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Development of the palaeogeography of Pangaea from Late Carboniferous to Early Permian

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Abstract

Old and new, well and poorly known lithofacies data of Moscovian and Artinskian age in the area limited by the Arctic Sea and central Africa, the Atlantic Ocean and the Ural Mountains have been plotted on present-day maps. The facies-mapped and related environments allow the recognition of a series of different palaeogeographic/palaeotectonic units including the Late Carboniferous to Early Permian deep sea to Palaeo-Tethyan ocean, west Siberia–Kazakhstan continent and Uralian–Kazakhstan–Tienshan arc, Uralian foredeep up to the closing arms of the Uralian ocean, Precaspian basin, Russian platform, Donets rift basin, Caucasian–Moesian–Dobrogean–Polish–Oslo basinal belt, syn-tectonic Hercynian foreland basin, intramontane post-Hercynian basins, Iran–Anatolian–Hellenic–Dinaric–Carnic basin branching into the Hungarian seaway, Apennine basin branching into the Cantabrian and South Portuguese basins, Oman–Iraq–Levantine–Sicily deep basin and inferred oceanic sea-way, south Peri-Tethyan platform basins, east Arabian cratonic basin, and north African intracratonic basins. The subsidence/sedimentation trends of these units were correlated and compared through lithostratigraphic logs, bathymetric curves and uncorrected cumulative stratigraphic curves. The sets of original and processed data were used to test two different palaeodynamic models, a Pangaea A model, static from Late Carboniferous to Triassic, and a mobile Pangaea B model with different dextral displacements between Laurussia and Gondwanaland in the same time interval. The best fit for our data requires a strike-slip offset of about 800 km from Moscovian to Artinskian time. This model implies a *first quasi-Pangaea* or *Pangaea B* assembly at the Carboniferous/Permian transition, an *ephemeral Pangaea B break-up* driven by an Early Permian oblique rift across the Mediterranean to Caribbean areas, and a final *Pangaea A* assembly in the Mid-to-Late Permian. The two palinspastic maps describing the model have been cross-checked by comparison with an independent set of biogeographic features of Late Carboniferous and Early Permian. Overall floral, reptile, and marine benthic organism distribution is consistent with the Early Permian trans-Pangaea seaway inferred from facies and palaeodynamic analyses.

Keywords: Moscovian; Artinskian; palaeogeography; palaeobiogeography; Pangaea B; trans-Pangaea seaway

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1. Introduction

The traditional view of Carboniferous through Triassic history was that of a consolidated stable Hercynian Europe gradually peneplanated and eventually remobilised and fragmented at its southeastern margin. Bullard's Pangaea fit (Van der Voo, 1973) for the early Jurassic Atlantic Pangaea break-up after the assembly of Pangaea had been applied back to the Mid-to-Late Carboniferous. Neither compelling data on dextral post-Hercynian transcontinental shear zone (Arthaud and Matte, 1977; Burg et al., 1994) nor innovative palaeomagnetic models, such as Irving's Pangaea fit (Irving, 1982; Morel and Irving, 1981) were able to graze the general acceptance of the static Bullard's model in palinspastic map reconstructions (Briden et al., 1974; Hallam, 1983; Scotese, 1984, 1994; Golonka et al., 1994; Scotese and Langford, 1995; Ziegler et al., 1997; Ziegler, 1988, 1989). Only a few authors have considered the adoption of Irving's model in some different time/space versions to best fit their geological data (Muttoni et al., 1996; Rau and Tongiorgi, 1981; Ricou, 1996; Stampfli, 1996; Stampfli et al., 2001; Vai, 1994, 1997).

Curiously, important field geological information available since the 19th century was basically neglected, although major palaeogeographic and palaeobiogeographic problems were raised (Gemmellaro, 1887; Glenister and Furnish, 1961; Gobbett, 1973; Kahler, 1974; Skinner and Wilde, 1966; Stevens, 1984). This information, which is inconsistent with the traditional view referred to above, has been updated in the last decade (Catalano et al., 1991; Di Stefano and Gullo, 1997; Kozur, 1995; Kozur et al., 1996). All this information has been carefully reconsidered when compiling the Moscovian and Artinskian maps for the Atlas Peri-Tethys (Dercourt et al., 2000). Many new additional data have been derived from grey literature on the poorly known late Palaeozoic rocks of the circum-Mediterranean to Middle East area (IGCP Projects 5 and 276). Objectives of this paper are (1) to highlight the features of the first two maps of the Atlas Peri-Tethys, (2) to back up a palinspastic model widened to a hemiglobal area east and west of the map limits, and (3) to discuss this model with alternative ones.

2. Methods and constraints

No Carboniferous to Early Permian maps were compiled in the Atlas Tethys by Dercourt et al. (1993). So the first Artinskian and Moscovian maps are plotted here on a present-day grid using available and many new data partly unpublished and partly collected through IGCP Projects 5 and 276 (Sassi and Zanferrari, 1990). Figs. 1 and 2 show the distribution of validation points used for constraining the mapped units. Many points represent the average of spot sections measured or drilled in the surrounding area. Because of the uneven distribution of the points, the palinspastic colour maps (at 10 million scale) are largely inferential (Vai, in Dercourt et al., 2000). Major open questions, lack of data, and speculations are shown by blank areas in the 10 million maps. The reader may refer to these maps for location of regional data not shown in Figs. 1, 2, 6 and 7).

Different palaeomagnetic solutions to be adopted for the palinspastic base maps have been tested to obtain the best fit with the geologic meaning of the facies and environmental belts reconstructed on the present grid. A static Bullard-type and two mobile Irving-type solutions have been considered. One of the latter demanded about 2000 km of right-lateral longitudinal shear movement between Gondwanaland and Laurussia. In the other, an offset of about 800 km was assumed, which is a figure double that adopted by Ricou for the Murgabian map (Late Permian) in the Atlas Tethys (Dercourt et al., 1993). This second mobile solution provided the best fit to the data available. If the entire movement occurred during the Permian, the average rate of displacement would have been 1.6 cm/yr. Of course, the commonly accepted palaeolatitudinal position of the cratons, although conforming to the kinematic model adopted, has been maintained. Displacements less than 1°, largely within the resolution of the measured palaeopoles, have been allowed. As known, the confidence level of the palaeomagnetic method is about 5° (or about 500 km on a

Fig. 1. Location of major palaeogeographic/palaeotectonic units distinguished plus validation points used for the Moscovian map on a present-day grid. Land areas (blank), continental basins (dotted), brackish to shallow marine basins (light shaded), deep marine to oceanic basins (medium shaded), volcanics (dark shaded), orogenic fronts (thicker lines), possible ephemeral marine connections (arrow).

great circle). Further details on constraints and time slice definition and resolution are found in Vai and Izart (2000a,b).

3. Palaeogeographic/palaeotectonic units

Several palaeogeographic/palaeotectonic units can be distinguished on the Moscovian and Artinskian validation point maps on a present-day grid (Figs. 1 and 2). Each of them is characterised by a relatively homogeneous stratigraphic succession, suggesting a common palaeogeographic domain and a specific tectofacies/palaeotectonic setting. Some of them included in the peri-Tethyan margins have not been deformed since the time of deposition/emplacement, or following their Hercynian consolidation. Others, included in the Palaeo-Tethyan to Tethyan areas, have been involved in the Cymmerian and Alpine orogenic cycles.

Units distinguished on the maps are listed anticlockwise and roughly characterised. Additional information and sources are available in Vai and Izart (2000a,b).

3.1. Late Carboniferous to Early Permian deep sea to Palaeo-Tethyan ocean

A deep sea environment was extrapolated into the Moscovian map from Early Permian Oman radiolarites. It is also consistent with the scattered thin-bedded siliciclastic turbidites, mudstones and radiolarian chert reported from the Aral Sea, Turan, Pamir–Tienshan area. The deep sea belt continued both westward (see units below) and northward (Uralian ocean). Evidence of northwestward subduction of the inferred Palaeo-Tethyan oceanic lithosphere is found in the large andesitic belt developed in both maps at the southern margin of the Kazakhstan continent and possibly in the Precaucasian Hercynian orogenic belt.

The Palaeo-Tethyan ocean was a large western lobe of Panthalassa left over after the Hercynian suturing of Rheic and Uralian oceans, and partly confined to the east by the Far East blocks (Metcalfe, 1999; Metcalfe et al., 1999; Robertson and Dixon, 1984; Robertson et al., 1996; Sengör, 1984, 1985; Stampfli and Pillevuit, 1993; Stampfli et al., 2001; Yan and Yin, 2000; Yin, 1997).

3.2. West Siberia–Kazakhstan continent and Uralian–Kazakhstan–Tienshan arc

Continental basins are characterised by the Angaraland-type flora which is sharply different from floras found elsewhere. Huge 2–4-km-thick andesites cover the large Balkash Lake area (Seitmuratova et al., 1997) in the core of the volcanic arc.

3.3. Uralian foredeep to closing arms of the Uralian ocean

The Uralian foredeep developed west of the volcanic chain by northward propagation of the oblique collisional belt and by subsidence triggered from the load of westward migrating nappes (Fokin et al., 2001). Evidence of previous eastward subduction of oceanic lithosphere is found in the long belt of thick andesitic bodies punctuating the Urals orogen. Westward migration of the turbidites depocentre mirrors the flexural rollback of the East European platform margin. The foredeep merged southward with the Precaspian basin.

3.4. Precaspian basin

It originated as an over 700-km-wide cratonic depression after the post-Baikalian early Palaeozoic extensional plate reorganisation. The infilling sequence shows subsidence pulses (rifts) of increasing importance through the Palaeozoic, including a Late Carboniferous to Permian interval.

The deep marine part of the basin was open to the southern end of the Uralian foredeep. The basin shows a deepening in the Early Permian, and an increasing sedimentation rate with time. This is consistent with the closure of the Uralian ocean and the channeling of detrital supply from the Uralian high chain to the Precaspian basin through the Uralian foredeep.

The basin occupies a special position at the southern edge of the Uralian orogen and at the eastern end of the north Hercynian front. A major tear-off in the sublithospheric mantle has to be

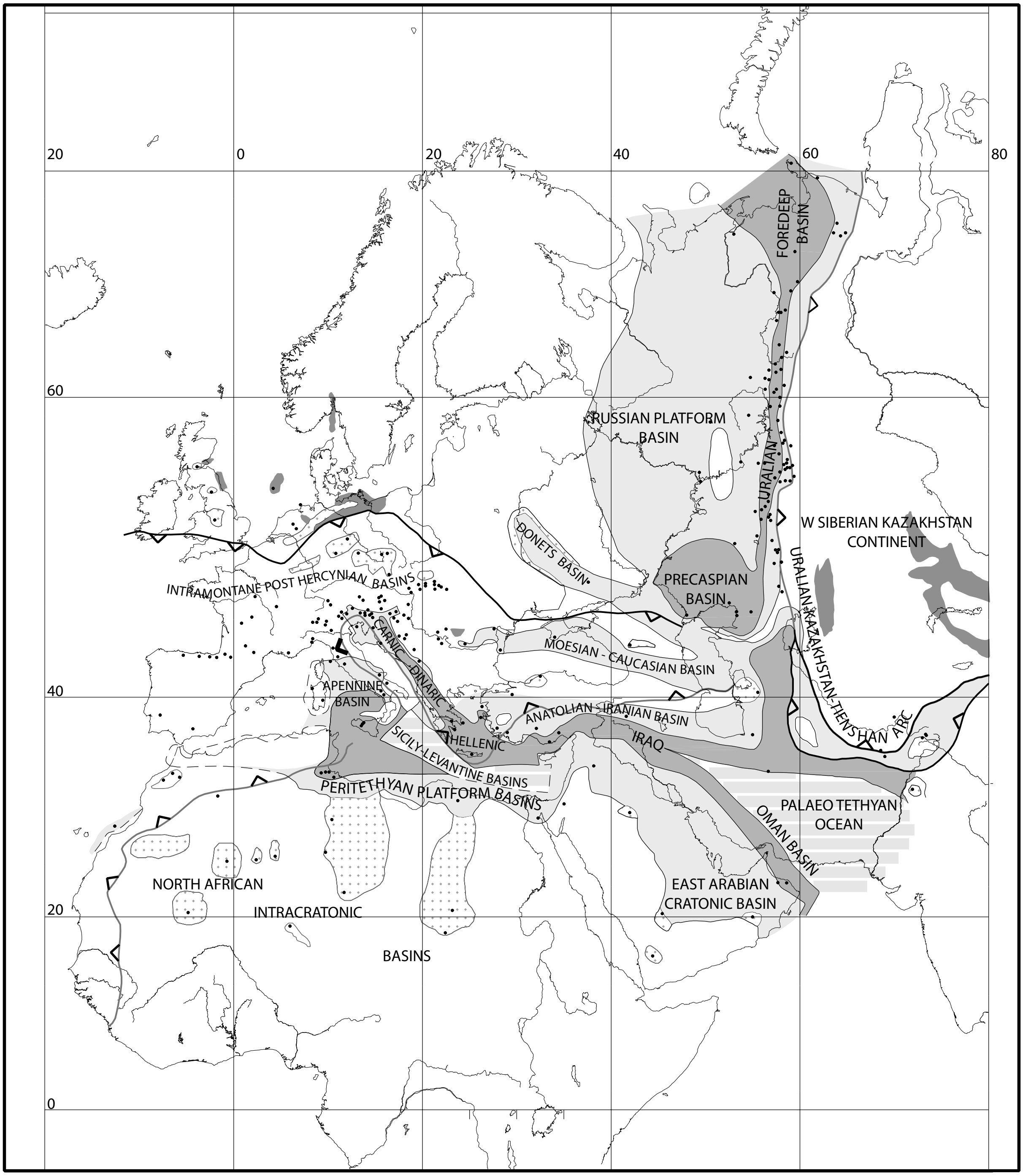

Fig. 2. Location of major palaeogeographic/palaeotectonic units distinguished plus validation points used for the Artinskian map on a present-day grid. Land areas (blank), continental basins (dotted), brackish to shallow marine basins (light shaded), deep marine to oceanic basins (medium shaded), volcanics (dark shaded), orogenic fronts (thicker lines), possible ephemeral marine connections (arrow).

assumed at the crossing of the eastward subducting European plate and the northward subducting Palaeo-Tethyan plate. It may have played a major role in the subsidence history of the basin.

3.5. Russian platform

This very large area acted as a stable to slightly extensional platform with dominantly marine sediments thinning westward.

3.6. Donets rift basin

This is a reactivated survivor of the Middle to Late Devonian Pripyat–Dnieper–Donets–Donbas–Karpinsky aulacogen (rift) during Carboniferous and Permian. In spite of being physically connected with the Precaspian basin its evolution was rather different and independent. Compressional inversions reported as Hercynian in Stovba and Stephenson (1999) occurred only during the Cimmerian (latest Triassic–earliest Jurassic) and Alpine (latest Cretaceous–earliest Tertiary) cycles (R. Stephenson, personal communication; Stephenson et al., 2001).

3.7. Caucasian–Moesian–Dobrogean–Polish–Oslo basinal belt

This narrow belt paralleling the SSW margin of the East European platform was cyclically inundated by marine water during the Permo–Carboniferous glaciation. Marine conditions persisted only in the eastern part during Permian (Yanev, 2000). The narrow seaway was bounded on the south by the Zonguldak continental basin having strong Euramerian floral affinity (Kerey et al., 1985).

3.8. Syn-tectonic Hercynian foreland basin

This basin occupied a wide elongated foredeep to foreland flexural depression at the northern front of the Hercynian chain. Cyclic marine bands punctuated the continental sequence as late as Westphalian C, suggesting connection westward to the closing Rheic ocean and/or eastward to the Moesian–Caucasian seaway.

3.9. Intramontane post-Hercynian basins

Small, thick, continental, tectonic collapse-related to pull-apart basins punctuated the inner zones of Hercynian Europe during Late Carboniferous (up to 6 km thick) and the outer zones during Early Permian (up to 2 km thick). The extensional collapse migrated in the same time from inner to outer zones. It was replaced gradually by transtensional fragmentation with discrete pull-apart basin development during the Early Permian rift mirrored by bimodal volcanism. The decreasing thickness of basin fill with time suggests a smoothing of the Hercynian chain relief.

3.10. Iran–Anatolian–Hellenic–Dinaric–Carnic basin branching into the Hungarian seaway

It consisted of carbonate platforms increasingly supplied with siliciclastic detritus westward approaching the Hercynian orogen. Floral and faunal affinity with the Russian platform and the Tethyan regions is very strong (Vai, 1994, 1997). Thus, the Hungarian seaway is viewed as a branch connecting this basin directly to the Russian platform sea. During the Early Permian rift, the Hungarian seaway was inverted.

3.11. Apennine basin branching into the Cantabrian and South Portuguese basins (possibly extending into the Amazon basin)

These three basins presented both carbonate platform and deep marine deposits during Late Carboniferous. The three basins occupied the position of residual Hercynian foredeep reaching a quiescent stage at the end of the Carboniferous. The Apennine basin was sharply separated from the Carnic basin through the Adriatic peninsula, which was morphologically welded with Hercynian Europe and also represented a geological appendix (AP) of the African platform, having been consolidated during Panafrican times (Vai, 1991). Close faunal relations to the South American Amazon basin characterise the three basins. It is worth stressing the poor faunal affinity with the Carnic Alps, which at present are so close.

Marine connections with the Amazon basin are suggested either through northwest Africa to the Florida offshore or (less easy) through central-west Africa (via Djado and/or Erdi basins).

During the earliest Permian, the Cantabrian and South Portuguese basins were uplifted and eroded. Shelf conditions instead persisted at the margin of the Apennine basin from Tuscany to Gargano, Lucania and Sosio, whereas part of Sicily underwent rapid subsidence down to abyssal plain deposits in the Sicani basin (Catalano et al., 1991; Kozur, 1995; Di Stefano and Gullo, 1997). At that time, the Adriatic peninsula was still welded to Hercynian Europe and no longer united with the African platform based on continental vs marine facies distribution. It would be better called a European morphologic promontory, from which the Mesozoic Adria microplate would then originate.

3.12. Oman–Iraq–Levantine–Sicily deep basin and inferred oceanic seaway

Scattered but clear evidence of this basin is found during Early to Middle Permian in Oman (Hawasina nappes), Kurdistan, Greek islands, Sicily and Tunisia, all these areas being involved in a rift followed by pre-Triassic compressional shear and deformation. The Oman ophiolite suggests northward transition to oceanic crust (Blendinger et al., 1990). It was speculated that the oceanic crust buried beneath the thick sediment cover of the Ionian and Levantine seas partly emplaced first during the Early Permian rift. Similarly, a narrow shallow seaway is thought to have separated Gondwanaland from Laurussia during the same Early Permian rift (Vai, 1994, 1997; see also discussion below).

3.13. South Peri-Tethyan platform basins

These basins form a cluster of anastomosing basins developed upon the rim of the north African platform. They represent the Panafrican foreland of the Hercynian orogen. Limnic to paralic basins were confined to Morocco. Carbonate platform conditions passing to siliciclastic shelf landward covered the large remaining area during Late Carboniferous. Extension to transtension-related subsidence dominated this belt. Regression and major northward displacement of the coastline occurred during Early Permian, accompanied by drowning of carbonate platforms, rapid deepening and increasing subsidence at the Sicily–Levantine–Tethyan edge.

3.14. East Arabian cratonic basin

During Late Carboniferous, most of the Arabian peninsula was a site of thin continental deposition. The Arabian platform was stable or slightly uplifting except for its northwestern edge (Sinai and Palmyra rifts) (Guiraud et al., 2001; Sawaf et al., 2001). During Early Permian, marine shelf deposits extended over one half of the peninsula, basinal facies being limited to Oman. This regional transgression reflects the Early Permian Tethys spreading with passive continental margin extension and subsidence.

3.15. North African intracratonic basins

Two types of basins punctuate the northern part of the African plate: (a) subcircular less than 1000-km-wide basins developed on the west African Proterozoic craton in limnic to alluvial conditions from Late Carboniferous to Early Permian; (b) N–S elongated over 1000-km-wide fault-controlled depressions developed near the margins of the west African and central African Proterozoic cratons or inside the intervening Panafrican belt; in these depressions the Late Carboniferous is largely marine and the Early Permian is continental, except for the Egyptian, Libyan and south Tunisian coastal area.

4. Correlation and subsidence/sedimentation trends

In spite of the problems still existing in definition of chronostratigraphic units within the considered time span, the level of chronologic correlation among the successions of the different palaeogeographic units (areas) listed in Section 3 has considerably improved. This was possible through the efforts of the Peri-Tethys and other

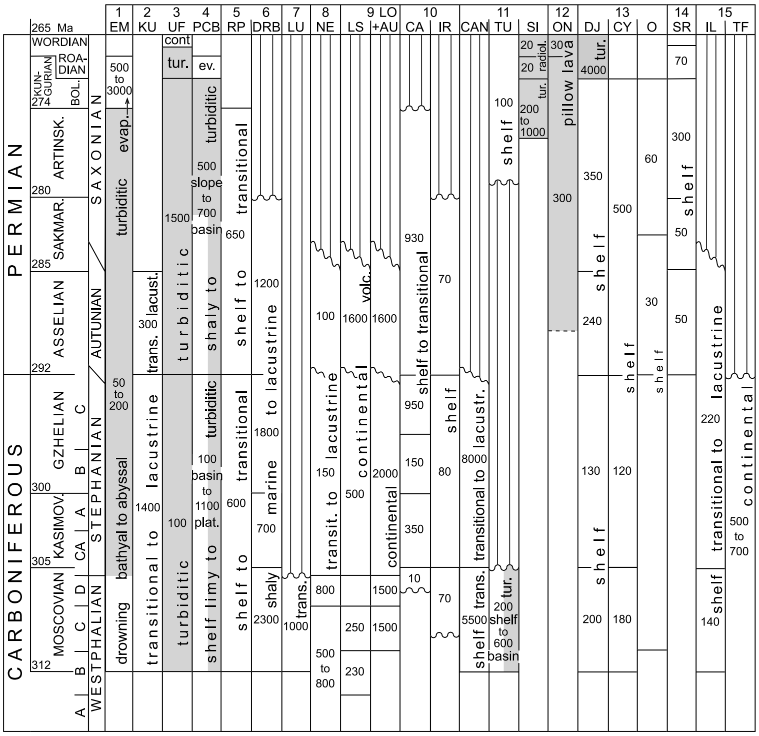

Fig. 3. Thickness (m) and environmental assignment (deep marine environments are shaded) of deposits in selected basins representative of major palaeogeographic/palaeotectonic units distinguished (see text). CA, Carnic Alps; CAN, Cantabria; CY, Cyrenaica; DJ, Djeffara; DRB, Donets rift basin; EM, Emba; IL, Illizi; IR, Iran; KU, Kuznetsk; LO, Lorraine+Autun; LS, Lower Silesian; LU, Lublin; NE, Netherlands; O, Oman; ON, Oman nappes; PCB, Precaspian basin; RP, Russian platform; SI, Sicily; SR, Salt Range; TF, Tindouf; TU, Tuscany; UF, Uralian foredeep. Time scale basically adopted from Menning (1995) and Menning et al. (2000).

international programmes (Pangea, IGCP Projects 343, 359, and 369, etc.), as well as the activity of pertinent bodies within the International Commission on Stratigraphy (Glenister et al., 1999; Izart et al., 1998; Jin et al., 1997; Wardlaw et al., 1999).

However, the degree of chronologic resolution and knowledge of sedimentation history of the individual basins and areas is quite different from place to place. So, for the purpose of this paper, the modal successions of the palaeogeographic/palaeotectonic units distinguished (Figs. 1 and 2) were compared simply by using their lithostratigraphic logs (Fig. 3), bathymetric curves (Fig. 4) and uncorrected cumulative stratigraphic curves (present thickness of stratigraphic units vs time) (Fig. 5) in order to obtain characteristic trends. No attempt to apply backstripping techniques and corrections (Watts and Ryan, 1976; Steckler and Watts, 1978) was made, because reliable data were available for only a few of the areas considered (e.g. Izart and Vachard, 1994; Brunet and Cloetingh, 2003). The purpose of this paper, in fact, is more to compare sedimentation/subsidence trends rather than quantify the corresponding rates. Nevertheless, after some cross-checking, it was felt that corrected subsidence curves would not differ significantly from those compiled in Fig. 4. For a correct reading of the stratigraphic curves (Fig. 5), a corresponding diagram of bathymetric curves for the selected units (and basins) was inferred from faunal/floral assemblages, sedimentary structures and facies recognised (Fig. 4). Furthermore, long-term eustatic sea level changes (after Ross and Ross, 1987, 1988, 1995) were reported in Fig. 5. In

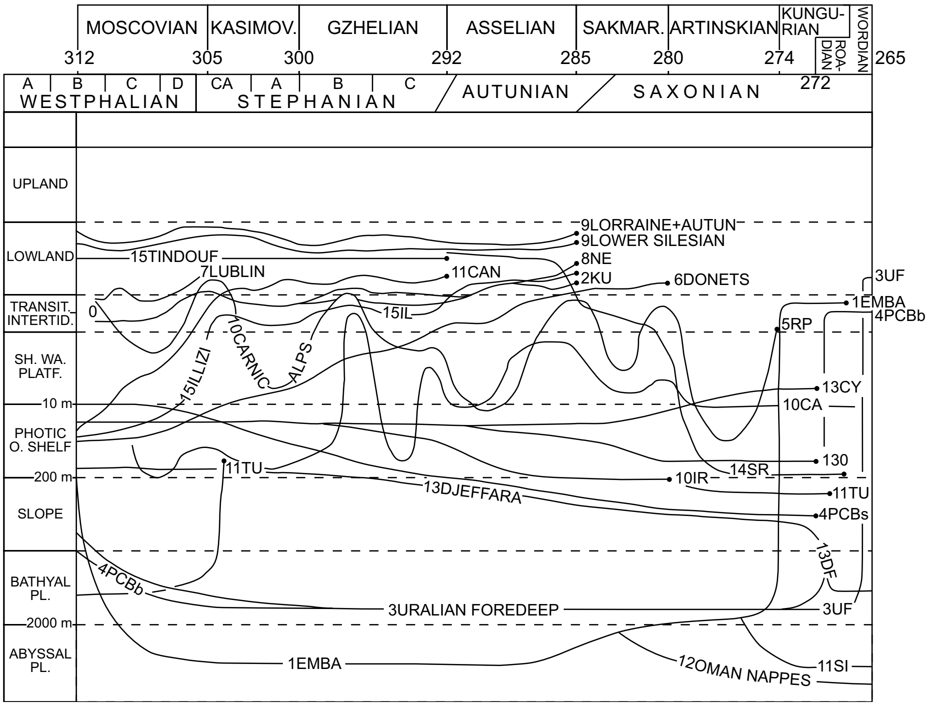

Fig. 4. Bathymetric curves of selected basins inferred from environmental reconstructions. See Fig. 3 for legend.

this way, subsidence related to sediment loading can be qualitatively separated from thermo-tectonic subsidence.

A careful comparative evaluation of data compiled in Figs. 3–5 gives the following results.

Synthetic lithostratigraphic logs (Fig. 3) show that, out of selected basins, deep water (bathyal to abyssal) deposits occur in the Latest Carboniferous to Early Permian of the Emba, Uralian foredeep and Precaspian basins, and in the Early to Middle Permian of the Sicily, Oman (nappes) and Djeffara basins.

Bathymetric curves (Fig. 4) indicate three basic types of evolution.

(1) Basins with almost constant bathymetry, modulated by eustatic-climatic sea level oscillations (intermediate frequency). Two groups can be distinguished: (a) continental lowland to transitional basins with relatively smaller-amplitude modulations; the group comprises the Lorraine+Autun, lower Silesian, Lublin, Netherlands, Kuznetsk and Tindouf basins, which correspond to intramontane post-Hercynian basins of the northern Peri-Tethys plus intracratonic basins of the southern one; (b) transitional to shallow marine basins with relatively higher-amplitude modulations; examples are the Donets rift, Russian platform, Cantabria, Carnic Alps, Illizi, and Salt Range basins, which correspond to different types of basins (aulacogen, foreland, foredeep) on both northern and southern Peri-Tethys margins.

(2) Weakly to strongly deepening basins. (a) The margins of many basins have bathymetry ranging from shelf to slope, as the Tuscan, Iran, and Precaspian basins, probably located not far from oceans. (b) The centre of the basins has bathymetry ranging from shelf to bathyal–abyssal plain, as Djeffara, Precaspian, Uralian foredeep, Sicily and Oman nappe basins, which were located at or in connection with Permo–Carboniferous oceans.

(3) Basins with long-lasting deepening followed by rapid shallowing because of infilling, as the Emba, Precaspian, and Uralian foredeep basins.

Fig. 4 also shows two groups of basins. One comprises basins exhausted before the Artinskian,

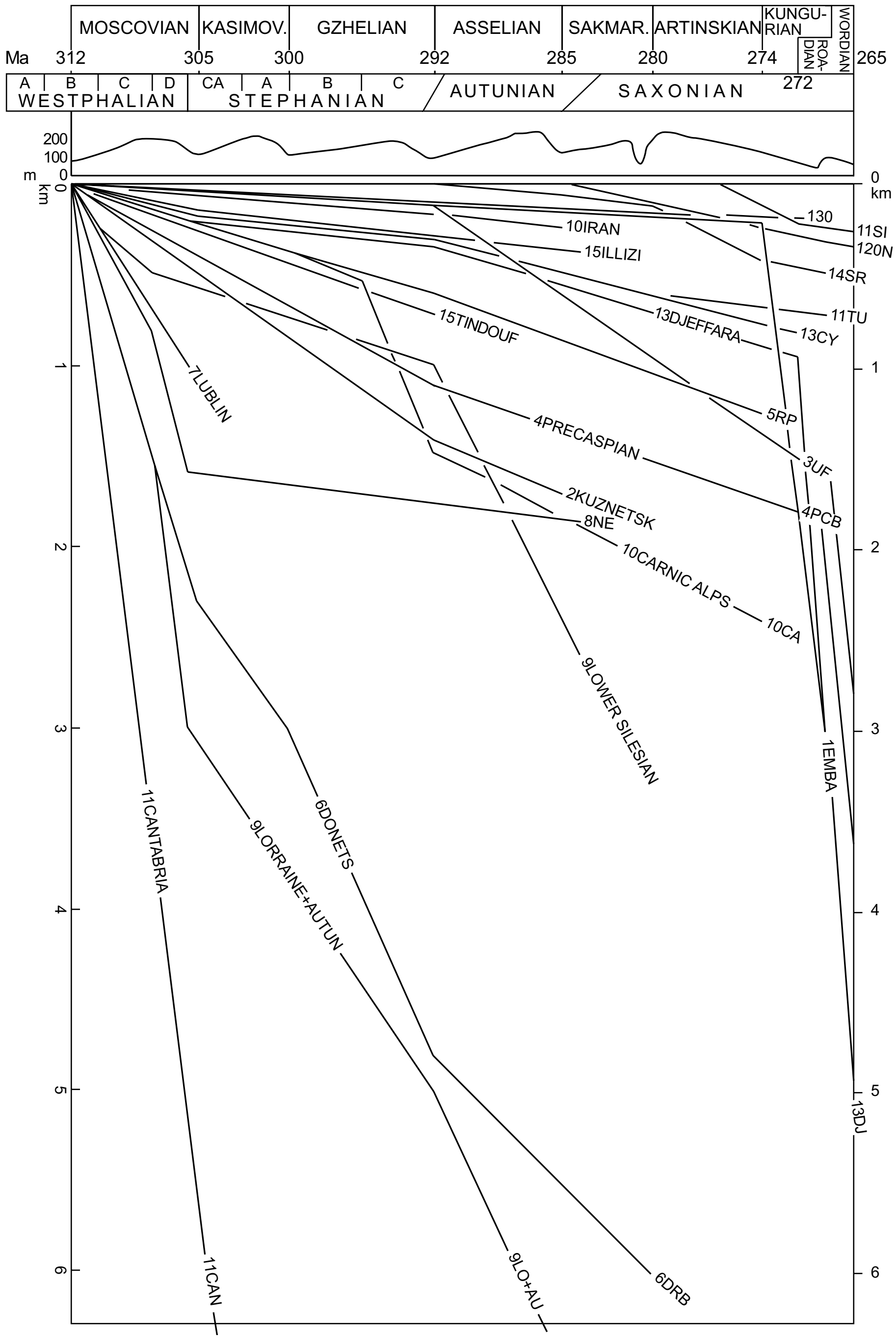

Fig. 5. Uncorrected cumulative sedimentation curves of selected basins. Long-term eustatic sea level changes after Ross and Ross (1987, 1995). After addition of water depth, uncorrected total subsidence would be obtained. See Fig. 3 for legend.

the other those continuing sedimentation after the Artinskian. The first is located apart from and the second near the Peri-Tethys margins. A third group of basins originated only during the Permian or was reset since.

Also the cumulative stratigraphic curves (Fig. 5) are very interesting, especially when evaluated concurrently with bathymetric information (Fig. 4). Different types of basins can be distinguished.

(1) Highly supplied (stuffed) highly subsiding basins, as the Cantabrian, Lorraine+Autun, Lublin, Netherlands and Donets basins, including foredeep, intramontane and rift basins. Tectonothermal and sediment load subsidence were active processes there. The basins became inactive at different times during the Late Carboniferous or Early Permian.

(2) Poorly supplied (starved) highly subsiding basins, as the Sicily and Oman nappe basins, where most of the subsidence was accommodated by the increasing water depth. The dominant process there was tectonic subsidence during Early Permian. A variant in this group is represented by the Emba basin showing a jump in sedimentation rate (from about 5 to 1400 m/Myr) after the Artinskian, resulting in a rapid infilling of the basin. This is a signal of twofold inversion: reduced subsidence and increased detrital supply. Notice that the same process occurred shortly afterwards in the Uralian foredeep and Precaspian basins.

(3) Medium-supplied medium-subsiding basins, as the Iran, Salt Range, Cyrenaica, Tuscany, Illizi, Tindouf, Russian platform, Uralian foredeep, Precaspian, Kuznetsk and Carnic Alps basins. Subsidence was controlled by tectonic, sediment-loading and tectonic-loading processes according to the different types of basins clustered in the group. A variant in this group is represented by the Djeffara basin which, like the Emba and other basins, shows a jump in sedimentation rate (from about 20 to 4000 m/Myr) after the Artinskian, leading to rapid infilling of the basin by reduced subsidence and increased detrital supply. This extremely rapid turbidite infilling provides evidence of water depth over 3 km in the Djeffara basin, and, by analogy, in the continuing starved Sicily basin close nearby.

An additional important general remark derives from the diagram (Fig. 5). Two major jumps (normally positive) in sedimentation rate are indicated by many curves near the base of the Permian, and by the Emba and Djeffara curves shortly after the Artinskian. The two events, referred to as the Early Permian rift and Middle Permian inversion respectively in Section 5, represent major perturbations in the sedimentary regime and the controlling stress pattern of western Tethyan and Peri-Tethyan areas.

A further remark is concerned with the imbalance of the northern vs southern Peri-Tethyan margins. Stuffed highly subsiding basins (Iberia, France, Germany) are dominant or exclusive in the northern margin. As an explanation, the northern margin is largely part of the Hercynian orogen, whereas the southern margin is mainly out.

5. A model for late Palaeozoic dynamic evolution of the Laurussia/Gondwanaland hinge area

The set of data presented (Figs. 1–5) requires a Carboniferous to Permian evolutionary model quite different from those currently available, to reach a satisfactory fit such as that simplified in Figs. 6 and 7 or detailed on maps 1 and 2 in the Atlas Peri-Tethys (Dercourt et al., 2000).

The classical concept of Pangaea dates back to Snider-Pelligrini (1858), Sacco (1906) and Wegener (1912, 1915). The majority of Permo–Carboniferous palinspastic restorations issued in the last decades are characterised by adoption of this static Pangaea A-type configuration undifferentiated from that assumed for the immediately pre-Jurassic break-up fit (Bullard et al., 1965). This means that the northwestern Gondwanaland margin was already welded to Laurussia and stable since the Late Carboniferous (Bosellini and Hsü, 1973; Biju-Duval et al., 1976; Golonka et al., 1994; Hallam, 1983; Lawver and Scotese, 1987; Lottes and Rowley, 1990; Ross and Ross, 1988; Scotese, 1984, 1994 (in Klein and Beauchamp, 1994); Scotese and McKerrow, 1990; Scotese and Langford, 1995; Sengör et al., 1988; Smith and Briden, 1977; Ziegler et al., 1979, 1997; Ziegler, 1988, 1989; Zonenshain et al., 1987, 1990).

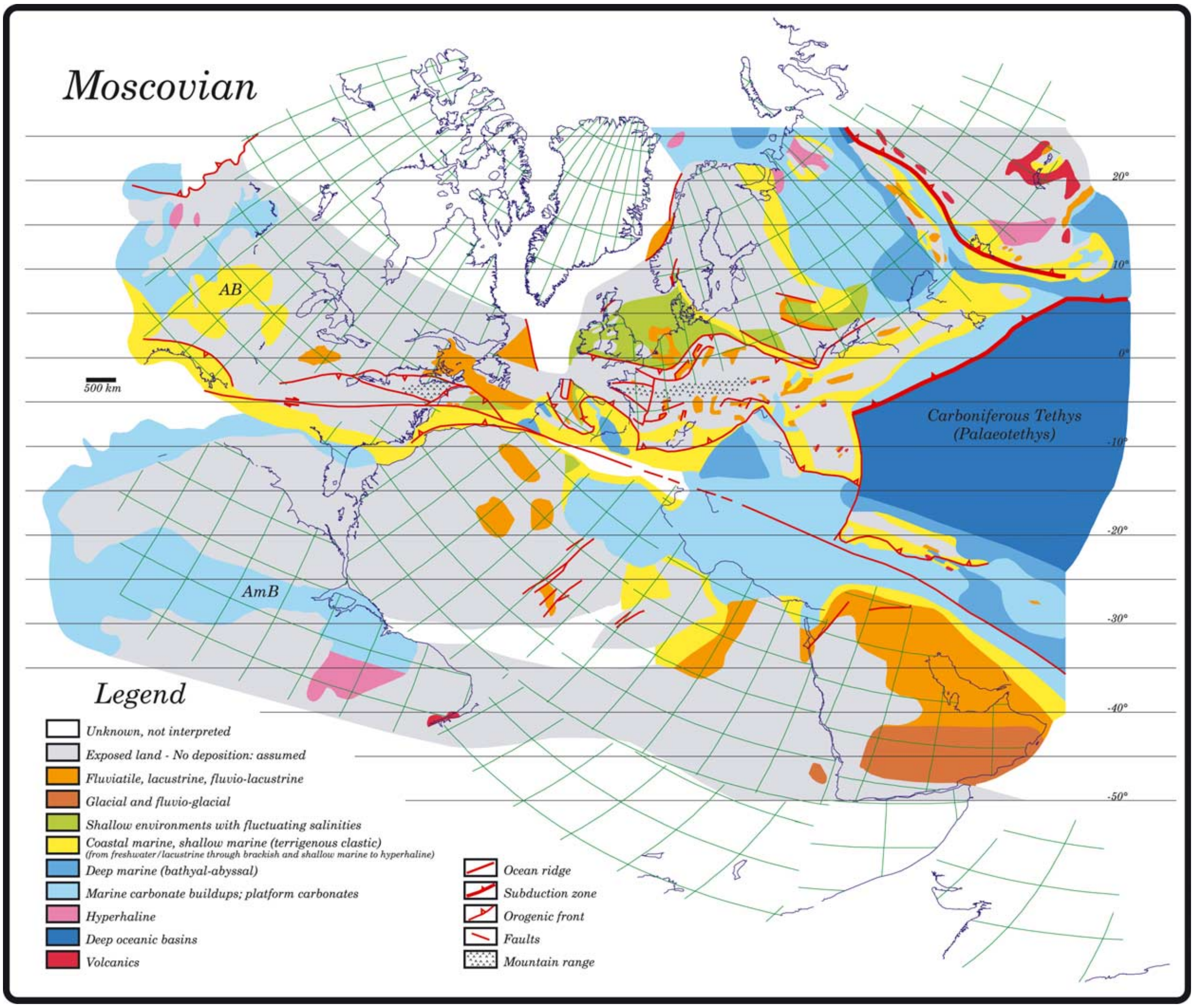

Fig. 6. Moscovian map (modified and expanded after Dercourt et al., 2000). AB, Anadarko basin; AmB, Amazon basin.

A few authors, however, have suggested a less static Pangaea configuration implying a certain degree of strike-slip displacement between Laurussia and Gondwanaland in different places (Lefort and Van der Voo, 1981; Muttoni et al., 1996; Rau and Tongiorgi, 1981; Ricou, 1996; Robertson and Dixon, 1984; Smith and Woodcock, 1982; Stampfli and Pillevuit, 1993; Stampfli, 1996; Stampfli and Mosar, 1999; Stampfli et al., 2001; Vai, 1994, 1997, 1998).

A simple model able to explain most of the old and new stratigraphic and structural data of the former Hercynian orogen and the Gondwanaland and Laurentia supercontinents facing the orogen was proposed by Vai (1994, 1997, 1998). The model assumes a mobile Pangaea B setting evolving gradually to Pangaea A (Morel and Irving, 1981; Van der Voo, 1973).

A nearly continuous dextral strike-slip displacement between Laurussia and Gondwanaland from Carboniferous to Triassic provides the basic kinematic frame. A stepwise evolution can be summarised as follows (Vai, 1998, fig. 16).

(1) Mid to Late Carboniferous oblique (transpressive) convergence between south Laurussia and the northern Gondwanaland margins gener-

ates the large European (Variscan) and Mediterranean segment of the Hercynian orogen. Meanwhile, the frontal (compressive) convergence between east Laurussia and west Siberia margins as well as the northwestern Gondwanaland and southeastern Laurussia margins results in the first linear Uralian and Mauretanid segments of the Hercynian orogen (Vai, 1991). The ensuing orogenic uplift leads to the following stage.

(2) At the Carboniferous to Permian transition a first Pangaea assembly is reached. I have called it *quasi-Pangaea* or *Pangaea B* (northwestern margin of South America facing the southeastern margin of North America). The Siberia continent is still separated from the supercontinent by the closing Uralian ocean.

(3) However, a steady state is not yet achieved and the two components of the system again start moving apart by dextral divergent (transtensional) megashear, involving at the same time Europe, North America and the Mediterranean area. The peak of this shear pulse occurs with the Early Permian rift. The rift can be genetically related to the opening of the Permian or Permo–Triassic Tethys branch and the faster convergence of the Palaeo-Tethys ocean (Fig. 7).

(4) The Early Permian oblique rift produces an *ephemeral Pangaea B break-up* aborted after a short time (about 20 Myr). The dextral transtensional displacement results in bringing northwestern Africa in front of northeastern America.

(5) The mid-Permian post-Artinskian change from transtensional to transpressional stress regime leads to Alleghenian collision of northwestern Africa with northeastern America. At the same time any residual oceanic area becomes closed in the Urals, and the *full Pangaea A1* stage is reached. An explanation for the mid-Permian changing stress regime may be the activation and a first pulse of the intra-Gondwanan Somaliland rift (see below).

(6) During Late Permian to Early–Middle Triassic, a 20° clockwise rotation in the future Gulf of Mexico area occurs and the full Pangaea A2 stage is accomplished.

(7) The dextral strike-slip regime possibly begins its gradual inversion with the Mid–Late Triassic Carnian to Norian rift (Courel et al., 2000). The Mid-Jurassic Atlantic and Tethys opening is driven by a major sinistral strike-slip rift (Kent et al., 1995).

In this overall picture the key role in changing the palaeostress regime is played by the activation of more or less documented new oceanic spreading centres. This model is in good agreement with the increasing number of regional and general palaeomagnetic studies favouring the Pangaea B approach in recent years (Muttoni et al., 1996; Ricou, 1996; Torcq et al., 1997; Besse et al., 1998).

6. The Moscovian and Artinskian maps

The following is a description with discussion and interpretation of data compiled on the two maps based on the model above (Figs. 6 and 7).

6.1. Moscovian map

The Moscovian (312–305 Ma) to Late Carboniferous map (Fig. 6) shows large epicontinental seas in the Russian platform, in northern Europe from Poland to Ireland, and in northern Africa; they range from fully marine to brackish-paralic or evaporitic conditions. Even inside the Hercynian front, Europe and the Near East areas appear to have been fragmented by a pattern of en échelon seaways open to the southeast and closed to the northwest. They are the Oslo–Polish branching into the Moesian to Caucasian, the Carnic–Dinaric continuing or branching into the Anatolian–north Iran, the Cantabrian and south Portuguese continuing or branching into the Apennine seaway; a conjugate branch of the Carnic–Dinaric seaway is represented by the trans-Hungarian seaway; trend and size of these seaways are quite comparable with the Donets rift basin still active during the Moscovian (Figs. 1 and 6) (Al Youssef and Ayed, 1992; Briand et al., 1998; Coquel et al., 1988; Demirtasli, 1990; Izart et al., 1996, 1998; Jenny et al., 1978; Kora, 1998; Kovácks et al.,, 2000; Lys, 1988; Massa and Vachard, 1979; Monod, 1977; Nedjari, 1982; Olaussen et al., 1994; Pasini and Vai, 1997; Protic et al., 2000; Ramsbottom et al., 1978; Vachard et al., 1993; Vai and Venturini,

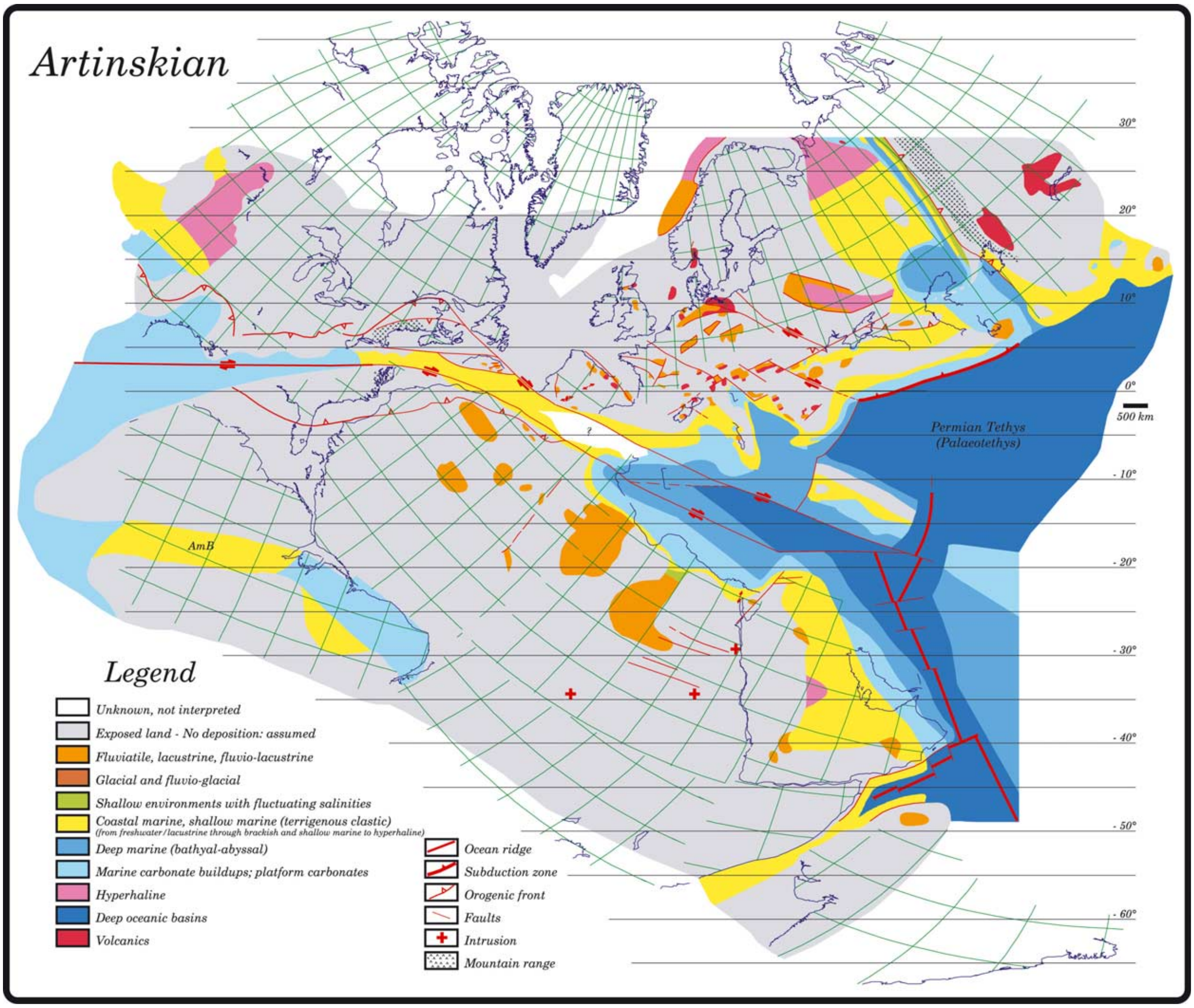

Fig. 7. Artinskian map (modified and expanded after Dercourt et al., 2000). AmB, Amazon basin.

1997; Vai and Izart, 2000a,b; Vozarova, 1998; Yanev, 2000; Zdanowsky and Zakova, 1995; Ziegler, 1988). So, at the end of the Hercynian orogeny, Europe and its northern (Caledonian) and southern (Panafrican) foreland areas were still involved in a thalassocratic regime, as a consequence of one of the global transgression maxima (Paproth, 1987).

Deep sea basins developed at the migrating western front of the Uralian chain (Uralian foredeep), at the southern front of the Kazakhstan arc (a belt north of the Turan Sea), at the crossing of the west Uralian and north Hercynian fronts (Precaspian basin), and at the south Hercynian front with the Cantabrian, south Portuguese, Apennine and Hellenic basins (Demirtasli, 1990; Ensepbaev et al., 1998; Filipović, 1995; Izart et al., 1998; Oliveira et al., 1983; Papanikolau and Sideris, 1990; Pasini and Vai, 1997; Ramovš, 1990; Villa, 1985; Wagner and Winkler-Prins, 1985). Basins floored by oceanic crust occupied the large Palaeo-Tethys or Carboniferous Tethys (Fig. 6) to the east and some branches of the northern Urals (Nikishin et al., 1996). The Palaeo-Tethys had a northward subduction margin and a southern passive margin at the northern limit of the Turkish (Kirsehir) and Iran blocks.

Isolated continental basins characterise two quite different settings: (a) small-size post-tectonic collapse basins (Donsimoni, 1981; Doubinger et al., 1995; Korsch and Schäfer, 1995; Kovács et al., 2000; Oplustil and Pesek, 1998; Vozarova, 1998); (b) large intracratonic northwest Africa (Tindouf, Taoudenni, Reggan and Iullemedden) (Conrad, 1985; Legrand-Blain, 1985; Massa, 1985) and northeastern America (Sidney and Fundy) (Pascucci et al., 2000; Ziegler, 1988) basins. Both areas show long-wave elliptic to subcircular depressions inside two major bulges.

Maxima of Moscovian subsidence and sedimentation are found in the Cantabrian, Lorraine, and Donets basins (Fig. 5). Palaeobiogeographic data show strong faunal affinity of the Tuscan Apennine and Cantabrian basins with the South American Amazon basin for corals, brachiopods and conodonts (Ferrari et al., 1977), whereas fusulinids have Russian platform affinity (Villa, 1985; Davydov, this volume). This contrasts with the stronger Russian platform affinity of the Carnic–Dinaric faunas (Ferrari et al., 1977; Krainer and Davydov, 1998; Pasini and Vai, 1997; Vai and Venturini, 1997). On the other hand, Europe, eastern North America and Moroccan Meseta have Euro-Amerian floral affinity, which is unknown in north Africa (except Morocco) and Gondwanaland.

Unlike some recent reconstructions (Golonka et al., 1994; Scotese, 1994; Scotese and Langford, 1995; Ziegler et al., 1997), physiographic evidence of a high-elevation mountain chain as a consequence of the Hercynian orogeny is found only in the Silesian–Moldanubian–Saxo-Thuringian–Central Massif and Armorican belt (Becq-Giraudon and Van den Driessche, 1994), and in the inner Appalachian belt (Fig. 6).

Convergence and/or subduction was prominent only along the west Siberian to southwest Kazakhstan margin, as shown by the impressive system of south Uralian to Balkash–Tienshan–Pamir volcanic arcs. Convergence also occurred along the Mauretanian margin. The large area in between was dominated by extension to small-scale transtension, with both land and shallow sea characterised by a tight network of small-scale pull-apart basins, and by a relatively low isostatic uplift of the Hercynian belt, except for the innermost tectonic zones. This is consistent with small volumes of deep-seated granitic intrusions and few isolated volcanic centres.

6.2. Discussion

The South American faunal affinity of the Apennine and Cantabrian basins with the Amazon basin recorded at the level of corals, brachiopods and conodonts (Ferrari et al., 1977) requires some kind of marine connection. The most likely, although far and partly speculative (see below), way would be through the northwest Africa offshore and the future Caribbean and Cordillera areas with a shorter loop along the west Africa offshore. An even more speculative alternative would be away from the Djado and/or Kufra basins through the Gulf of Guinea.

A shelf sea extended at least 1000 km west of the Djeffara basin to the Mezarif basin in northwest Africa. Its possible continuation westward and northward into the Apennine, Cantabrian and south Portuguese basins is hampered by the large Atlas Maghrebian chain. This is a south-verging thrust system with systematic décollement and sole thrusting at the Triassic evaporite horizon, thus burying tectonically the extent of its autochthonous deposits.

The relatively poor faunal affinity between the Apennine and the Carnic–Dinaric basins is explained by the barrier represented by an Adriatic peninsula physiographically connected to the Hercynian Europe (Vai, 1994, 1997). However, at the lithospheric level the Adriatic–Apulia area can be regarded as an African promontory (Channel et al., 1979) following the Panafrican crustal consolidation (Vai, 1991).

6.3. Artinskian map

Major changes and a contrasting regional trend appear in the Artinskian (280–273 Ma) to Early Permian map (Fig. 7) (Paproth, 1987). Western Europe, north Africa and even the Russian platform show the important regression taking place, with some fluctuations, up to the top of the Permian consistently with the global trend (Dickins, 1985, 1997; Ross and Ross, 1987, 1995). Even in

Europe the Permian is mostly regarded as a regressive period with the exception of the mainly continental mid-Permian Gardena (New Red) Sandstone and the evaporitic Late Permian Zechstein transgressions (Cassinis et al., 1997, 2000; Ziegler, 1988). A prominent common retreat of the coastline sometimes up to the extinction of basin is clear for the Cantabrian, Hercynian foredeep, trans-Hungarian, Donets, Russian platform, Murzuk and Kufra basins. This is even more remarkable considering that the map (Fig. 7) is drawn at one of the high sea level fluctuations (Fig. 5).

A major exception to this general trend is shown in the elongated belt at the southern front of the Hercynian orogen eastward of Tunisia and Sicily. In this belt evidence is common of progressive deepening of sediments accumulating on sialic crust stretching up to the emplacement of oceanic crust. Thin-bedded deep sea turbidites, mudstones, bathyal nodular limestones, abyssal radiolarian shales and chert, frequent large-scale olistostrome and olistoliths containing platform carbonates and seamount condensed pelagic limestone elements are found in the Sicani Mts. of Sicily (Broquet et al., 1966; Catalano et al., 1991; Di Stefano and Gullo, 1997), in Crete and many other places of the Dinaric–Hellenic–Turkish area (Baud et al., 1991; Krahl et al., 1986; Kozur et al., 1998; Papanikolau and Sideris, 1990; Ramovš, 1990; Vogl, 1913), in Kurdistan and northeastern Iraq (Vašicek and Kullmann, 1988), and in Oman (Béchennec, 1988; Blendinger, 1988; Blendinger et al., 1990). Transgression and deepening are also documented on the eastern Arabia margin (Alsharhan and Nairn, 1997; Al-Saad et al., 1991) (Fig. 7). Further east, this type of sediment is common in suture zones that separate allochthonous sialic terranes in Tibet, Yunnan, Thailand and Malaysia. These suture zones are remnants of the Palaeo-Tethys (Metcalfe, 1996; Metcalfe et al., 1999). Departures from the global Permian regressive trend (Ross and Ross, 1987, 1995) are shown also in the eastern Tethys (Chen et al., 1998).

A special feature of this elongated belt is to bear evidence of submarine synrift instability (frequent and large olistostrome/olistolith intercalations) and postrift deformation (shearing, folding, soft sediment disruption) taking place before or during early Triassic in a permanent deep marine environment.

The belt characterised by the Early Permian deepening trend can be divided into two oblique rifts.

The major rift developed at the northern margin of Gondwanaland and was identified by the Oman–Iraq–Levantine–Sicily deep basin (see above) and the Permo–Triassic Tethys. Its continuation across continental crust west of Tunisia is tectonically buried by the Atlas Maghrebian thrust system (Vai, 1997) (see below). However, transtensional effects are present north of the rift in the small post-tectonic basins of Iberia and Moroccan Meseta (El Wartiti et al., 1990), and south of the rift in the south Moroccan Tarfaya offshore (Nahim and Jabour, 1997). Pre-Triassic, possibly Permian thick synrift deposits of unknown environment are suggested from a seismic line (Fig. 8). Further evidence of this major oblique rift was provided by the Permian fusulinid limestone block found inside Cretaceous turbidites in a Cuba drill hole (A.J. Boucot and H. Kozur, personal communication).

The minor rift paralleled part of the first one to the north and was identified by the Hellenic–Dinaric–Carnic basin west of the western tip of the Permian Palaeo-Tethys. It continued with the dextral megashear across the post-Hercynian Europe (Arthaud and Matte, 1977), as shown by many small, pull-apart, discrete continental basins (Ziegler, 1988). Large bodies of epiplutonic granitoids and related bimodal volcanics were emplaced along narrow fault belts within Hercynian Europe mainly in the time interval from 300 to 275 Ma and especially from 290 to 280 Ma (Cassinis et al., 1997, 2000; Vai et al., 1984; Ziegler, 1988). At the same time, outside the north Hercynian front, large masses of alkaline volcanics were emplaced from the Oslo graben to Poland and the British Isles (Geluk, 1997; Ziegler, 1988). The evidence is clear that this magmatism is related to the Early Permian Mediterranean oblique rift (Vai, 1994) which is consistent with the dextral trans-European trans-Atlantic megashear (Arthaud and Matte, 1977), and is not related to the Hercynian orogeny (Dal Piaz, 1993).

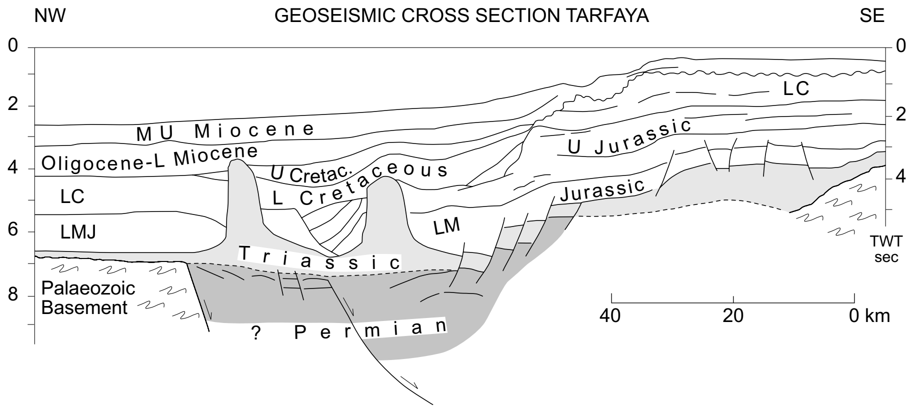

Fig. 8. Geoseismic section across the Tarfaya Moroccan offshore (modified after Nahim and Jabour, 1997).

Beginning in the Permian, the relief produced by the Hercynian orogeny was smoothed out as shown by intramontane basins increasing in number and expanding in size in comparison with the Late Carboniferous. Unlike other restorations (Golonka et al., 1994; Scotese, 1994 in Klein and Beauchamp, 1994; Ziegler et al., 1997), Hercynian Europe was eroded to form an essentially upland to lowland peneplain as suggested by the wide extent of the onlapping tabular Late Permian to Permo–Scythian New Red magnafacies (Cassinis et al., 2000). In contrast, high mountain ranges characterised the Uralian and outer Appalachian orogenic segments during the Artinskian (Fig. 7).

Convergence and/or subduction was still prominent only along the Uralian to Kazakhstan margins with the volcanic arc limited to Kazakhstan. Convergence also occurred at the Appalachian and Mauretanid–Moroccan margins (Dallmeyer, 1982; Piqué et al., 1993; Sougy, 1969).

6.4. Discussion

Key items to be discussed after comparison of the Artinskian and Moscovian maps (Figs. 6 and 7) are: occurrence of a seaway west of Sicily and Tunisia, occurrence of oceanic crust in the Ionian and Levantine seas as early as Early Permian, original pertinence of the Moroccan Meseta to Gondwanaland or Laurussia, and relation of deposits to the Late Carboniferous–Early Permian glaciation.

West of Sicily and Tunisia, the strong effects of the Early Permian rift are only apparently sharply interrupted. In fact, both E–W-trending facies limits and gentle fold axes of the Permian Djeffara marine deposits, unconformably sealed by late Cretaceous transgressive rocks, plunge and continue in the subsurface beneath the Tunisian front of the south-verging Atlas thrust system. A minimum shortening of the Atlas chain is estimated at 100 km (Frison de la Motte, personal communication, 1997). This allows for a 100-km-wide *seaway* continuing westward of the Djeffara basin. Moreover, as the Atlas thrust system detached at the Late Triassic evaporite level, there is an additional chance that the width of the present chain (200–300 km) was occupied, at least partly, by the same seaway. Direct evidence for that reasonable speculation should be sought by drilling in the Atlas subsurface or in the northwest Africa offshore, as shown for the Tarfaya segment (Fig. 8). A Tarfaya-like seismostratigraphic setting was discovered recently on deep trans-Adriatic CROP profiles, where there are thick synrift deposits correlated with the Carnic–Dinaric Permian to Permo–Carboniferous basin fills (Vai and Venturini, 1997). This is additional evidence of the effects of the Early Permian rift. Instead, the Rif and Kabylia teleallochthonous nappes, displaced from farther north, contain discontinuous continental Permian deposits suggesting a northern limit to this seaway.

The Moroccan Meseta (western and central Morocco microplate) was placed into the Laurussian southeastern margin mainly because of its Hercynian structural paradox. In conventional Late Carboniferous reconstructions of the Hercynian structural zones a sharp contrast arises from the east Moroccan internal zone being interposed between the west Moroccan low metamorphic external zone to the northwest and the non-metamorphic Anti-Atlas zone to the southeast (Piqué et al., 1998; Vai, 1980). The puzzle is solved by placing the Moroccan microplate at the Newfoundland–Nova Scotia margin.

A further problem arises from the Early Permian Mediterranean–Atlas–Caribbean (or north Gondwanan) oblique rift which cuts at a very low angle the northeastern part of the Appalachian–Mauretanid segment of the Hercynian orogen and parallels the southern Hercynian front eastward. This is not uncommon; see e.g. the Norwegian–Greenland Sea rift along part of the Caledonian axial zone during Late Carboniferous and Permian (Ziegler, 1988), and the Tyrrhenian rift that broke along part of the Alpine suture zone in late Miocene and Pliocene (Patacca et al., 1993; Vai, 1992). Extension can be active in the internal zones while compression still affects the external ones (Sartori, 1990). The case of the north Gondwanan rift is easy to explain taking into account the prominent transpressional character of the Hercynian orogen in the European–Mediterranean and also northwest Africa segments (Badham, 1982; Dallmayer and Lécoché, 1990; Piqué et al., 1993; Vai, 1980, 1991; Vai and Cocozza, 1986). In such a setting any change in the large-scale stress patterns results in rotation of displacing vectors of the two supercontinents, which accounts for change from transpression to transtension and vice versa. In this case, the onset of Permo–Triassic Tethys opening east of Arabia in the Early Permian (Fig. 7) generates the short-lived north Gondwanan oblique rift and opening. With the first activation of the west India–Somaliland rift during Late Permian, which originates the Ogaden to Malagashy basin system (Wisser, 1997; Wopfner and Casshyap, 1997), the north Gondwanan rift stopped, the northwest Africa seaway disappeared and the two facing continents welded to form the full Pangaea (Vai, 1994).

The Late Carboniferous through Early Permian glaciation affected part of Gondwanaland (Barron and Fawcet, 1995; Crowell, 1995; Crowley, 1994; Parrish, 1995; Witzke, 1990) producing important biotic and cyclostratigraphic (modulation) effects (Izart et al., 1998; Ross and Ross, 1988; Ziegler, 1990; Ziegler et al., 1997). Surprisingly enough, the late Early (end Artinskian) to Middle Permian (Wordian) deglaciation did not overcome the global regressive trend, as shown also by the global sea level curve (Ross and Ross, 1987, 1995) (Fig. 5). Similarly, the beginning of the glaciation was not registered during the Moscovian transgression. It seems that the glacio-eustatic factor was negligible compared to the geotectonic ones (such as spreading and uplift) controlling the worldwide transgressions and regressions during Late Carboniferous and Permian. Unlike Weevers and Powell (1987) and Wisser (1997), it is reasonable to assume that spreading rates were low during Late Carboniferous and especially Permian times, as shown by the transgressive conditions being limited to narrow rift belts (Fig. 7).

There is an increasing consensus to recognise the oceanic nature of the crust flooring the Ionian and the eastern Mediterranean Levantine seas (except for the Syrte basin, the East Mediterranean rise and the Eratosthenes microblock). The oceanic crust is overlain by a thick sedimentary blanket (Cantarella et al., 1997; Catalano et al., 2001; Finetti, 1982; Finetti and Del Ben, 1986; Finetti et al., 1997; Makris et al., 1986; Saioni, 1996). Also the crustal features of the Levant continental margin support this view. In fact, the nearly normal continental crust of Negev (35 km thick) is replaced northward by the thinned crust of Judea–Samaria (25 km) and further northward by the semioceanic crust of Galilee–Lebanon (Ben Avraham and Ginzburg, 1990). The commonly restored N–S-trending rift and spreading directions of the Levant basin are not supported by this N–S crustal gradient suggesting rather an E–W oblique spreading axis (Garfunkel, 1998). The age assumed for this oceanic crust (Scandone et al., 1981) is according to different authors Permian (Ben Avraham and Ginzburg, 1990; Catalano et al., 1991; Saioni, 1996; Vai, 1994), Late Permian to Early Triassic (Stampfli et al., 2001), Triassic (Finetti, 1982; Finetti and Del Ben, 1986;

Finetti et al., 1997), middle Jurassic (Cantarella et al., 1997), Cretaceous (Dercourt et al., 1986), Tertiary or even Messinian (Aubouin, 1965; Fabricius and Hieke, 1977). Lacking direct documentation, the following points are important in constraining the emplacement age of the Ionian and Levantine crust.

(a) Young Tertiary emplacements are contradicted by the cold nature of this crust (heat flow less than 1 HFU) and the unusually thick (up to 11 km) sedimentary cover (Giese and Morelli, 1975; Erickson and von Herzen, 1978).

(b) Evidence of deepening and/or accelerated subsidence of the continental margin sequences before Tertiary are known only during Late Cretaceous, Late Triassic to Early Jurassic, Middle Triassic, and especially Early to Middle Permian (Catalano et al., 1991; Charier et al., 1988; Gumati and Kanes, 1985; Vai, 1994).

(c) The many-kilometres-thick sedimentary cover overlying this oceanic crust, in relatively starved to partly supplied basins, requires an old age of the crust. Attempts to trace basinward calibrated seismic profiles from the Levant continental margin (Garfunkel, 1998, fig. 2) show that pre-Jurassic sequences are considerably thickened in the basins surrounding the abyssal plain. This feature is consistent with the fact that the Early to Middle Permian is the interval when the most common occurrence of platform deepening, drowning and collapsing is documented in the areas surrounding the Ionian and Levantine seas.

(d) Symmetric magnetic anomalies have not been mapped in the Ionian and Levantine oceanic crust. This would be expected if a Permian even low-spreading-rate emplacement of this oceanic crust had occurred during the long-lasting Permo–Carboniferous reversed polarity (Kiaman) superchron (Menning, 1995; Menning et al., 2000)

(e) A subcircular belt with special features follows the margins of the Ionian and part of the Levantine seas separating them from the platform margins to the south and the mountain chains to the north. Unlike platforms and chains where the crystalline basement and or early Palaeozoic rock are known at the surface or in the subsurface, the special belt does not contain sedimentary rocks older than Permian nor evidence of sialic basement. This holds for Sicily, the Southern Apennine chain, the south Hellenic islands, and some of the Taurus nappes (Demirtasli, 1990; Papanikolau and Sideris, 1990; Vai, 1994). Of special interest is the evidence provided by the diatremes of Mesozoic to Quaternary age cutting across the Triassic to Recent sedimentary sequence of the Hyblean plateau (east Sicily) where only mafic to ultramafic exhumed xenoliths are found. This suggests a pre-Triassic emplacement of an oceanic lithosphere at the east Sicily to Malta escarpment (Scandone et al., 1981; Bianchi et al., 1989; Vai, 1994).

(f) The main rifting axis located in the south Mediterranean Ionian and Levantine areas during the Carboniferous to Permian times jumped or propagated northward into the Lagonegro, La Spezia, Pindos, Budva and similar basins on both sides of the Adriatic European promontory during Middle and Late Triassic (Vai, 1994). The final northward jump of the rift axis occurred during the Middle Jurassic in the Alpine Carpathian areas coupled with the dramatic change from dextral to sinistral strike-slip motion.

7. Palaeobiogeographic cross-check

The set of data compiled within the assumed palaeogeographic framework are internally consistent with the interpretation provided, although some of them are still inferential. The model of a stepwise dynamic Pangaea development, with an aborted Early Permian Pangaea break-up and a Late Permian Pangaea restoration is convincingly supported. In particular, a trans-Pangaea seaway connecting western Tethys with Panthalassa through the Mediterranean, the northwest Africa offshore and the Caribbean area would fit well the observed lithofacies, geometric and structural data. To cross-check independently the model above (F. Cecca, pers. commun.), the major palaeobiogeographical scenarios during Late Carboniferous to Permian times were tested against our maps (Figs. 9–11).

The Late Carboniferous (about Westphalian) distribution of Euramerian floras (Fig. 9) was

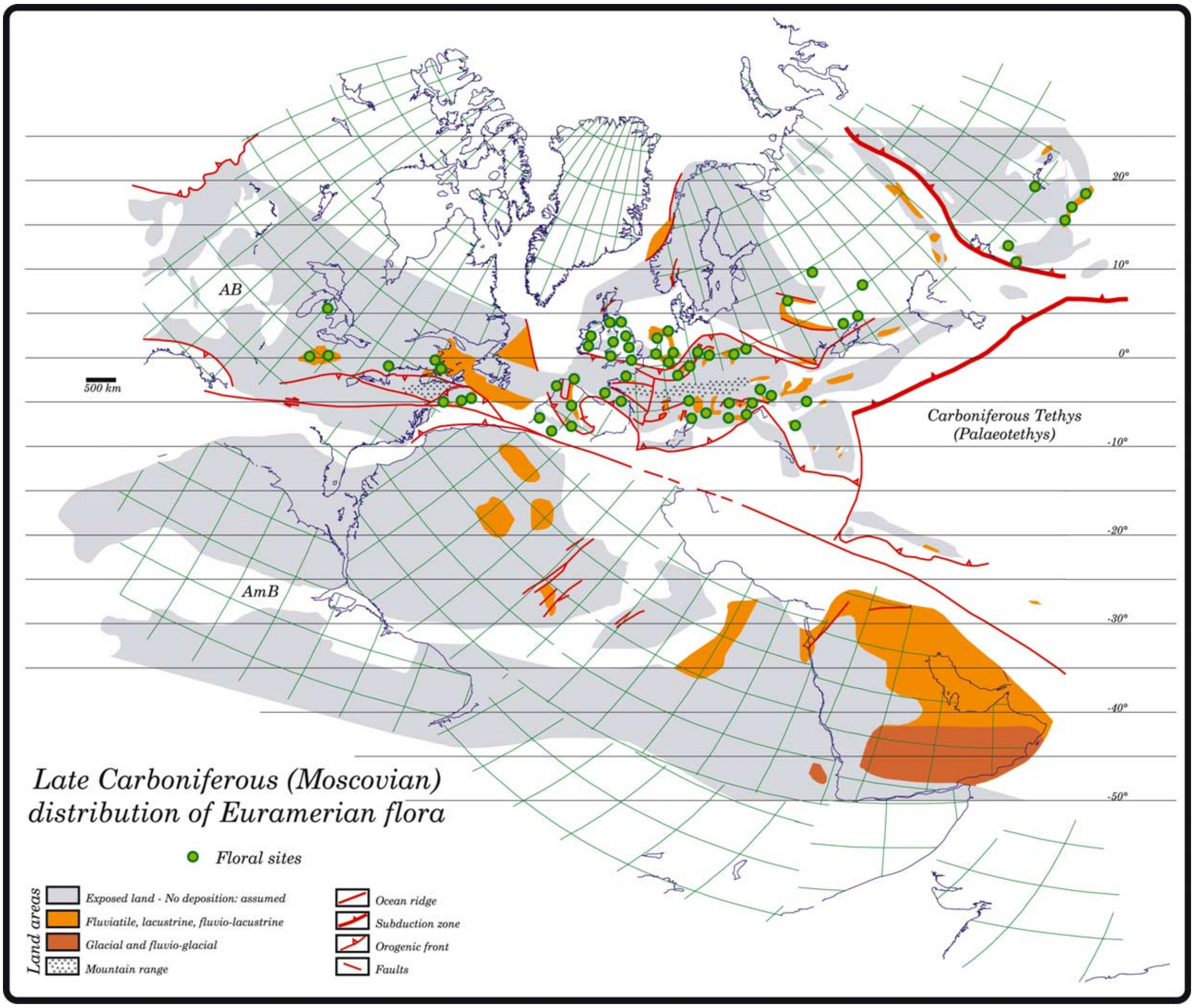

Fig. 9. Late Carboniferous (about Westphalian or Moscovian) distribution of Euramerian floras. (Marine areas in white; compare to Fig. 6.)

compiled after Chaloner and Meyen (1973) and updated (Chaloner and Creber, 1988; Meyen, 1987; Ziegler, 1990; Utting and Piasecki, 1995). Although great caution is needed to correctly implement data on floral distribution for palaeogeographical purposes (Laveine et al., 1999, 2000; Ziegler, 1990), the Euramerian realm is sharply confined to the south by the trans-Pangaea seaway (Fig. 9). The Euramerian flora was related to the Carboniferous tropical rainforest (Ziegler, 1990) fitting well the palaeoequatorial belt in our map (Fig. 6). Euramerian floral outcrops are limited to Laurussia, the seaway acting as a barrier to expansion of this flora to palaeoequatorial land areas of Gondwanaland.

The Early Permian reptile distribution (Fig. 10) mirrors the floral one (Fig. 11). Late Carboniferous reptilians are known from Laurussia (including recent findings in the Carnic Alps, Tuscany and Sardinia), but are lacking in Gondwanaland (Schneider, 1996). According to data compiled from Romer (1973) and Millsteed (1995), there is a rich Early Permian continental reptile fauna in Laurussia, whereas the Gondwanaland record is restricted to the family Mesosauridae (of uncertain, possible amphibian adaptation) sharing a

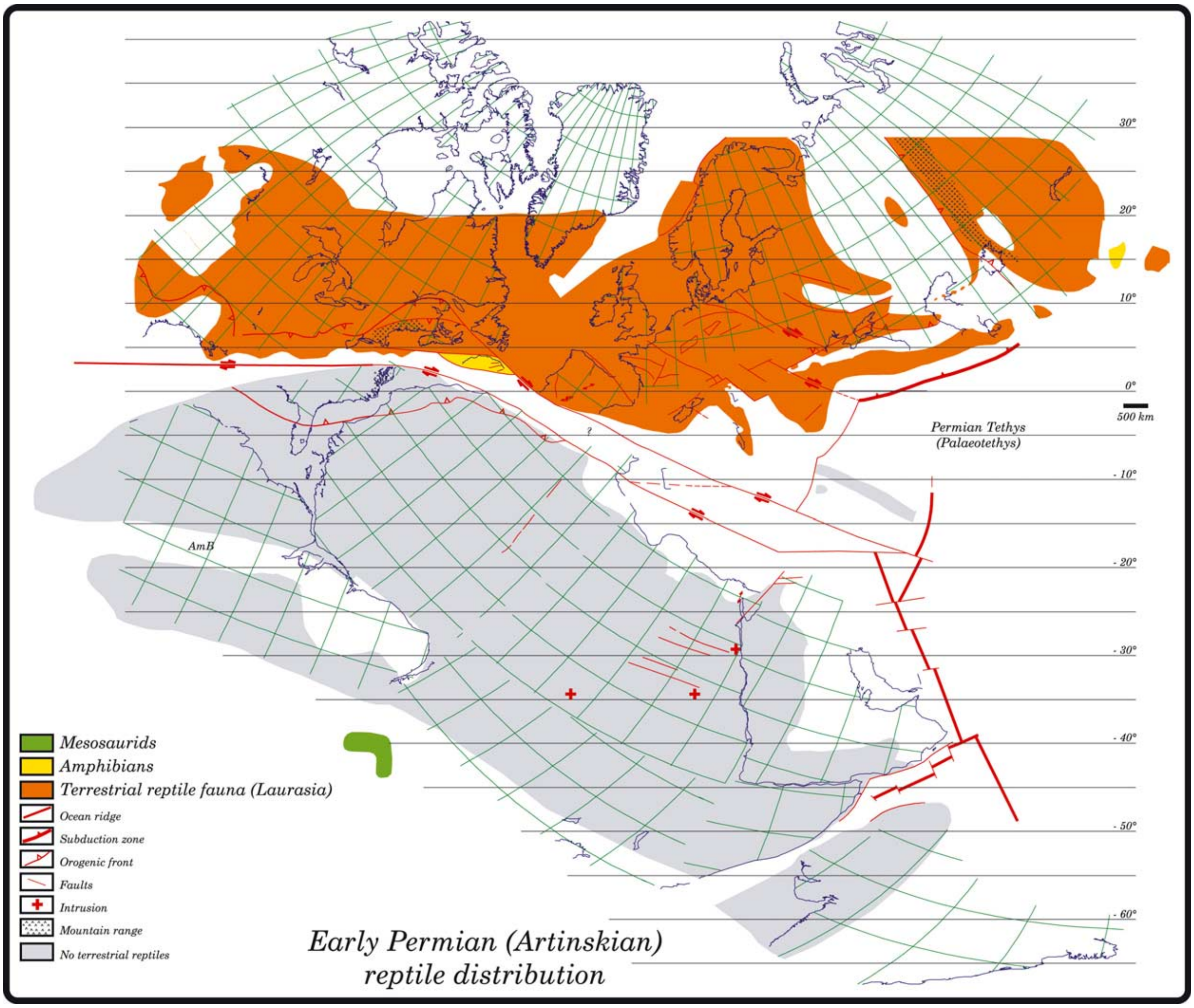

Fig. 10. Early Permian (about Artinskian) reptile distribution. (Marine areas in white; compare to Fig. 7.)

limited area of Brazil and South Africa, and to amphibians in Kashmir and Morocco. In Kazanian/Tatarian time, a first increase in diversity of reptiles occurs in Gondwanaland following the establishment of continental migration routes from Russia. This means that terrestrial reptiles do not appear in the fossil record of Gondwanaland until the Late Permian (Olson, 1979). A full reptile interchange between Laurasia and Gondwanaland is achieved only with the latest Permian, despite the dominance of terrestrial sediments in that area during late Palaeozoic.

The Early Permian (about Artinskian) floral distribution (Fig. 11) shows part or most of the four major realms of that time: the Euramerian (more precisely subdivided in North American and Atlantic), the Angaran and Cathaysian realms for the Laurasian northern floras (Chaloner and Meyen, 1973; Plumstead, 1973) and the *Glossopteris* realm for the Gondwanan southern flora. It is worth mentioning the strong floral fragmentation in the northern hemisphere, contrasting with the relative uniformity of the *Glossopteris* flora in the southern one (Gondwanaland). The map shows a sharp boundary between the *Glossopteris* realm on the Gond-

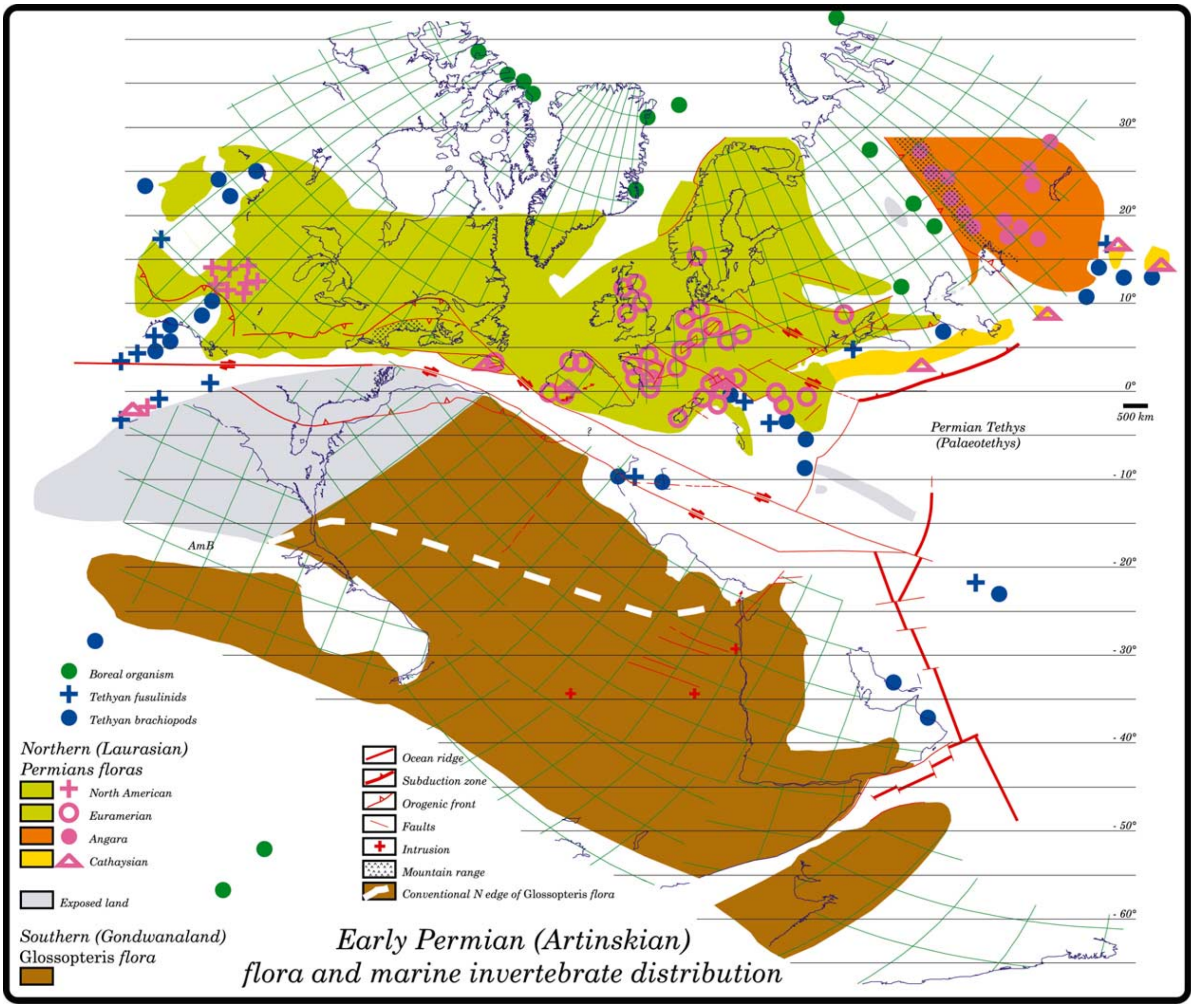

Fig. 11. Integrated distribution of Early Permian (about Artinskian) floral realms and marine benthic invertebrates. (Marine areas in white; compare to Fig. 7.)

wanan supercontinent and all other realms in the Laurussian+Asian supercontinent. The *Glossopteris* realm is sharply limited to the north by the inferred trans-Pangaea seaway. The factor controlling the distribution seems to be the physical barrier to the north rather than latitude and climatic belt. In fact, the map is consistent with the distinction of different climate-related biomes (Ziegler, 1990) within the Gondwanan (*Glossopteris*) realm. The segmentation of the North American/Atlantic, Angaran and Cathaysian floras is partly geographic (distance, physical barriers) and partly latitudinal, with the Cathaysian flora indicating a clear oceanic climatic control, along with the anticlockwise dispersal, amalgamation and accretion of Gondwanan fragments (terranes) toward Asia (Metcalfe, 1998; Laveine et al., 1999). A stable vs time-changing ratio between elements of different bioprovinces (mixed faunas and floras) in discrete blocks (terranes) of southeast Asia allows reconstruction of different patterns of dispersal (Shi et al., 1995; Shi and Archbold, 1998). Mixed floras are also known from northwest Venezuela, west Texas (Ricardi et al., 1997), Spain, Morocco, Niger and Gabon (Broutin et al., 1990, 1998). According to Ziegler

(1990), the Atlantic (Euramerian) realm floras were peripheral to the rainforests (drier with younging Permian age), the Cathaysian was a tropical realm with ever-wet and summer-wet forests, the Angaran was a cool temperate to cold realm (except for the Subangaran Kazakhstan province having a mediterranean climate), and the Gondwanan was a cool temperate realm with extensions to warm temperate on one side and cold to glacial on the other side. As the *Glossopteris* flora was not directly controlled by climate, it was potentially able to spread over the Laurussian supercontinent. This migration, however, was hampered by the trans-Pangaea seaway until Late Permian (Broutin et al., 1995, 1998; Berthelin et al., 1999) when the seaway disappeared (see above) and a route for floral incursions and interchange between the northern and southern hemispheres was open.

The Early Permian distribution of some marine benthic groups, such as the warm Tethyan brachiopods and fusulinids vs boreal forms compiled after Gobbett (1973) and Stehli (1973) and updated (Shi et al., 1995; Stevens et al., 1990), is also shown in Fig. 11. A similar climatic setting is provided by the distribution of Permian sponges (Rigby and Senowbari-Daryan, 1995) and even of Permian conodonts (Henderson and Mei, 2000). Earlier authors have suggested westward migration to explain the typical Tethyan fauna occurring in Bolivia, Venezuela, central America and Texas (Gobbett, 1973; Stehli, 1973; Stevens, 1984). The migration would have been easy and direct along the inferred Early Permian trans-Pangaea seaway. Long-shore migration of benthic organisms through the high-latitude belts was not allowed as documented by the occurrence of boreal organisms settling the Arctic, north Canadian and South American coastal belts. Even the Tethyan fauna occurring in the Uralian sea during the earliest Permian was replaced by boreal forms as soon as the Karpinsky swell interrupted the connection to the Tethys (Davydov and Leven, 2003). Transoceanic eastward migration of Tethyan benthos at the larval stage, even if theoretically possible, in our case was hampered by the reconstructed westward pattern of global warm currents (Ross and Ross, 1990), as well as by the evidence of westward migration from a Russian platform spreading point. The alternative suggestion made by some authors (Ross and Ross, 1990; Yan and Yin, 2000; Hongfu Yin, personal communication) that Tethyan faunas could have travelled eastwards across Panthalassa along with Tethyan terrane dispersal (if at all compatible with the time available by the given displacement rate and the date of welding) would be useless for the Tethyan faunas contained in parautochthonous units.

Overlapping the data provided by the groups of organisms discussed above, an integrated distribution at the Early Permian is obtained (Fig. 11). It shows that the palaeobiogeographical check is well consistent with the trans-Pangaea seaway inferred from facies and palaeodynamic analyses. The restored palaeogeographic setting and especially the trans-Pangaea seaway, temporarily open during the Early to Middle Permian, explain the different palaeobiogeographic problems and the overall setting, including the Late Permian to Triassic evolution.

8. Summary and conclusion

Current Permo–Carboniferous palaeogeographic restorations follows a static Pangaea A-type approach. Available data, augmented by new Moscovian and Artinskian data plotted on a present-day map, within the framework of PTP (the Peri-Tethys Programme), show:

(1) A Moscovian thalassocratic regime over Hercynian Europe and its northern Caledonian and southern Panafrican foreland. Epicontinental seas were present in the Russian platform, northern Europe and north Africa. Hercynian Europe and the Near East were fragmented by en échelon seaways closed to the northwest and open to the southeast: the mid-Black Sea–Caucasian, the Iran–Anatolian–Dinaric–Carnic–trans-Hungarian, the Apenninic–Cantabrian, and the south Portuguese seaways. Convergence/subduction was active along the west Siberian–southwest Kazakhstanian volcanic margin, and along the Mauretanian margin. In between extension to small-scale transtension dominated.

(2) In the Artinskian, major coastline retreat and regression are shown worldwide: in northwestern Europe, north Africa, Donets and Russian platform basins. However, the Tunisia–Oman belt shows deepening sedimentation up to the emplacement of oceanic crust to the east. Granitoids and bimodal volcanics are common in narrow belts inside and outside Hercynian Europe. This is consistent with uplift/collapse of Hercynian Europe and the following fragmentation related to the Early Permian dextral megashear. The axis of the oblique rift leading to the seafloor spreading of the Permian Tethys is placed along the Oman–Levantine–Ionian seaway. This south Mediterranean belt is the best location for the large dextral slip transforming a Pangaea B-type setting into a Pangaea A-type, convergence still being active at the Uralian and Mauretanian margins.

(3) An Early Permian shallow marine seaway connecting the Permian western Tethys to the eastern Panthalassa in the Texan–Bolivian areas would fit the Permian migration of Tethyan fusulinids and brachiopods through Cuba into central America. It would also prevent exchange of floras and continental tetrapods between Laurussia and Gondwanaland. This model is consistent with a temporary transtensional dextral slip between Gondwanaland and Laurussia lasting some tens of million years when the last shortenings occurred in the Ouachita–Allegheny–Mauretanian and Uralian belts. Outcomes of the model for Pangaea evolution are: (a) a first quasi-Pangaea stage at the Carboniferous/Permian boundary; (b) an aborted Pangaea break-up stage during the Early Permian (separating Laurussia from Gondwanaland by a shallow sea from Tunisia to Cuba); and (c) the full Pangaea stage during Late Permian to Triassic.

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