki 2004a, b; Popov et al. 2002) leaving the precise location of the Brunovistulian terrane open to interpretation. The Cambrian paleobiogeographic data, however, suggest its proximity to Baltica (Fatka and Vavrdova 1998; Vavrdova et al. 2003; Belka et al. 2000; Winchester et al. 2002).

Summarizing, even though there may be the best fit of the Vendian petrological data with the Amazonia, the Baltoscandinavian affinity of microflora and ichnofauna in the lower Cambrian molasse of the Brunovistulian terrane is hard to reconcile with the model of the Vendian location in the Pan-African belt of Gondwana. Consequently, both paleobiogeographic, petrologic and paleomagnetic interpretations require a careful scrutiny.

Origin of the Silurian exotics and the lower Paleozoic detrital micas

The provenance of the Silurian pelagic graptolite shales and limestones near Stínava, the only sedimentary record before the early Devonian opening of the extensional basins on the Brunovistulian basement in Moravia, is an enigma. We can propose three models for their origin. In the first model, the Silurian facies were deposited in a back-arc basin of a hypothetical, consumed Brunovistulian magmatic arc. This may be supported by pebbles from the andesite-dacite-rhyolite volcanic suite and granites with assumed Brunovistulian affinity present in the Lower Viséan Korenec conglomerate of the Protivanov unit (Zachovalova 2003), the magmatic-arc suite of tholeitic arc basalts, low-K calc-alkaline andesites and metakeratophyres present in the Devonian Vrbno facies domain (Patocka and Hladil 1997; Janousek et al. 2006) and Silurian magmatic activity in the Brunovistulian terrane (Prichystal 1999). The presence of the Silurian volcanic back-arc has also been discussed for the southern margin of the Rhenohercynian Zone in Germany (Meisl 1990; 1995, Franke et al. 1995) and in both cases, the arcs may record the progressive closing of the Rheic Ocean.

In the second model situation, the Silurian facies were thrust over the Brunovistulian terrane during an early Variscan docking, by analogy with the Armorican units at the base of the Giessen–Harz nappe (Fuchs 1976; Franke and Oncken 1995; Dörr et al. 1992a, b; Dörr and Franke 2002), and the clastic lower Paleozoic material of the Protivanov Fm. was derived from a colliding Lugodanubian (Armorican) upper plate.

In the third situation, the synorogenic flysch of the Protivanov Formation as well as Silurian rocks were deposited at the margin of the Rheic Ocean (Letovice Ocean of Höck et al. 1997) between the Brunovistulian and the Lugodanubian terranes (Fig. 4). In this case the association with the Drahany facies domain may be a result of the tectonic stacking.

The presence of the Ordovician, Silurian and early Variscan (391–371 Ga) detrital micas in the synorogenic siliciclastics of the allochthonous Protivanov, Andelska Hora, Horni Benesov formations and the Mirov unit (Schneider 2002) may indicate derivation from either the Brunovistulian magmatic arc (now eroded or buried under the Moldanubian nappes) or the upper plate of the Lugodanubian (Armorican) group of terranes or both. Similar ages of the major episodes of granitoid magmatism have been recognized in the Saxothuringian Zone (Dörr et al. 1992a, b) and in the eastern part of the Armorican Mid-German Crystalline Rise (Anthes and Reischmann 2001). Ordovician, Silurian and early Variscan ages of detrital micas and zircons derived from the Tepla–Barrandian terrane are also reported from the Saxothuringian flysch (Schäfer et al. 1997).

There is growing evidence that the Brunovistulian terrane was already accreted to Baltica in the Cambrian (Fatka and Vavrdova 1998; Vavrdova 2006; Belka et al. 2000, 2002; Nawrocki et al. 2004a; Winchester et al. 2002). In this case the presumed early Paleozoic magmatic activity in the Brunovistulian terrane cannot reflect rifting and crustal thinning at the northern Gondwana margin as was the case with the Armorican terranes. Together with the presence of detrital spinels with a back-arc MORB affinity, whose provenance Copjaková et al. (2005) attributed to the Letovice–Rehberg Ocean (part of the Rheic Ocean), the white micas present in the Protivanov, Andelska Hora and Horni Benesov formations testify to the eroded levels of the Lugodanubian and/or Moravian–Silesian nappes.

Comments on the Variscan structure of the Brunovistulian terrane

There are differences in the large-scale internal architecture between different marginal parts of the Brunovistulian terrane. In its southern part (Drahany Upland), the Variscan continental subduction of the Brunovistulian terrane under the Moldanubian terrane was relatively shallow, contributing to widespread thin-skinned tectonic style of the Moldanubian nappes (Schulmann et al. 1991). A belt of highly correlating gravity and magnetic anomalies in the southeastern part of the Bohemian Massif (Gnojek and Hubatka 2001; Bielik et al. 2006; Lenhardt et al. 2007) indicates the possible continuation of the Brunovistulian terrane up to the Pribyslav and Vitis zones, i.e. about 30–60 km west from its boundary on the surface. In the northern part (Jeseníky Mts.), the angle of continental subduction was steeper and the bulldozing effect of the Keprník and Desná units (‘‘crustal boudins’’) contributed to markedly greater shortening and propagation of the Culm nappes much further to the east than in the southern Drahany area (Kumpera and Martinec 1995; Schulmann and Gayer 2000). Here, the subsurface limits of the Brunovistulian terrane are again indicated by a sharp step in magnetic and gravity fields in the Cervenohorske sedlo zone (Dudek 1980; Gnojek and Hubatka 2001; Bielik et al. 2006), even though on the surface the Brunovistulian terrane can be traced even more to west in the Keprnik nappe (Schulmann and Gayer 2000).

The collision between the Lugoanubian terrane and the Brunovistulian Terrane, connected with the closure of the Rheic Ocean (Fig. 4), is thought to have commenced during the early Carboniferous (Schulmann and Gayer 2000). This view may be supported by the age of the Letovice ophiolite amphibolites ranging from 354 to 328 Ma (MacIntyre et al. 1993) and tectonic incorporation of the upper Devonian limestones in the nappes (Hladil et al. 1987). An alternative interpretation regarded the diastrophic late Devonian Mohelnice Formation (Mirov Culm) located in hangingwall of the Moldanubian overthrust as synorogenic sediments connected with the closure of the Rheic Ocean in late Devonian (Kalvoda 1995, 1998; Hladil et al. 1999). However, new data on early Carboniferous cooling ages of detrital white micas ranging from 362.6 to 344.9 Ma (Schneider 2002) put the previously assumed Devonian age of the Mohelnice Formation in question. To sum up, the Rheic Ocean between the Lugodanubian and the Brunovistulian terrane (Fig. 4) was presumably closed in the early Carboniferous, but due to uncertain tectonostratigraphic and paleogeographic affinities of both the Mírov Culm and the Letovice ophiolite this question remains still open (Franke and Zelazniewicz 2002; Misar et al. 1984; Mísar and Dudek 1993).

Devonian and Carboniferous record and the correlation with the Rhenohercynian zone in Germany

As it was already discussed in detail by Kalvoda et al. (2002, 2003), in the Devonian to the Permian, the Brunovistulian terrane showed close ties both in terms of its sedimentary and faunistic records to the East European Craton, the Moesian terrane and terranes of the Istanbul Zone (Fig. 11) where the continuation of the Rhenohercynian zone to the east can be anticipated.

The Devonian preorogenic and Carboniferous synorogenic sedimentation also bears close similarity to the Rhenohercynian Zone of Germany. As both regions had a different Late Proterozoic to Early Paleozoic history (Franke 1989; Meissner et al. 1994; Franke 1995a; Kalvoda 1995; Finger and Steyrer 1995; Pharaoh 1999; Belka et al. 2000, 2002; Winchester et al. 2002; Kalvoda et al. 2002, 2003), the parallelism reflects the similar geotectonic regime of the passive continental margin of Laurussia (the lower plate with respect to the active margins of the

Armorican Terrane Assemblage; Oncken and Weber 1995; Oncken et al. 1999; Tait et al. 1997; Babek et al. 2006). In both areas, crustal extension and differential subsidence were accompanied by felsic and basic within-plate volcanism (Floyd 1995), contributed to the development of structural lows and rise in a system of halfgrabens (Engel et al. 1983; Franke 1989; Lütke 1990; Franke 1995b; Oncken and Weber 1995; McCann 1999; Kalvoda 1998; Kalvoda and Melichar 1999) and eventually led to pelagic sedimentation (''Herzynische Fazies''). Nevertheless, there are also some differences. While the northern highs of the Caledonian Old-Red continent furnished most of the Rhenohercynian basin with siliciclastic input throughout the Devonian (Engel et al. 1983; Franke 1995b), in the Moravo–Silesian basin the coarse siliclastic input was much lower, related to recycling of materials from the Brunovistulian basement during rifting (Kalvoda 1995; Nehyba et al. 2001) and limited thus only to the basal parts of the Devonian sequences.

The Devonian reefoid limestones and Famennian to Viséan hemipelagic limestones and calciturbidites of the western part of the Moravian Karst facies domain, located on a more mobile basement, can be correlated with the Velbert and Stavelot–Venn anticlines (Franke et al. 1975; Dreesen et al. 1985) while the eastern part of the Moravian Karst Facies with the northern part of the Rhenohercynian autochthon of Germany (McCann 1999).

The facies of the preflysch phase of the Rhenohercynian autochthon in Germany, characterized by the middle–late Devonian reef growth and the Famennian to early Carboniferous hemipelagic limestones, pelites and cherts (Engel et al. 1983; Bender et al. 1993; Franke 1995b) resemble the Ludmirov transitional facies. In particular, the absence of volcanites suggests resemblance with the sedimentary record W of the river Rhine (Franke 1995b).

Most of the lithostratigraphic units of the Brunovistulian cover correspond to the parauthochtonous units of the Rhenish Massif (Fig. 12). To find analogies to the allochthonous Giessen–Harz nappe and other units (Dörr and Preiss 1982; Dörr 1986, 1990, 1998; Stibane et al. 1984; Franke 1995b; Dörr and Franke 2002) is very difficult, as the history of the southern margin of the Rhenohercynian Zone in Moravia is not as well known as it is in Germany. In Germany, the existence of the Rhenohercynian (Lizard–Giessen–Harz) Ocean is well documented by the MOR-type basalts at the base of the Giessen–Harz nappe (Grösser and Dörr 1986; Platen et al. 1989; Dörr 1990) and the presence of the enigmatic Mid German Crystalline Rise (MGCR), which formed its southern active margin (Franke and Oncken 1990; Oncken and Weber 1995; Franke 2000). While the Rheic Ocean between the Laurussia and the Armorican terrane assemblage closed in Germany in late Silurian–early Devonian (Franke et al. 1995; Tait et al 1997), this closure most probably did not occur before the earliest Carboniferous at the eastern margin of the Bohemian massif (Schulmann et al. 1991; Mazur et al. 2006).

In Moravia, the attenuated continental crust with narrow segments of oceanic crust of the Vrbno and Drahany facies (Prichystal 1993), analogous to the crust of the Rhehohercynian Ocean in Germany, seems to be connected with bac-arc rifting at the margins of the Brunovistulian terrane (Patocka and Valenta 1996; Janousek et al. 2006) related to the subduction of the Rheic Ocean beneath the Brunovistulian terrane (see Fig. 4). This is in accord with the conclusions of Konopasek et al. (2002) who argued for a bivergent subduction of the Rheic Ocean (Matte et al. 1990) at the eastern margin of the Bohemian massif. A similar scenario is also assumed in the Sudetes by Mazur et al. (2006).

In the Drahany facies, the tectonic slices of the lower Devonian to Tournaisian fossiliferous shales (often with Bohemian affinity, Chlupac et al. 2002), crinoidal and hemipelagic limestones and shales with radiolarian cherts, all alternating with submarine volcanics associated with the allochthonous flysch nappes, correlate with the tectonically displaced slices of the Rhenohercynian allochthon (Franke 1995b) rather than with the more distal autochthonous sequences of the Dill and Lahn synclines (Fig. 12). In our assumption, the Drahany facies was deposited on a thinned passive southern margin of the Brunovistulian terrane (Figs. 4, 5), where narrow segments of oceanic crust are anticipated (Prichystal 1993). The structural position of the

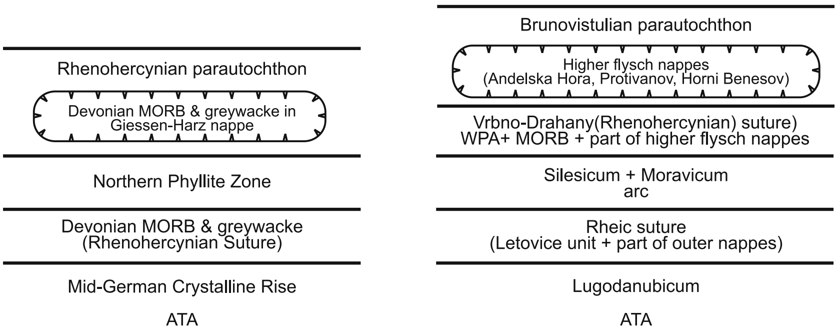


Fig. 12 Diagrammatic representation of tectonic relationships at the contact of the Armorican terrane assemblage and Laurussia in Germany and Czech Republic. German part modified after Franke (2002)

Silurian exotic sediments and the incomplete, base- and top-cut-off successions of the Devonian Drahany facies is best documented in the central part of the Drahany Upland where tectonic shavings carried along the base of Protivanov nappes are interpreted, based on detailed mapping and structural studies of the Drahany Upland (Chadima and Melichar 1998; Babek et al 2006; Figs. 3, 9).

The record of the Vrbno facies represents a part of the thickened, underplated parautochthonous Keprník and Desná domes of the Silesicum (Schulmann and Gayer 2000). In many respects it may be similar to the data from the Northern Phyllite Zone of the southernmost part of the Rhenohercynian belt (Klügel et al. 1994) where both WPB and MORB are indicated with possible Late Silurian–Early Devonian volcanic arc (Meisl 1990; Meisl 1995; Floyd 1995) interpreted by Franke (2000) as a part of the Armorican Terrane Assemblage. The oldest member of the Vrbno facies, the metamorphosed shallow marine Drakov quartzite represents the ''Rheinische Fazies'' (Langenstrassen 1983) and shows close ties to the Taunus quartzite or the Herdorf group of the Ardenno–Rhenish area (Chlupac 1975, 1981, 1989). Other parallels to the Northern Phyllite Zone may be seen in the metamorphosed Velke Vrbno upper allochthon at the margin of the Silesicum (Schulmann and Gayer 2000) which contains limestones of presumably early Devonian age (Hladil and Cejchan 1994; Hladil et al. 1999) and the possible presence of Silurian shales and Ordovician quartzites accreted to a volcanic arc during the Devonian is anticipated (Hladil et al. 1999). Such a sequence may suggest a Bohemian and Armorican rather than Brunovistulian provenance (Hladil et al. 1999); however, Kröner et al. (2000) include at least a part of Velke Vrbno high-grade rocks in the Brunovistulian terrane. Consequently, the faunistic, lithologic and tectonic record of the Silesicum may resemble the situation in the Northern Phyllite Zone where the slices of both Avalonian and Arnorican units are juxtaposed (Franke 2002).

The onset of a compressional tectonic regime and the transition to synorogenic sedimentation shows a distinct heterochroneity and polarity from the middle Tournaisian in the W (Drahany facies domain) to the late Viséan in the E (eastern part of the Moravian Karst facies domain). It coincided with the beginning of the tectonic emplacement of the Lugodanubian nappes in the W about 350–340 Ma (Schulmann et al. 1991; Hartley and Otava 2001; Schulmann et al. 2005). The progradation of the synorogenic sediments was controlled by the advancing active margin, in a manner similar to the Rhenohercynian belt.

The MORB-type volcanics, the Bohemian-type Silurian and Devonian fossils, the Hercynian facies and the Ordovician, Silurian and early Variscan white micas (Schneider 2002) associated with the structurally highest flysch Protivanov, Andelska Hora and Horni Benesov nappes speak in favour of their correlation with the Rhenohercynian allochthon (Fig. 12; Franke 1995b; Dörr et al. 1999). This interpretation is reinforced also by the overthrust of the Andelska Hora nappe over the Vrbno unit of the Silesicum (Chab 1990; Schulmann and Gayer 2000); however, there are some differences. Other evidence such as the later onset of synorogenic clastic sedimentation (and higher representation of calciturbidites may favour a correlation with the Rhenohercynian autochthon of Lahn and Dill area (Birkelbach et al. 1988; Bender et al. 1993; Dörr and Franke 2002). In our opinion, however, the arguments for the correlation with Rhenohercynian allochthon are stronger and the age differences can be explained by the separate evolution of the Brunovistulian and the Avalonian terrane during the Variscan plate convergence (Kalvoda et al. 2002, 2003).

Conclusions

1. The affinity of the late Proterozoic Brunovistulian terrane is difficult to assess. The Cambrian lithological and paleontological data show possible close ties to the Malopolska Terrane and East European Craton.
2. The middle to late Paleozoic evolution shows close ties to the East European Craton, especially to the terranes located at its eastern margin, e.g. the Moesian and terranes of the Istanbul zone (Fig. 11).
3. So far, there is no evidence for the extension of the Mid-German Crystalline rise as far as the eastern margin of the Bohemian Massif. The outermost part of the Brunovistulian terrane is distunguished in the units of the Silesicum, the Desna nappe with the Vrbno facies domain, the Keprnik nappe with its Devonian sedimentary cover and, partly, the Velke Vrbno unit, which presumably contains tectonic slices of both the Brunovistulian and Armorican provenance. Consequently, the Silesicum is correlated with the Northern Phyllite Zone at the sourthern tip of the Avalonian terrane (Fig. 12).
4. The early Carboniferous closure of the Rheic Ocean by bivergent subduction seems to offer a simpler scenario for the tectonic evolution at the margin of the Brunovistulian terrane than at the margin of Avalonia (Fig. 4).
5. In the Devonian, we can distiguish a period of extension/transtension, which was connected with basin differentiation into horsts and grabens. It was the result of the Devonian rifting, which led to the development of attenuated continental crust with narrow segments of oceanic crust in the marginal parts of the Brunovistulian terrane (Rhenohercynian basin; Figs. 4, 5). A significant role of back-arc rifting related to the closure of the

Rheic Ocean is supposed especially in early and middle Devonian. The proposed presence of an early Paleozoic active Brunovistulian margin, now destroyed or covered by the Lugodanubian nappes, and the presence of a Silurian back-arc basin needs further investigation.

6. The extension phase was followed by a compression/transpression phase and the deposition of Viséan to lowermost Namurian synorogenic flysch (Culm facies) and Namurian to Westphalian paralic and limnic molasse in remnant and foreland basins.
7. The oldest allochthonous flysch nappes (Andelska Hora, Protivanov, Horni Benesov formations) associated with the Drahany preorogenic facies were derived from the Rheic/Rhenohercynian remnant basin (Fig. 4) while the younger parautochthonous formations (Myslejovice, Moravice, Hradec–Kyjovice) associated with the Ludmirov and Moravian Karst preorogenic facies were deposited in the Variscan foreland basin.
8. The latest Paleozoic record can be traced in the uppermost Carboniferous to lower Permian postorogenic sediments of the extensional/transtensional Boskovice Graben.
9. The similar geotectonic position of the Brunovistulian and the Avalonian terranes at the southern margin of Laurussia contributed to similar tectonostratigraphic development of the Devonian and Carboniferous of the ‘‘Rhenohercynian Zone’’ whose extension is anticipated in the Istanbul terrane of NW Turkey.
10. Most of the paraautochthonous units of the Brunovistulian Devonian and Carboniferous cover correlate with the paraautochthonous units of the Rhenish Massif. Correlation of the allochthonous Andelska Hora, Horni Benesov and Protivanov nappes with the Rhenohercynian allochthon of Germany is proposed (Fig. 12).

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