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Permo–Triassic intraplate magmatism and rifting in Eurasia: implications for mantle plumes and mantle dynamics

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Abstract

At the transition from the Permian to the Triassic, Eurasia was the site of voluminous flood-basalt extrusion and rifting. Major flood-basalt provinces occur in the Tunguska, Taymyr, Kuznetsk, Verkhoyansk–Vilyuy and Pechora areas, as well as in the South Chinese Emeishen area. Contemporaneous rift systems developed in the West Siberian, South Kara Sea and Pyasina–Khatanga areas, on the Scythian platform and in the West European and Arctic–North Atlantic domain. At the Permo–Triassic transition, major extensional stresses affected apparently Eurasia, and possibly also Pangea, as evidenced by the development of new rift systems. Contemporaneous flood-basalt activity, inducing a global environmental crisis, is interpreted as related to the impingement of major mantle plumes on the base of the Eurasian lithosphere. Moreover, the Permo–Triassic transition coincided with a period of regional uplift and erosion and a low-stand in sea level. Permo–Triassic rifting and mantle plume activity occurred together with a major reorganization of plate boundaries and plate kinematics that marked the transition from the assembly of Pangea to its break-up. This plate reorganization was possibly associated with a reorganization of the global mantle convection system. On the base of the geological record, we recognize short-lived and long-lived plumes with a duration of magmatic activity of some 10–20 million years and 100–150 million years, respectively. The Permo–Triassic Siberian and Emeishan flood-basalt provinces are good examples of "short-lived" plumes, which contrast with such "long lived" plumes as those of Iceland and Hawaii. The global record indicates that mantle plume activity occurred episodically. Purely empirical considerations indicate that times of major mantle plume activity are associated with periods of global mantle convection reorganization during which thermally driven mantle convection is not fully able to facilitate the necessary heat transfer from the core of the Earth to its surface. In this respect, we distinguish between two geodynamically different scenarios for major plume activity. The major Permo–Triassic plume event followed the assembly Pangea and the detachment of deep-seated subduction slabs from the lithosphere. The Early–Middle Cretaceous major plume event, as well as the terminal–Cretaceous–Paleocene plume event, followed a sharp acceleration of global sea-floor spreading rates and the insertion of new subduction zone slabs deep into the mantle. We conclude that global plate kinematics, driven by mantle convection, have a bearing on the development of major mantle plumes and, to a degree, also on the pattern of related flood-basalt magmatism.

Keywords: Eurasia; Rifts; Mantle plumes; Flood-basalts; Environmental crisis; Permian; Triassic; Plate tectonics; Geodynamics

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1. Introduction

The extrusion of voluminous flood-basalts, which characterize so-called "Large Igneous Provinces" (LIPs), is generally attributed to the upwelling of large-scale mantle plumes (super-plumes), which rise from the core–mantle boundary and impinge on the lithosphere (Coffin and Eldholm, 1992, 1994; Kent et al., 1992; Wilson, 1992, 1997; Dobretsov, 1997a,b; Courtillot et al., 1999). Dynamic processes controlling the development of such LIPs, as well as their relation to plate tectonics, are subjects of on-going discussion. Moreover, recently, serious questions have been raised on the validity of the entire mantle plume concept (Smith and Lewis, 1999; Sheth, 1999).

In this paper, we present evidence for the association of Permo–Triassic, presumably super-plume related volcanism in Asia and China with a major plate boundary reorganization. Evidence of this plate reorganization is preserved in the sedimentary record of numerous Permo–Triassic rift systems and major flood-basalt provinces (Veevers, 1995; Ziegler et al., 2001; Ziegler and Stampfli, 2001). We suggest that this plate reorganization, which involved the detachment of deep-reaching inactive subduction slabs and the development of new subduction zones, was accompanied by a reorganization of the global mantle convection system during the early phases of which mantle convection was not fully able to provide the necessary heat transfer from the Earth's core to the surface. This was presumably accompanied by a thickening of the D″ layer at the core–mantle boundary and the ejection of material from it that rose in the form of mantle-plumes to the base of the lithosphere, triggering the extrusion of voluminous flood-basalts (Larson, 1991; Storey et al., 1992; Prevot and Perrin, 1992).

Latest Palaeozoic to Early Mesozoic times correspond to a period of global plate boundary and plate kinematics reorganization that marks the transition from the amalgamation of Pangea to its break-up. It has been inferred that this lithosphere-scale reorganization was accompanied by a reorganization of the deep mantle convection system (Ziegler, 1993; Nikishin and Ziegler, 1999; Ziegler et al., 2001).

The Permo–Triassic transition also coincides with major changes in the Earth's environment (Hardie, 1996; Faure et al., 1995; Dobretsov, 1997a,b; Kozur, 1998; Yin and Tong, 1998) that are held responsible for a major mass extinction, presenting a sharp incision in the evolution of life (Sepkoski, 1990; Erwin, 1994; Walliser, 1995; Moldowan et al., 1996; Alekseev, 1998; Kozur, 1998; Yin and Tong, 1998). Tectonic and magmatic events that occurred at the Permo–Triassic boundary have been discussed in numerous papers and were considered as possible factors responsible for the observed environmental changes (e.g. Ziegler, 1990; Courtillot, 1994; Khain and Seslavinsky, 1994; Veevers, 1995; Veevers and Tewari, 1995; Walliser, 1995).

The Permian and Triassic evolution of Western Europe is relatively well known (Ziegler, 1990; Dercourt et al., 1993; Yilmaz et al., 1996). Data for Eurasia and other continents have been discussed in many publications (e.g. Khain and Seslavinsky, 1994; Milanovsky, 1996; Veevers, 1995; Yin and Tong, 1998). In this paper, we attempt to evaluate the tectonic stress regime of Eurasia at the Permo–Triassic transition, and partly also of other continents, in an effort to assess the potential role of tectonic and magmatic processes as controlling factors of the observed large-scale biological and environmental changes (Kozur, 1998).

Preliminary data show that at the Permo–Triassic boundary regional tensional stresses built up in an area that extended from Western Europe to Eastern Siberia and from Taymyr to Middle Asia. These stresses controlled the development of a system of rifted basins. At the same time, voluminous intracontinental flood-basalts were extruded, for instance in Siberia (West Siberia rift system, Tunguska, Taymyr and Kuznetsk flood-basalts), in the Pechora Basin and in southern China. This magmatic activity, partly associated with rifting, may reflect a discrete pulse of plume activity (Dobretsov, 1997a,b; Nikishin and Ziegler, 1999). We propose that Permo–Triassic rifting and mantle-plume activity was dynamically related to the terminal–Palaeozoic–Early Mesozoic global plate boundary reorganization that followed the assembly of the Pangea supercontinent (Ziegler and Stampfli, 2001). However, as the correlation of Late Permian and Early Triassic series has to contend with major biostratigraphic uncertainties (Menning, 1995), we are unable to provide a precise correlation of the different events and, thus, have to accept a potential error of the order of 1–5 million years. For the Permian and Triassic, we have adopted the numerical time scale of Menning (1995).

Many authors have addressed the problems of plume-related magmatism and continental rifting (White and McKenzie, 1989; Larson, 1991; Storey et al., 1992; Coffin and Eldholm, 1992, 1994; Dobretsov, 1997a,b; Courtillot et al., 1999; Smith and Lewis, 1999). In this paper, we focus on three specific issues, namely: (1) can mantle plumes simultaneously affect the lithosphere of different continents and oceans, (2) do global plate kinematics influence the pattern of plume-related magmatism, and (3) what is the duration of activity of a specific plume. One of the best examples, which can be used to address these problems, are the intracontinental Permo–Triassic flood-basalt and rift provinces of Eurasia. These provide an opportunity to compare the timing of rifting and flood-basalt extrusion and to assess their geodynamic setting.

2. Permo–Triassic rift systems and flood-basalt provinces of Eurasia

At the end of the Palaeozoic, assembly of the Pangea supercontinent was completed with the suturing of Laurussia (Baltica and Laurentia) and Gondwana along the Variscan–Appalachian orogen, of Laurussia and Kazakhstan–Siberia along the Uralian orogen, and of Kazakhstan and Eastern Siberia along the Central Asian orogen (Ziegler, 1989, 1990; Kazmin and Natapov, 2000) (Fig. 1).

The distribution of the main Permo–Triassic flood-basalt provinces and rift systems of Eurasia are shown in Fig. 2. They are largely superimposed on Palaeozoic orogens, as in the West-European and Atlantic regions (Ziegler, 1990), in Western Siberia (Surkov, 1995), in the Scythian belt along southern margin of the East-European Craton (Nikishin et al., 1998a,b, 2001), in the Timan-Barents Sea region (Shipilov and Tarasov, 1998; Gavrilov, 1993; Stupakova, 1996), and

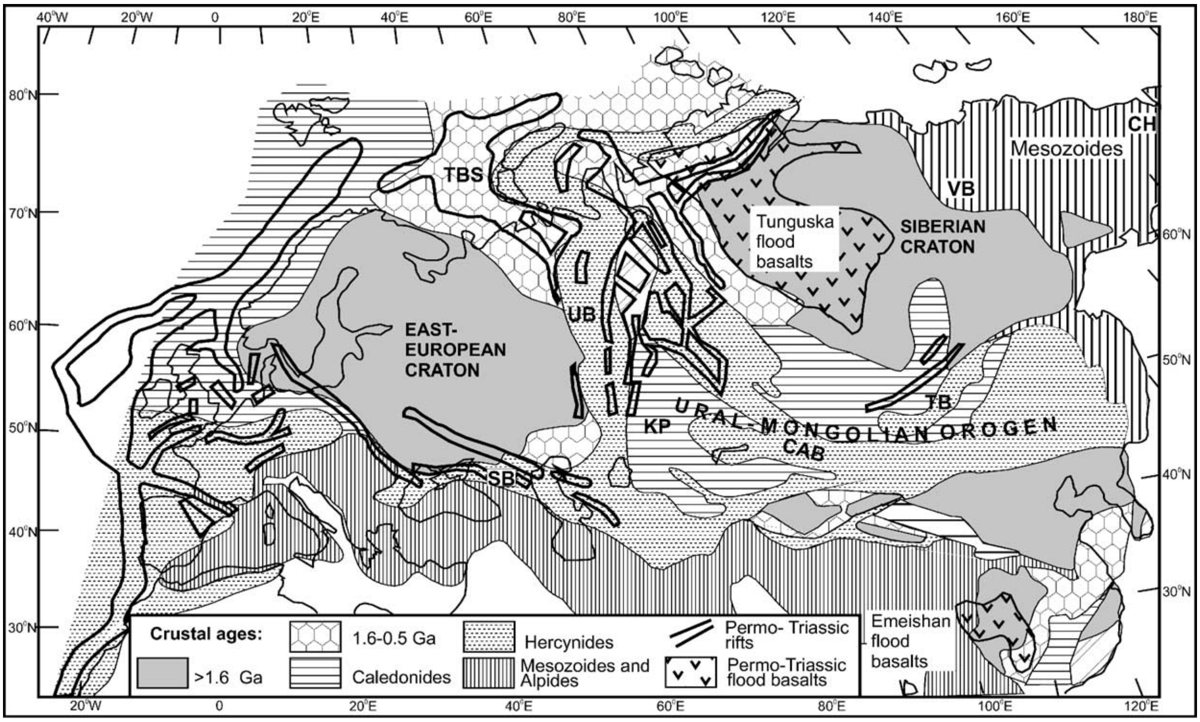


Fig. 1. Basement provinces of Eurasia, showing location of Permo-Triassic continental rifts and flood basalt provinces. CAB—Central Asian belt; CH—Chukotka terrane; KP—Kazakhstan plaeocontinent; SB—Scythian belt; TB—Transbaikal region; TSB—Timan–Barents Sea region; UB—Ural belt; VB—Verkhoyansk belt.

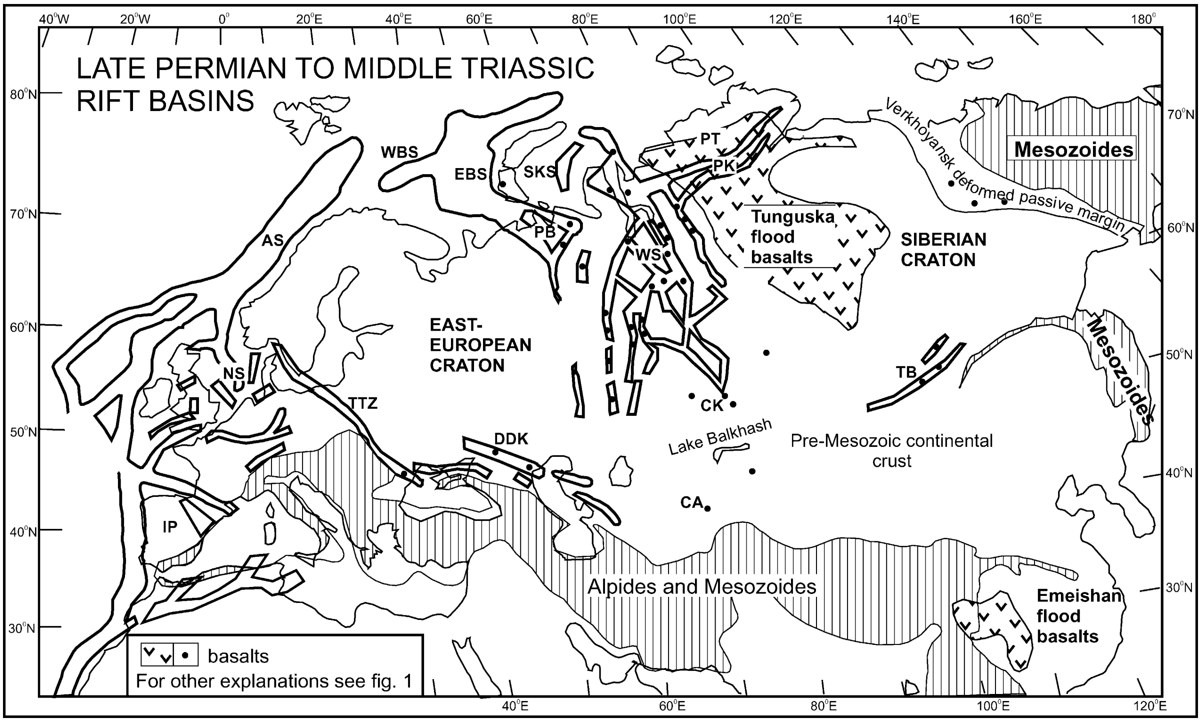


Fig. 2. Late Permian to Middle Triassic rifted basins and intraplate food basalt provinces. AS—Atlantic shelf rift systems; CA—Central Asia; CK—Central Kazakhstan; DDK—Dniepr–Donets–Karpinsky rift belt; EBS—East Barents Sea rift system; IP—Iberian Peninsula; NS—North Sea rift system; PB—Pechora Basin rift system; TP—Taymyr flood basalt province; SKS—South Kara Sea rift system; TB—Transbaikal; TTZ—Teisseyre–Tornquist Zone; WBS—West Barents Sea rift system; WS—West Siberian rift system; PK—Pyasina–Khatanga rift system.

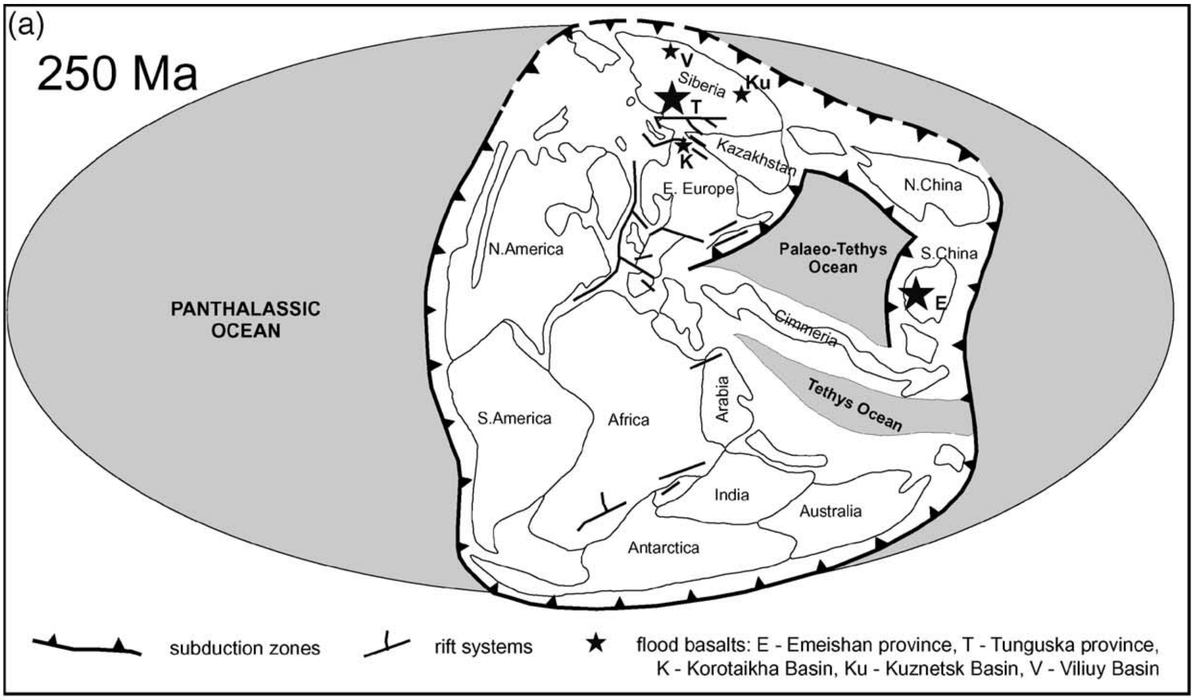
in the Transbaikal region (Milanovsky, 1996). By contrast, the Permo–Triassic rifts of Gondwana are superimposed on its Pan-African sutures (Guiraud and Bosworth, 1999; Delvaux, 2001; Ziegler et al., 2001). The Permo–Triassic rifts of Laurasia and Gondwana form a broad network that transects the central parts of Pangea (Fig. 3a,b).

In the following sections, we discuss the Permo–Triassic rift systems and magmatic provinces of Eurasia.

2.1. West and East Siberian regions

The main basement provinces of Siberia are illustrated in Fig. 1. A simplified Early Triassic palaeotectonic/palaeogeographic reconstruction of Siberia is provided by Fig. 4. The East Siberian Craton consists of Archean to Early Proterozoic crust (older than 1.6 Ga; Condie and Rozen, 1994; Khain and Nikishin, 1997) whereas the Kazakhstan palaeo-continent contains only Proterozoic and Palaeozoic crustal elements (Zonenshain et al., 1993).

The Ural–Mongolian orogen forms the suture between the East European and Siberian cratons and between the latter and the Palaeozoic Kazakhstan collage of subduction–accretion complexes (Sengör et al., 1993). The main branches of the Ural–Mongolian orogen superbelt are the Uralian and the Central Asian (or Altaid) orogenic belts. The Ural–Mongolian superbelt includes domains of Late Precambrian, Early Palaeozoic, and Late Palaeozoic crust (Khain, 1981; Zonenshain et al., 1993; Khain and Nikishin, 1997). A Late Palaeozoic collisional orogenic pulse affected most of the Ural–Mongolian superbelt. During this collision, all Late Precambrian and crustal domains were overprinted and exhumated by more than 2 km (Surkov, 1995). During Middle Permian times, the Uralian orogen was of a Himalayan type. From the Permo–Triassic transition onwards, Siberia formed part of Laurasia (and Pan-



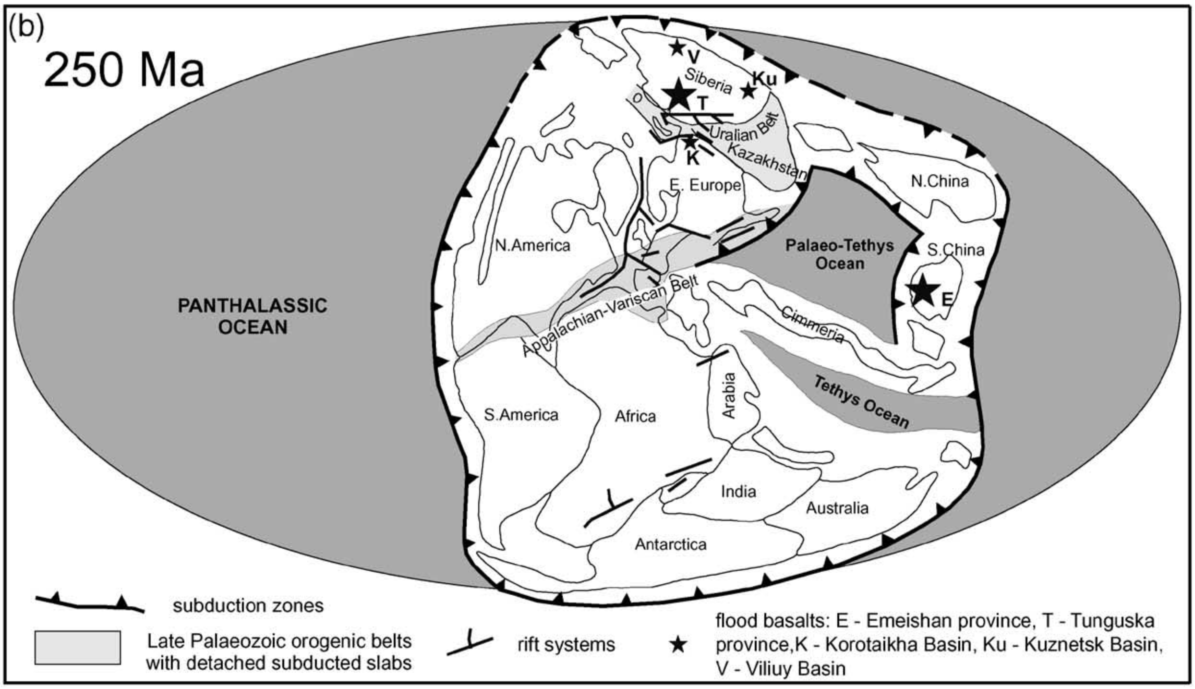


Fig. 3. (a) Permo–Triassic reconstruction (250 Ma), simplified after Scotese and Golonka (1993), showing location of Permo–Triassic continental rifts and flood basalt provinces. (b) Same with possible regions where deep-reaching subduction slabs were detached from the lithosphere.

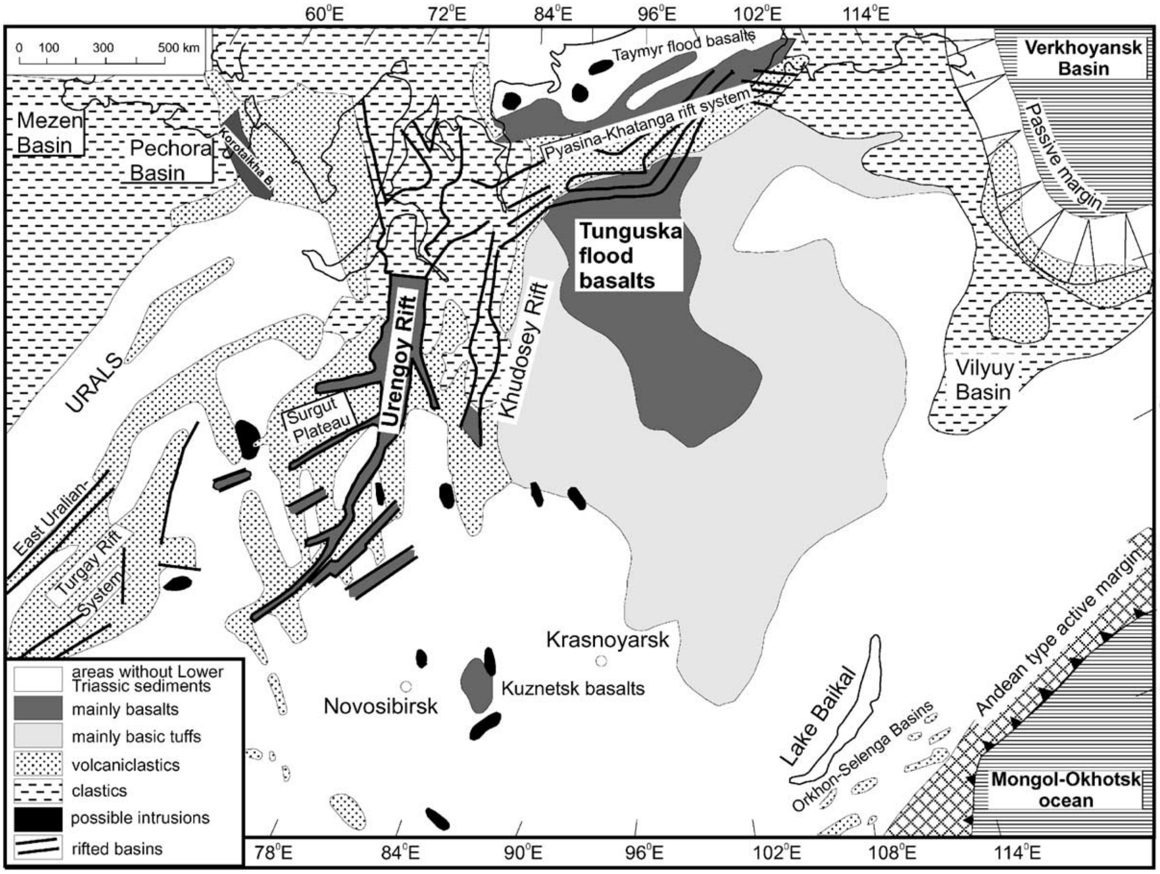


Fig. 4. Permo–Triassic rifts and magmatic provinces of Siberia. The Tyumen SG-6 well is located within the Urengoy rift (for details, see Fig. 5).

gea) and was flanked to the E–NE (in present coordinates) by the Verkhoyansk passive margin (Fig. 3). The area of this margin was affected during Middle–Late Devonian times by rifting, culminating at the beginning of the Carboniferous by crustal separation and the development of a passive margin, which persisted until the Jurassic (Parfenov, 1984; Yapaskurt, 1992). During Late Palaeozoic to Jurassic–Cretaceous times, the southeastern margin of Siberia (in present coordinates) was occupied by an Andean-type orogen that faced the Mongol–Okhotsk ocean (Parfenov, 1984; Zonenshain et al., 1993; Delvaux et al., 1995).

At the Permo–Triassic transition, Siberia was the site of major tectonic and magmatic activity, as shown by the development of the West-Siberian, Pyasina–Khatanga and South Kara Sea rift systems, the Tunguska, Taymyr and Kusnetsk flood-basalt provinces, the Transbaikal volcanic zone, the Central Kazakhstan–Central Asia region of dyke swarms and local basalt volcanism, as well as by volcanic activity on the Verkhoyansk passive margin and in the Vilyuy Basin (Fig. 4).

2.1.1. West Siberian Basin

The West Siberian Basin hosts one of the world's largest hydrocarbon provinces (Nesterov et al., 1990). It is underlain by a major Permo–Triassic to Triassic rift system that is well documented by extensive geophysical and well data (Surkov and Zhero, 1981; Surkov, 1995; Zonenshain et al., 1993; Peterson and Clarce, 1991). This rift system consists of three main branches, here referred to as the Urengoy, East Uralian-Turgay and Khudosey systems of rifted basins (Fig. 4).

The *Urengoy Rift*, also referred to as the Koltogor–Urengoy or Pur Trough rift, is the most prominent feature of the West Siberian rift system (Surkov and Zhero, 1981; Kontorovich, 1994; Surkov, 1995; Zonenshain et al., 1993). It is largely superimposed on the Central Asian Late Palaeozoic orogen. Although the structure of the Urengoy rift is mainly known from geophysical data, different authors show different configurations for it (Surkov, 1995; Kostyuchenko, 1992; Kazakov et al., 2000). The Urengoy (or Tyumen) SG-6 (or SD-6) superdeep well was recently drilled within the Urengoy rift to a total depth of 7502 m (Figs. 5 and 6), but failed to reach the basement (Khakhaev and Kaplun, 1995). In its lowermost parts, this well penetrated nearly 1000 m of mainly tholeiitic and alkali basalts and tuffs of Late Permian (Tatarian) and Early Triassic (Induan to Olenekian) age, as indicated by palynological data (Purtova, 1995; Khakhaev and Kaplun, 1995; Kazakov et al., 2000). Magnetostratigraphic data (Westphal et al., 1998) reveal an N–R–N–R–N succession, suggesting that volcanic activity persisted over a considerable span of time, possibly until the end of the Induan (245 Ma) or Olenekian (241 Ma) (Kazanskii et al., 2000). However, there is still some disagreement between palaeontological and palaeomagnetic data. The basalts were extruded under subaerial continental conditions (Yapaskurt et al., 1994; Khakhaev and Kaplun, 1995; Kazakov, 1995).

Subsidence history modelling suggests that the Urengoy Basin underwent very rapid syn-rift subsidence during the latest Late Permian and Induan, terminating during the Middle Triassic (Fig. 5). Therefore, the duration of its rift stage was no longer than 10 million years. During Middle–Late Triassic times, the Urengoy rift was characterized by a lack of volcanic activity, gentle post-rift subsidence and continental clastic sedimentation. At the Triassic/Jurassic boundary, the Urengoy Basin underwent minor uplift and erosion (mild inversion) in conjunction with an important orogenic pulse that affected the Urals–Novaya Zemlya foldbelt (Nikishin et al., 1996; Milanovsky, 1996). Gentle post-rift subsidence resumed during the Hettangian.

The buried *East Uralian–Turgay rift system* (Fig. 4) is superimposed on the Late Palaeozoic Uralian orogen and consists of a belt of grabens, which parallel the tectonic fabric of the Uralides. These basins are filled with Early, and partly Middle(?) Triassic subaerial volcanics (predominantly basalts and subordinate rhyolites) and continental clastics. These basins were inverted at the Triassic/Jurassic transition in conjunction with major orogenic activity in the Novaya Zemlya and Taymyr fold-and-thrust belts (Milanovsky, 1996).

The buried *Khudosey rift* (Fig. 4) is superimposed on the boundary between the Siberian Craton and the Ural–Mongolian orogen. Early Triassic volcanics are known from deep wells drilled in the southern part of this rift (Kazakov, 1995; Surkov, 1995).

2.1.2. South Kara Sea rift system

The South Kara Sea Basin, which forms the northern, off-shore continuation of the West Siberian Basin, is superimposed on poorly defined pre-Mesozoic basement. In the central parts of the Kara Sea Basin, Mesozoic and Cenozoic sediments attain thicknesses of up to 10 km (Fig. 6; Shipilov and Tarasov, 1998). Seismic data indicate the presence of a system of Late Permian to Middle Triassic (Shipilov and Tarasov, 1998) or Triassic (Gavrilov, 1993) rifted basins beneath Late Triassic and younger sediments. Gravity and magnetic anomalies associated with these deeply buried grabens indicate that extensive syn-rift magmatic activity had occurred in them (Shipilov and Tarasov, 1998). Therefore, it is likely that the main rifting and volcanic activity in the South Kara Sea Basin, similar to its southward continuation in the West Siberian Basin, was confined to Tatarian–Early Triassic times; however, this assumption is not supported by well data.

2.1.3. Pyasina–Khatanga rift system and Taymyr flood-basalt province

Large parts of this Permo–Triassic rift system, which straddles the boundary between the Siberian Craton and the Late Palaeozoic to Early Mesozoic South Taymyr orogen, are deeply buried beneath Jurassic and Cretaceous sediments of the Pyasina–Khatanga Basin (Fig. 4). Correspondingly, the configuration of the Permo–Triassic Pyasina–Khatanga rift system is largely defined by geophysical data (Surkov, 1995; Kazakov, 1995; Vernikovsky, 1996). However, its northern part was compressionally deformed during the Late Triassic and again during the Jurassic–Early Cretaceous Khatangan orogeny. It crops out in the folded structures of the southern Taymyr Peninsula, exposing up to 2–3 km thick Late Permian and Early Triassic subaerial basaltic lavas and pyroclastics (Zonenshain et al., 1993; Vernikovsky, 1996; Drachev, 1999; Inger et al., 1999). Palaeontological data indicate an early Induan ($\pm$ 251–248 Ma) and early Olenekian ($\pm$ 245–243 Ma) age for two main volcanic units (Egorov and Vavilov, 1992). According to velocity data, volcanics partly fill the Pyasina–Katanga rift system (Surkov, 1995; Vernikovsky, 1996). Seismostratigraphic and palaeomagnetic data indicate that, as for the West Siberian Basin, rifting and magmatic activity in the Pyasina–Khatanga Basin was confined to Late Permian and Early Triassic times (Kazakov, 1995) and, thus, followed on the heels of the Late Carboniferous–Early Permian orogeny that had affected Taymyr (Vernikovsky, 1996; Inger et al., 1999). The Pyasina–Khatanga and Taymyr flood-basalt province can be considered as forming part of the Tunguska flood-basalt province (Vernikovsky, 1996; Kurenkov, 1997).

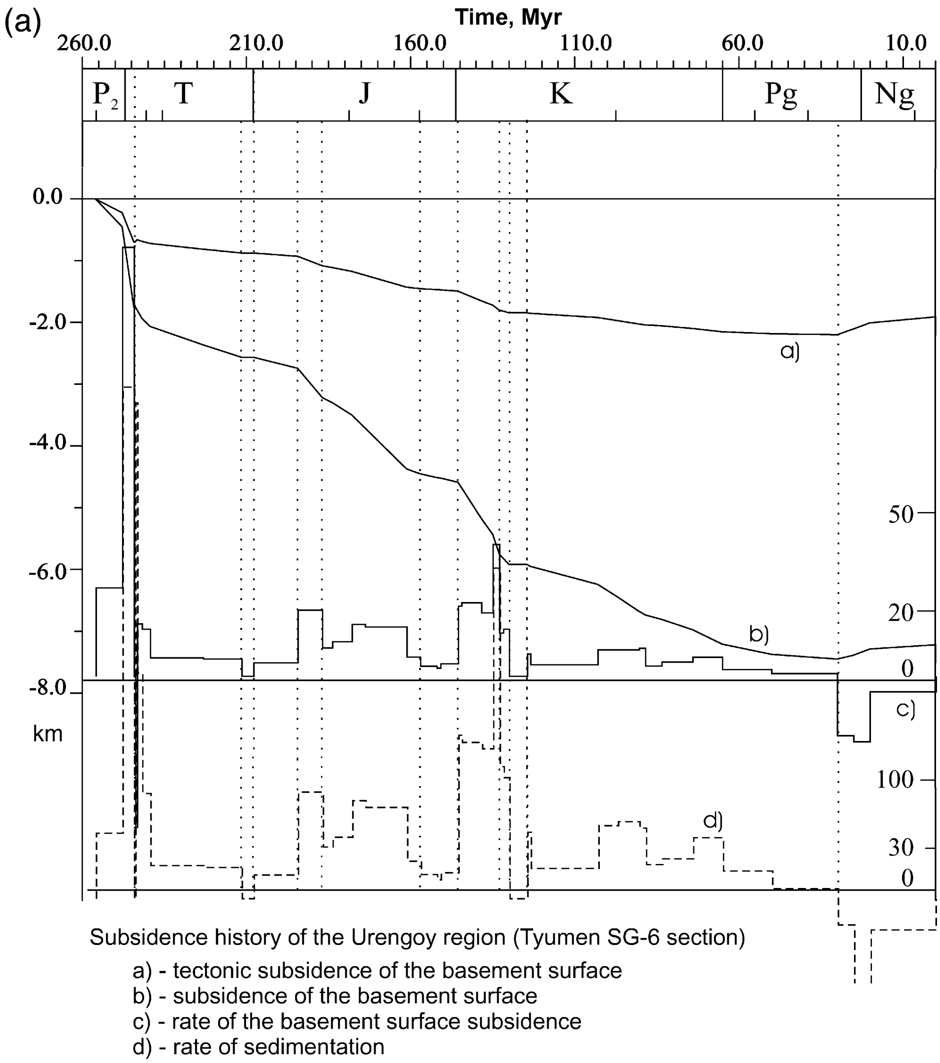


Fig. 5. Subsidence history of Urengoy rift derived from the Tyumen SG-6 super-deep well (a), and other wells (b), based on data of the Geological Survey of Russia.

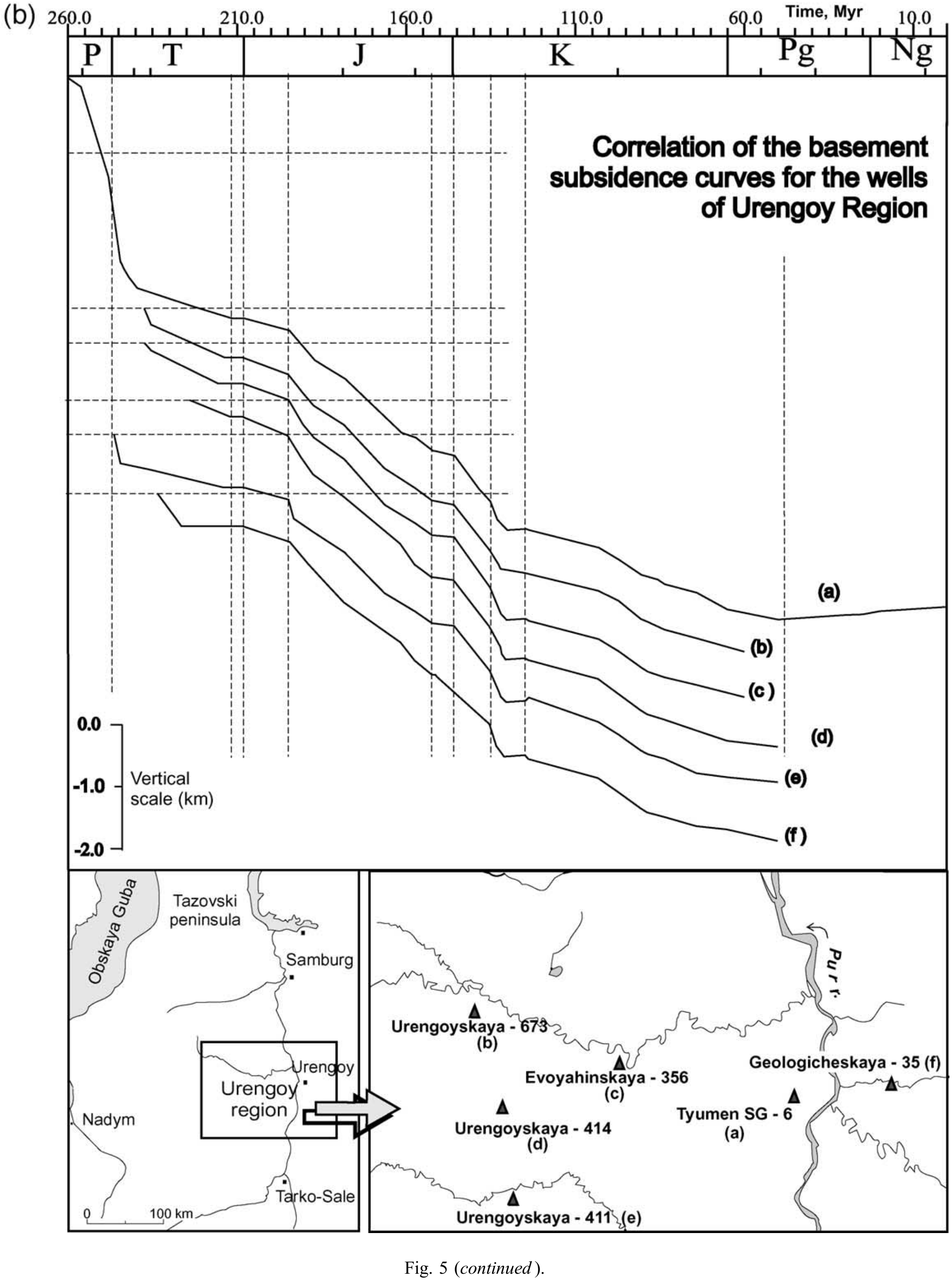


Fig. 5 (*continued*).

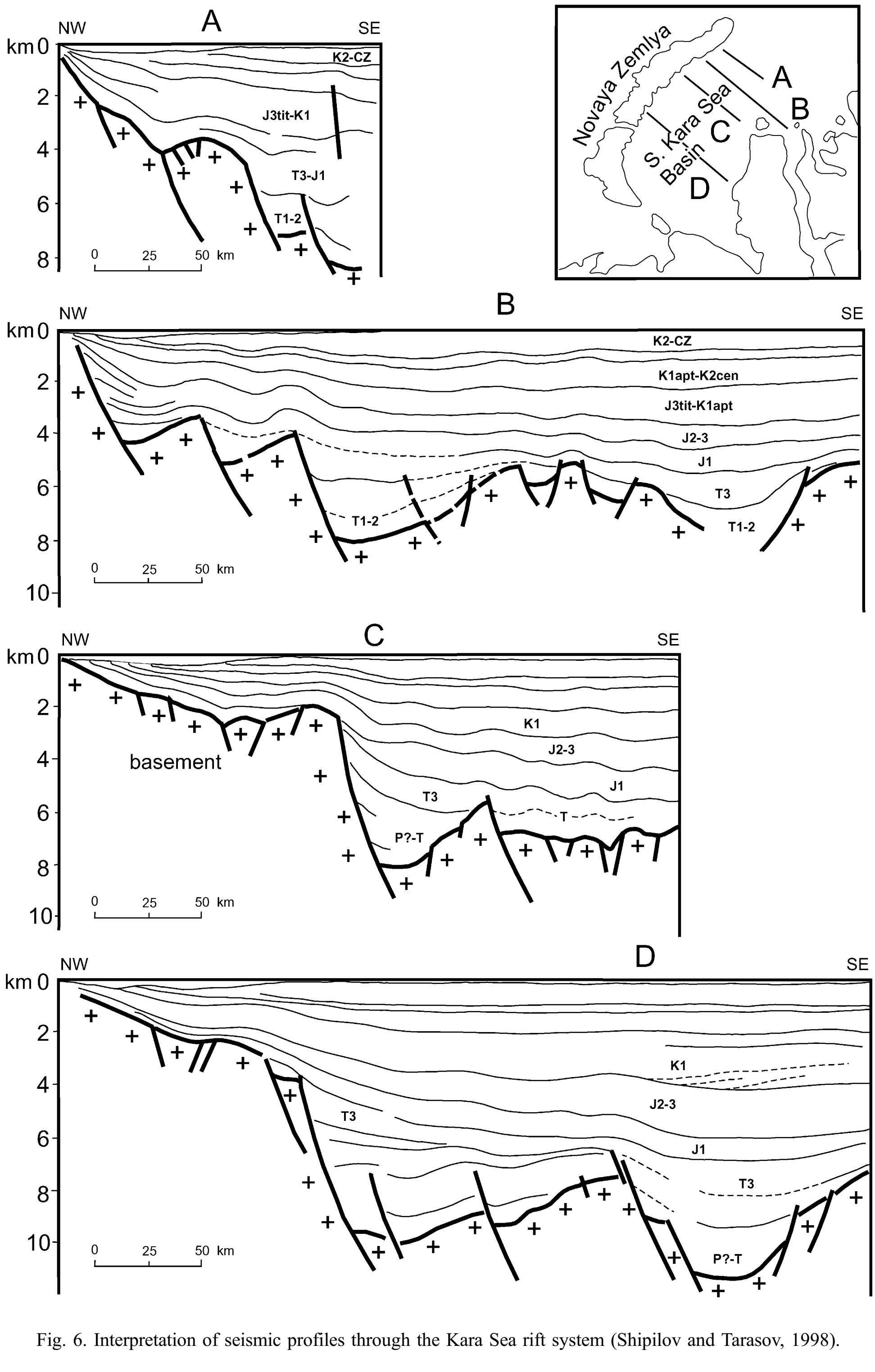


Fig. 6. Interpretation of seismic profiles through the Kara Sea rift system (Shipilov and Tarasov, 1998).

2.1.4. Tunguska flood-basalt province

The Tunguska flood-basalt province is the largest intracontinental flood-basalt province on Earth (Fig. 4). It is located in the NW part of the Siberian Craton (Polyakov, 1991; Hawkesworth et al., 1995; Kurenkov, 1997; Dobretsov, 1997a,b). The total area covered by volcanics exceeds 675,000 km^2, and the total volume of basic rocks exceeds 575,000 km^3 (Milanovsky, 1996). Well data suggest that for Tunguska alone, the total volume of basalt sills and dykes could amount to more than 748,000 km^3 (Vasiliev and Prusskaya, 1997). Together with the basalts of the West Siberian Basin and the Taymyr flood-basalt provinces, the total volume of extruded and intruded basalts amounts to 2,000,000 km^3 or more (Milanovsky, 1996).

The central part of the Tunguska province is mainly composed of subaerial basalt flows whereas a combination of tuffs and flows prevail in its peripheral parts. The thickness of this volcanic complex ranges from a few hundred meters to 2–2.4 km on the Putorana Plateau (close to Norilsk city), and 3.5 km along the northern margin of the province in the Maymecha–Kotuy region (Milanovsky, 1996). The entire volcanic complex, as well as the underlying Palaeozoic sequences, are intruded by dolerite and other basic dykes and sills. Available palaeontological data indicate that the lowermost tuffs and volcaniclastic sediments are late Tatarian in age and that the main part of the volcanics was extruded during the Induan (Lozovsky and Esaulova, 1998). The last minor volcanic pulse occurred probably during the Olenekian, as indicated by the occurrence of tuff horizons in marine sediments, northeast of the Tunguska province (Kazakov, 1995). Radiometric age determinations show that this magmatic province developed at the transition from the Permian to the Triassic (250 Ma; Renne and Basu, 1991; Campbell et al., 1992; Renne et al., 1995). On the other hand, magnetostratigraphic studies show that eruption of these volcanics commenced during the Late Carboniferous–Permian Reverse Superchron and continued after the lower Tatarian Illawarra reversal (265 Ma, Menning, 1995; Menning and Jin, 1998). This indicates a rather short time interval of perhaps less than 1 million years for the main phase of this eruptive cycle (Westphal et al., 1997; Courtillot et al., 1999). However, the total duration of volcanic activity spans the Tatarian to Olenekian times and thus lasted for about 5 million years. As some authors believe that magmatic activity persisted into the Middle Triassic (Surkov, 1995; Kazakov, 1995), additional data on its duration are needed. The geochemistry of the Tunguska flood-basalts suggests that they were derived from mantle plume material with an additional component coming from partly molten lower lithospheric mantle (Hawkesworth et al., 1995; Arndt et al., 1993).

2.1.5. Kuznetsk flood-basalt province

This province is located within the Kuznetsk Basin (Fig. 4), the fill of which consists of Devonian to Jurassic sediments that probably rest on Precambrian crust. During the Late Palaeozoic, this basin, forming the flexural foredeep of the Salair and Tom-Kolyvan orogens, was filled with molasse-type clastics. Early Triassic continental sediments contain volcaniclastic horizons and are covered by subaerial basalt flows; moreover, this basin was intruded by numerous basalt dykes and sills (Kazakov, 1995; Surkov, 1995). Stratigraphically, these volcanics rest directly on basal Triassic continental clastics, for which palaeontological data indicate an Induan age. Therefore, these volcanics are time equivalent to the main phase of the Tunguska flood-basalts (Lozovsky and Esaulova, 1998).

2.1.6. Verkhoyansk passive margin and Vilyuy Basin

Early Triassic Induan clastic sediments of the Verkhoyansk passive margin basin contain volcaniclastic material and minor basaltic flows and contemporaneous basaltic dykes (Fig. 4; Yapaskurt, 1992; Kazakov, 1995). The Vilyuy Basin is superimposed on a Middle to Late Devonian rift system (Gayduk, 1988). During Carboniferous to Jurassic times, development of this basin was governed by post-rift thermal subsidence. At the same time, this basin was incorporated into the shelf of the Verkhoyansk passive margin. In the eastern part of the Vilyuy Basin, Induan clastic sediments contain volcaniclastic material and a few basalt flows (Kazakov, 1995).

2.1.7. Transbaikal volcanic zone

During Late Palaeozoic to Mesozoic times, an active Andean-type orogen occupied the margin between the Siberian continent and the Mongol–

Okhotsk ocean (Figs. 3 and 4, Parfenov, 1984). To the northeast of this active margin, the Orkhon–Selenga back-arc rift basin developed probably during the Late Permian–Early Triassic (Yarmolyuk and Kovalenko, 1991; Delvaux et al., 1995; Gordienko, 1999). This basin was filled with subalkaline basalts, trachybasalts, rhyolites, tuffs, and clastic deposits; however, a detailed stratigraphy is not yet fully resolved. This sedimentary–volcanic sequence attains thicknesses of up to 3–5 km. Structural investigations in Transbaikal (Delvaux et al., 1995) indicate that Late Permian to Early Triassic northwest directed crustal extension controlled the subsidence of this basin.

2.1.8. Central Kazakhstan and Central Asia

The Early and Late Palaeozoic orogens of Central Kazakhstan and Central Asia (Fig. 1) underwent major Late Palaeozoic compression and formed mountain ranges during Permian times (Surkov, 1995). Central Kazakhstan is crossed by numerous small mafic dyke swarms, which cross-cut Late Palaeozoic granites. These dykes are mainly basaltic with geochemical characteristics close to those of the Tunguska flood-basalts (Spiridonov, personal communication, 2000). The age of these dykes is poorly constrained, but could be Early Triassic. In Central Kazakhstan, small remnants of Triassic basalts and rhyolite flows are associated with folded Palaeozoic strata. In Central Asia, close to the city of Tashkent, Early Triassic basalts cover a Palaeozoic calc-alkaline magmatic belt (Milanovsky, 1996). A relatively small area of Early Triassic basaltic volcanism occurs to the south of Lake Balkhsh.

2.2. Southern China

The Emeishan flood-basalt province is located along the western margin of the Precambrian Yangtze Craton of China and covers an area of at least 250,000 km^2 (Figs. 1 and 2). These trap basalts, which range in thickness from a few hundred meters to 5 km, were extruded at the transition from the Permian to the Triassic (Chung and Jahn, 1995). They unconformably overlie early Late Permian carbonates and, in turn, are covered by Triassic sedimentary sequences. Lower basaltic members are covered by trachyte and rhyolite flows and tuffs that yield radiometric ages of 251 ± 3.4 Ma (Chung and Jahn, 1995). In Southern China, the classical Meishan section exposes marine sediments that straddle the Permo–Triassic boundary; these contain biostratigraphically defined Permo–Triassic boundary bentonites that yield an age of 250 Ma (Chung and Jahn, 1995; Lozovsky and Esaulova, 1998; Yin and Tong, 1998). As this ash layer is probably related to the upper acidic member of Emeishan igneous province, the age of these basalts is very close to the Permo–Triassic boundary with the duration of volcanic activity being perhaps less than 1 million years (Chung and Jahn, 1995). Biostratigraphic correlations suggest, however, that the Emeishan volcanism commenced 8 million years before the Permo–Triassic boundary, at the end of the Guadalupian (Stanley and Yang, 1994; Courtillot et al., 1999). Yet, occurrence of the Illawarra magnetic reversal (265 Ma) in the lower Tatarian strata of Eastern Europe and in Middle Permian Guadalupian strata in North America (Menning and Jin, 1998) illustrates the problematic nature of standard biostratigraphic correlations (Ziegler and Stampfli, 2001). On this base, an earlier age was proposed for the Emeishan trap basalts (Stanley and Yang, 1994). Therefore, it is likely that flood-basalt volcanism in the Tunguska region and in Southern China was synchronous (Yin and Tong, 1998; Kozur, 1998).

2.3. Eastern Europe

Much of Eastern Europe is underlain by the Early Precambrian East–European Craton (Fig. 1; Khain, 1981; Bogdanova et al., 1996). Its southern margin is fringed by the Scythian Platform that is characterized by Carboniferous–Early Permian basement consolidation (Nikishin et al., 1996, 2001). The Carboniferous–Permian Uralian orogenic belt bounds the East–European Craton to the East. The Pechora–Barents Sea Basin, which flanks the East–European craton to the Northeast, is underlain by Late Precambrian basement (Khain, 1981; Nikishin et al., 1996). Permo–Triassic basins are widespread in Eastern Europe and include the Scythian Platform, the Dniepr–Donets, Peri–Caspian, Moscow–Mezen and Pechora–Barents Sea basins (Figs. 7 and 8; Lozovsky, 1992; Lozovsky and Esaulova, 1998; Nikishin et al., 1996, 2001).

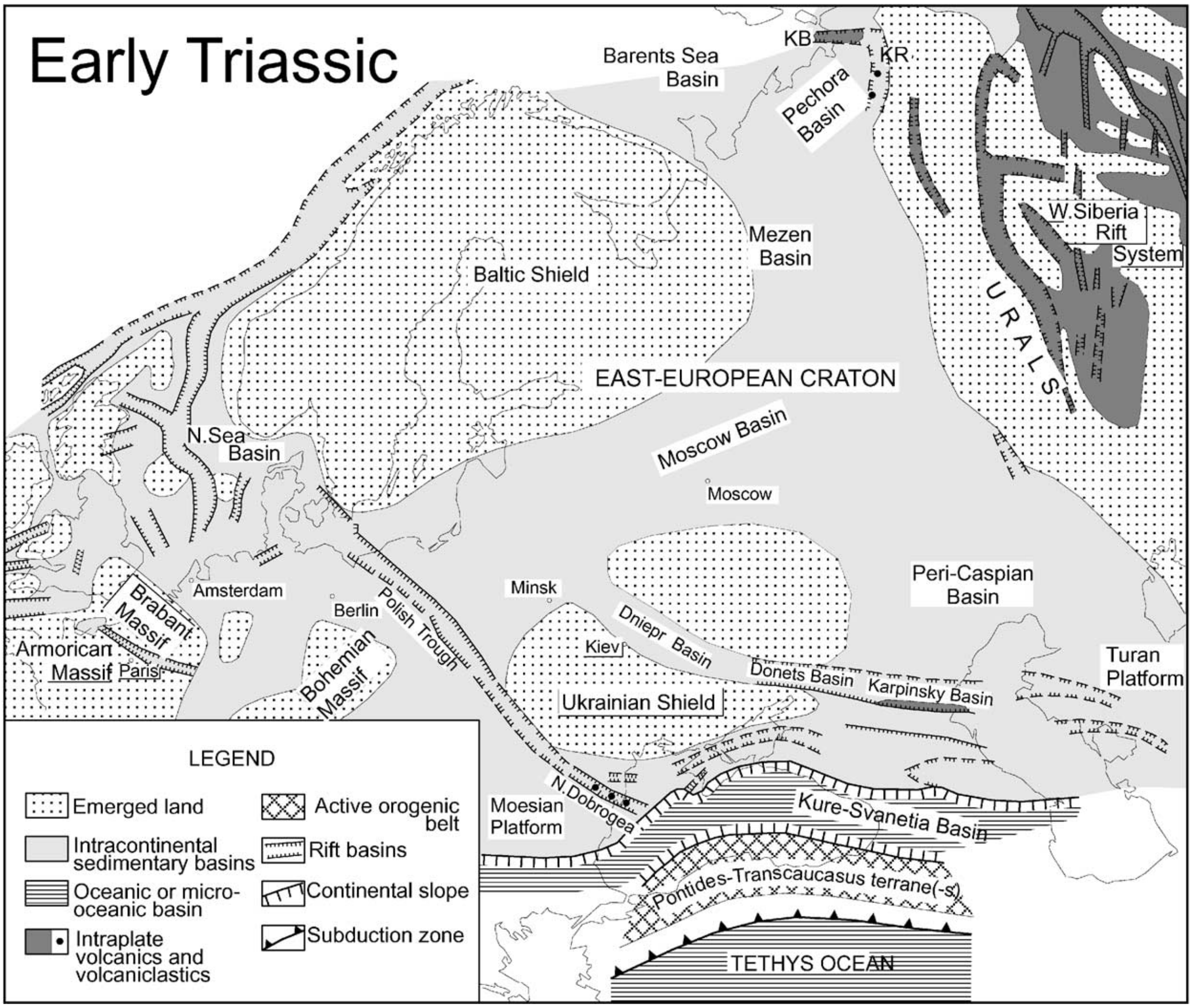


Fig. 7. Permo–Triassic rifts and basalt provinces of Europe. KB—Korotaikha Basin; KR—Kos'yu-Rogovaya Basin.

2.3.1. Scythian platform and Dniepr–Donets Basin

On the Scythian Platform, Triassic sediments are widely distributed (Nazarevich et al., 1986; Lozovsky, 1992), but were folded and partially eroded during the Late Triassic to Early Jurassic Early Cimmerian orogeny (Nikishin et al., 1998a,b, 2001). Early Triassic to lower Carnian strata occur in a number of rifted basins, comprising in the East Pre-Caucasus area the East Manych, Kayasula and Mozdok troughs, and in the West Pre-Caucasus-Crimea area, the Northern-Crimea-Azov and Novo-Fedorovsk troughs (Figs. 7 and 8; Lozovsky, 1992; Slavin, 1986; Nikishin et al., 1998b, 2001). These two areas are separated by the Stavropol High, located in the central Pre-Caucasus area. This high was probably arched up during the Early Cimmerian orogeny at the same time as the Triassic rifts on the Scythian Platform were inverted (Nikishin et al., 1998a, 2001).

In the eastern Pre-Caucasus area, sedimentation commenced under marine conditions during the Early Triassic and persisted at least until early Carnian times. Rifted basins were characterized by deeper-water conditions, whereas intervening unextended areas were occupied by carbonate platforms (Nazarevich et al., 1986; Nikishin et al., 1998a, 2001). In the East Manych Trough, located along the southern flank

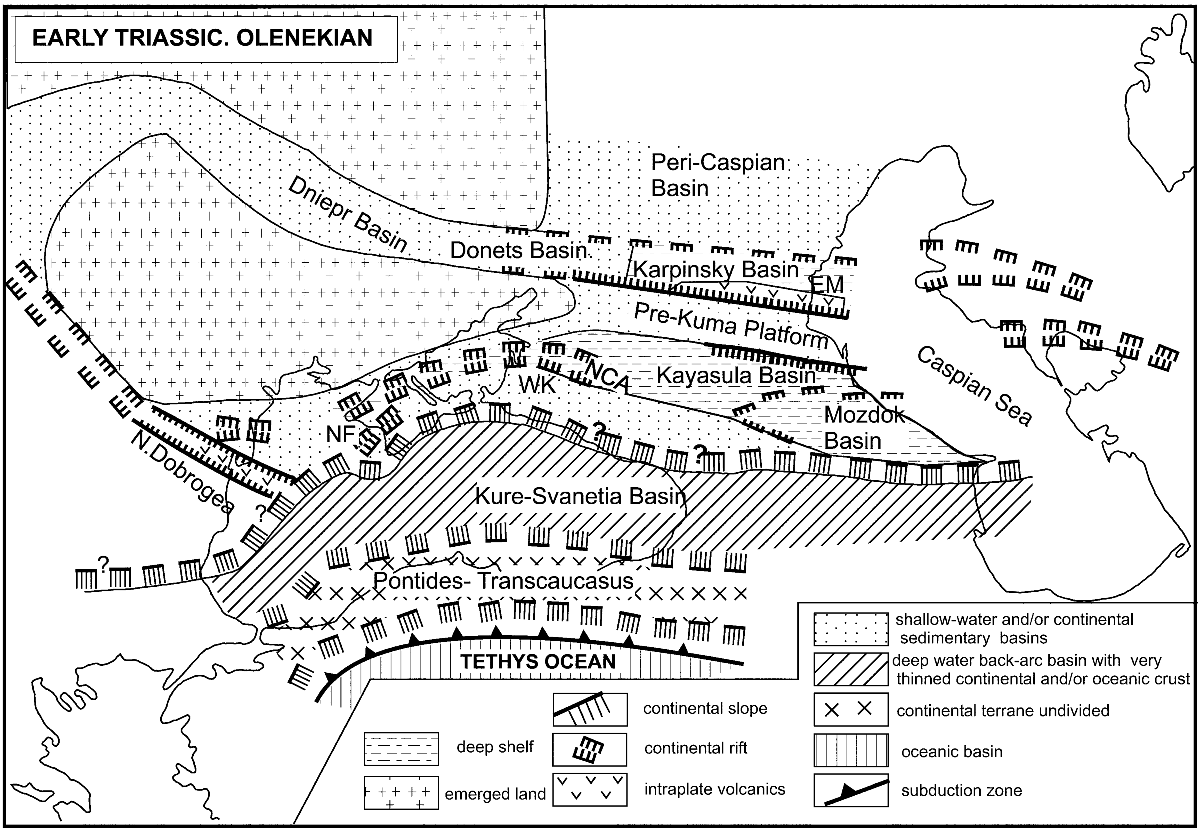


Fig. 8. Early Triassic palaeotectonic reconstruction of the Black Sea-Caucasus area. NCA—North Crimea–Azov Basin; EM—East Manych Basin; NF—Novo–Fedorovsk Basin; WK—West Kuban Basin (Nikishin et al., 2001).

of the Karpinsky Swell, shallow-water carbonates grade upward into deeper-water clays, marls, and carbonates. Subsidence of this rifted basin was accompanied by mafic–felsic bimodal volcanism (Nazarevich et al., 1986; Nikishin et al., 1998a, 2001). Sedimentation in the Kayasula Basin was carbonate dominated. The high, separating the Early Triassic East Manych and Kayasula troughs, was covered by the reef fringed Pre-Kuma carbonate platform. The presence of turbiditic sediments in the Mozdok Basin suggests that this trough formed part of an Early to Middle Triassic passive margin.

In the western Pre-Caucasus-Crimean area, the northern Crimea-Azov and Novo-Fedorovsk troughs contain Scythian(?) to early Norian turbiditic clastics, clays and carbonates (Slavin, 1986; Boiko et al., 1986). In the West-Kuban Basin, Early to Middle Triassic series consist of carbonates and clastics. In general, it appears that palaeowater-depths increased towards the area now occupied by the Great Caucasus (Boiko et al., 1986).

Quantitative subsidence analyses of selected wells from the Pre-Caucasus area and the Crimea show that both areas subsided rapidly during the Early and Middle Triassic (Bolotov, 1996; Nikishin et al., 1996, 2001). This rapid subsidence and the occurrence of Early–Middle Triassic basalts and bimodal volcanics are consistent with a rifted origin of these basins. Evolution of the Triassic rifts on the Scythian Platform has been related to steepening of the Palaeotethys subduction zone and back-arc extension. This caused opening of the oceanic Küre and Svanetia basins and the separation of the Pontides–Transcaucasus domain from the southern margin of the Scythian Platform (Figs. 7 and 8; Nikishin et al., 2001; Stampfli et al., 2001; Ziegler and Stampfli, 2001).

The Dniepr–Donets Basin, which originated as a Late Devonian volcanic rift, resumed its subsidence during the Triassic after it was uplifted during the Middle Permian inversion of the Donbass–Karpinsky zone (Milanovsky, 1996; Nikishin et al., 1996). Triassic continental clastics, attaining thicknesses of up to 500 m in the Dniepr Basin, rest unconformably on truncated Permian and Carboniferous sediments (Lozovsky, 1992). Subsidence analyses show that the Dniepr–Donets Basin experienced a phase of accelerated subsidence during Triassic times (van Wees et al., 1996; Ershov, 1997). Although there is no evidence for syndepositional Triassic extensional faulting, this subsidence phase may be tensional in origin, with Devonian salts providing for strain dissipation in the post-salt series. Moreover, the Triassic rise in sea level (Haq et al., 1988) was presumably a contributing factor.

2.3.2. Peri-Caspian Basin

The Peri-Caspian Basin, located northward adjacent to the Karpinsky Swell (Figs. 7 and 8), is bordered to the East by the Late Palaeozoic Uralian Orogen. Sediments contained in this basin exceed 20 km in thickness and rest in its central parts on a 10-km-thick high velocity layer made up of Devonian or Early Palaeozoic or even Riphean oceanic or highly extended continental crust (Zonenshain et al., 1993; Nikishin et al., 1996; Brunet et al., 1999).

According to the interpretations of Zonenshain et al. (1993) and Nikishin et al. (1996), the Peri-Caspian Basin originated as a Middle–Late Devonian rift that possibly proceeded to crustal separation. Its post-rift evolution began at the onset of the Carboniferous. During Late Carboniferous and Permian times, the Peri-Caspian Basin was incorporated into the flexural foreland basins of the Scythian and Uralian orogens. During the Kungurian, very thick halites infilled this basin. A minor erosional unconformity separates Permian and Triassic strata (Lozovsky, 1992). During the accumulation of up to 2- to 3.5-km-thick Triassic continental and marine clastics, diapirism of the Permian salts provided rapid lateral thickness changes. This renders it difficult to assess the true Triassic tectonic subsidence of the Peri-Caspian Basin. Although there is no evidence supporting a tensional control on its Triassic subsidence, a regional tensional setting is indicated by the evolution of the Scythian Platform and the occurrence of Triassic grabens within the southern Uralian system (Nikishin et al., 1996).

2.3.3. Moscow–Mezen Basin

The Triassic to Early Cretaceous Moscow–Mezen Basin is located in the north-central parts of the East European Craton to the north of Moscow (Fig. 7; Lozovsky, 1992). Throughout this basin, a minor erosional unconformity occurs between Late Permian (Upper Tatarian) and Early Triassic (Vetlugian) strata (Yaroshenko and Lozovsky, 1997; Lozovsky and Esaulova, 1998). This region was generally stable during Phanerozoic times and weak extensional events are evident during the Devonian (Nikishin et al., 1996; Fokin et al., 2001), whereas during Carboniferous–Permian times, the area subsided slowly. Quantitative subsidence analyses demonstrate a Late Permian (mainly Kazanian) subsidence acceleration (Nikishin et al., 1996) that may be related to the build-up of compressional stresses during the last phase of the Uralian orogeny. Termination of this rapid subsidence phase at the Permo–Triassic boundary may reflect post-compressional relaxation of these stresses. Over Devonian rifts, the thickness of Early Triassic strata is greater than in adjacent areas. Latest Permian–Early Triassic strata are dominated by continental clastics (Lozovsky and Esaulova, 1998).

2.3.4. Pechora–Barents Sea Basin

The Pechora–Barents Sea Basin (Fig. 7) is characterized by Late Permian–Triassic deltaic and marine clastic sequences that can exceed thicknesses of 10 km (Johansen et al., 1993; Shipilov and Tarasov, 1998; Gavrilov, 1993). The Phanerozoic history of this huge basin was complex and is still somewhat controversial. Moreover, since Devonian times, the evolution of the Pechora region and the Eastern and Western Barents Sea basins differ in some aspects.

The *Pechora Basin* was affected by rifting during Lochkovian and early Frasnian times. During the Carboniferous, this basin was characterized by post-rift subsidence. During the Permian Uralian orogeny, it was incorporated into a wide flexural foreland basin. At the same time, the Devonian grabens were partly inverted in response to the build-up of foreland compressional stresses (Nikishin et al., 1996; Fokin et al., 2001). At the transition from the Permian to the Triassic, some linear subbasins, characterized by rapid subsidence, formed (Nikishin et al., 1996; Korotaev, 1998). These are the partly inverted Devonian Pechora–Kolva rift, the Korotaikha Basin, located along the margin of the Late Triassic–Early Jurassic Pay–Khoy orogen, and the Kos'yu–Rogovaya trough in the Permian Polar–Uralian foredeep basin. Induan basalts occur in the Korotaikha and Kos'yu–Rogovaya basins. The age of this volcanism is constrained by palynological (Ilyina and Novikov, 1991; Lozovsky and Esaulova, 1998), radiometric (Rb–Sr) and palaeomagnetic data (Andreichev, 1992; Lozovsky and Esaulova, 1998). Basalt flows form two layers with the lower one being up to 12 m thick, and the upper one up to 90 m. Palaeontological data do not exclude the possibility that the first layer is latest Permian in age, whereas the second was extruded during the earliest Triassic (Kalantar, 1980; Lozovsky, 1992; Lozovsky and Esaulova, 1998). In the Korotaikha Basin, Induan–early Olenekian sediments attain thicknesses of up to 2000 m. Results of quantitative subsidence analyses and the observed magmatism indicate that a tensional event affected the area at the Permo–Triassic transition.

There is still considerable uncertainty about the age of crustal consolidation and the Early Palaeozoic evolution of the *East Barents Sea region* (Shipilov and Tarasov, 1998; Johansen et al., 1993; van Veen et al., 1993; Gavrilov, 1993). However, our investigations indicate that the main rifting cycle occurred during the Middle–Late Devonian and that since mid-Permian times the region was incorporated into the flexural foreland basin of the Novaya Zemlya orogen from which huge deltaic complexes prograded into a deep-water trough. Subsidence curves reflect a very rapid subsidence event close to the Permo–Triassic boundary (Nikishin et al., 1996; Korotaev, 1998). However, on seismic profiles, there is no evidence for major contemporaneous extensional faulting (Shipilov and Tarasov, 1998; Johansen et al., 1993). In this respect, it should be kept in mind that during the terminal Triassic–Early Jurassic final orogenic pulse of Novaya Zemlya, its foreland basin was compressionally overprinted. Therefore, it cannot be excluded that, similar to the Pechora area, the terminal Permian–Early Triassic subsidence of the Novaya Zemlya foreland basin was enhanced by a short extensional pulse. This is compatible with the observation that Early Triassic clastic sediments reach thicknesses of up to 3 km (Stupakova, 1996), with minor indications of fault reactivation on reflection-seismic data (Shipilov and Tarasov, 1998; Stupakova, 1996), and with the occurrence of Induan basalts in the Gusinaya Zemlya Peninsula of Novaya Zemlya (Kogaro et al., 1992) and possibly also in the eastern parts of the East Barents Sea Basin (Shipilov and Tarasov, 1998).

The *West Barents Sea region* is largely underlain by crust that was consolidated during the Caledonian orogeny (Fig. 1). Devonian collapse of the Caledonian orogen was accompanied by wrench tectonics accommodating a major sinistral displacement of Baltica relative to Laurentia–Greenland (Ziegler, 1988, 1989, 1990). A main rifting phase affected the West Barents Sea area during the Carboniferous (Johansen et al., 1993; Cecchi, 1993; Shipilov and Tarasov, 1998). Late Permian–Triassic marine and deltaic clastic sediments are widely distributed in the western Barents Sea (Nottvedt et al., 1993; van Veen et al., 1993). Subsidence curves reflect a rapid subsidence event at the Permian–Triassic boundary (Bergan and Knarud, 1993; Reemst, 1995). This was coupled with a first uplifting phase of the Loppa Ridge that forms the southeastern shoulder of the Bjørnøya graben. Generally, this event is related to a contemporaneous acceleration of rifting activity in the Norwegian–Greenland Sea area (Ziegler, 1988, 1990). In the Western Barents Sea region, there is no evidence for Permo–Triassic volcanic activity.

2.4. Atlantic, West European and West Tethyan Regions

Large parts of Western Europe and the North and Central Atlantic domain are underlain by basement that was consolidated during the Palaeozoic. The Caledonian Arctic–North Atlantic orogen forms the suture along which the Laurentia and Baltica cratons were welded together during the Late Silurian, thus forming Laurussia. Suturing of Gondwana to Laurussia became effective in the Variscan domain of Europe by late Westphalian times and in the Central Atlantic Appalachian domain by mid-Permian times (Ziegler, 1989, 1990).

2.4.1. Collapse of the Variscan Orogen

Following the Late Devonian–Early Carboniferous initial collision of the northwestern margin of Africa with Iberia, the collision front between Gondwana and Laurussia propagated eastward and southwestward during the Carboniferous main phases of the Variscan orogeny in conjunction with progressive closure of the Palaeotethys and the Protoatlantic oceans. By Westphalian time, Gondwana had collided with the American craton whereas to the East Palaeotethys was still open (Stampfli et al., 2001). Correspondingly, the western parts of the Variscan Orogen are inferred to have been characterized by a Himalayan-type setting (continent–continent collision) whereas its eastern parts, which find their prolongation in the Scythian Orogen, remained in an Andean-type setting (continent–ocean collision). By end-Westphalian times, crustal shortening ceased in the western parts of the Variscan orogen, but persisted in the Appalachian–Mauretanides–Ouachita–Marathon Orogen until mid-Permian times, controlling the Alleghanian orogeny. During the latest Carboniferous and Early Permian, convergence of Gondwana and Laurussia changed from an oblique collision to a dextral translation, culminating by Mid-Permian times in a Pangea A2 continent assembly (Ziegler, 1989, 1990; Matte, 1991; Stampfli et al., 2001). Details of this processes and collapse of the Variscan Orogen are discussed in numerous works (Ziegler, 1988, 1989, 1990; Arthaud and Matte, 1977; Coward, 1993; Marx et al., 1995; Plein, 1995; van Wees et al., 2000; Bertotti et al., 1993; Cassinis et al., 1997).

2.4.2. Opening of the Neotethys and Meliata–Maliak oceans

Following the Late Carboniferous–Early Permian subduction of the Palaeotethys sea-floor spreading axis beneath the eastern Variscan and Scythian orogens, increasingly older and mechanically stronger oceanic lithosphere was subducted. Correspondingly, increasingly larger slab-pull forces were exerted on the noncollisional northeastern margin of Gondwana. These largely controlled the development of a system of Late Carboniferous–Early Permian rifts along the northeastern peripheries of Africa and Arabia that culminated in the Mid-Permian detachment of the ribbon-like, composite continental Cimmerian terranes from Gondwana. Continued northward subduction of Palaeotethys beneath the southern margin of Eurasia accounted for the gradual northward migration of these terranes and the Late Permian and Triassic opening of Neotethys (Sengör et al., 1984; Robertson et al., 1996; Stampfli, 1996; Stampfli et al., 1991, 2001; Ziegler and Stampfli, 2001). Moreover, steepening and roll-back of the Palaeotethys subduction zone, which plunged northwards beneath the eastern Variscan and Scythian orogens, is thought to have controlled the Late Permian and Triassic opening of the oceanic Meliata–Maliak (Kozur, 1991; Stampfli et al., 1991), Crimea–Svanetia (Nikishin et al., 2001), Karakaya (Okay and Mostler, 1994) and Küre (Ustaömer and Robertson, 1997) back-arc basins (Stampfli et al., 2001; Ziegler et al., 2001).

2.4.3. Late Permian and Triassic Atlantic and Tethys rifts

Following their end Silurian consolidation, the Arctic–North Atlantic Caledonides were transected by a major axial sinistral shear zone that governed their Devonian collapse and the subsidence of an array of pull-apart basins in which continental Old Red series accumulated. Wrench deformations along this Arctic–North Atlantic mega-shear gave way to crustal extension during the Middle Carboniferous. Intermittent rifting persisted until crustal separation was achieved at the end of the Paleocene. During Permian times, the evolving Norwegian–Greenland Sea rift propagated southward, opening the way for the transgression of the Arctic Zechstein Seas into the Northern and Southern Permian basins of Western and Central Europe. At the Permian–Triassic transition, rifting accelerated and propagated into the North Sea Basin and during the Anisian into the North Atlantic domain to reach during the Carnian the Central Atlantic area (Ziegler, 1988, 1989, 1990). There is no evidence for Late Permian and Early Triassic volcanic activity in the Arctic–North Atlantic rift system.

During the Triassic, the Tethys rift system, which can be related to the opening of the Meliata–Maliak and Neotethys oceans, propagated westwards and interfered with the Norwegian–Greenland Sea rift system in the Northwest African–Iberian–North Atlantic area. In the process of this, Western and Central Europe, including the Variscan domain, were transected during Early Triassic times by a complex, multidirectional graben system. Some elements of it are superimposed on Permo–Carboniferous fractures (Ziegler, 1988, 1990; Stampfli et al., 1991, Stampfli and Merchant, 1997). Similar to the Permo–Triassic Arctic–North Atlantic rift, the Triassic rifts of Western and Central Europe are essentially nonvolcanic. On the contrary, the Late Permian and Triassic Tethyan rifts are characterized by considerable magmatic activity, related to partial melting of the lithospheric thermal boundary layer and upper asthenosphere in response to lithospheric extension and possibly the impingement of a plume (Pamic, 1984; Ziegler, 1988, 1990; Bonin et al., 1987; Bonin, 1989, 1990). In this respect, it should be noted that, on the basis of Nd and Sr isotope ratios, Dixon and Robertson (1999) concluded that the Triassic rift-related basalts of Greece, Western Turkey, and Cyprus bear a mantle-plume signature.

In the following, the main Permo–Triassic rifts of Western and Central Europe are reviewed (Figs. 2 and 7).

In the *Iberian Peninsula*, Late Permian and Triassic crustal extension controlled the subsidence of the Catalonian–Valencia and Central Iberian basins (Salas et al., 2001). Quantitative subsidence analyses show an initial Late Permian rifting phase that was followed by a second one that started at the Permo–Triassic transition and persisted at least into the Middle Triassic. Volcanic activity was restricted to the Middle and Late Triassic (Salas and Casas, 1993; van Wees and Stephenson, 1995; Arche and Lopez-Gomez, 1996; van Wees et al., 1998). Along the southern, Betic margin of Iberia, rifting activity commenced during the Middle Triassic and was associated with magmatic activity during the Late Triassic and Early Jurassic (Vera, 2001). On the other hand, the Pyrenean Basin, which subsided slowly during Late Permian to Middle Triassic times, experienced a major rifting pulse during the Late Triassic and Early Jurassic, punctuated by basaltic magmatism (Brunet, 1994; Curnelle, 1989; Le Vot et al., 1996).

On the *Atlantic shelves of France, Ireland and Great Britain*, an array of Late Permian and Triassic rifted basins is evident, including amongst others the Western Approaches, the Channel, Bristol Channel, North Celtic Sea, Porcupine basins, the Slyne–Erris Trough, Sea of Hebrides, and West Shetland basins (Naylor and Shannon, 1983; Ziegler, 1990; Parnell, 1992). Late Permian rifting is indicated by the southward advance of the Zechstein Sea and its ingression into the North Sea area via the Irish Sea Basin. Rifting activity accelerated during the Triassic, as evidenced by the accumulation of up to 3 to 6 km of continental clastics, containing evaporitic horizons in basins that opened into the Pyrenean rift zone. Only in the Erris Trough is minor evidence for Late Triassic volcanic activity.

In the *North Sea*, rifting activity commenced during the earliest Triassic and persisted intermittently until the Cretaceous. During the Late Permian, the North Sea area was occupied by the western parts of the Southern and Northern Permian basins in which Rotliegend clastics and Zechstein carbonate-evaporite series accumulated under tectonically quiescent conditions (Ziegler, 1988, 1990; Glennie, 1998; Taylor, 1998). During the Triassic, these intracratonic thermal sag basins were disrupted by the Viking, Central, Moray Firth and Horn grabens and the Horda and Norwegian–Danish half-graben (Ziegler, 1990; Fisher and Mudge, 1998). Throughout the North Sea Basin, Early Triassic sediments are represented by continental to lacustrine redbeds. In the main grabens, Triassic sediments attain thicknesses of 2 to 4 km. Despite significant crustal extension, there is only very minor evidence for Middle and Late Triassic magmatic activity (Ziegler, 1990).

The *Teisseyre–Tornquist Zone* marks the boundary between the Phanerozoic crust of Western and Central Europe and the Precambrian crust of the East European craton. This zone was reactivated by dextral wrench-movements during the Permo–Carboniferous tectono-magmatic cycle and propagated northwestward into the Late Precambrian crust of southern Scandinavia (Sorgenfrei zone) where it terminated in the highly magmatic Oslo Graben. The different segments of the Sorgenfrei–Teisseyre–Tornquist zone, which extends from Denmark through Poland and beneath the Eastern Carpathians to Romania, were tensionally reactivated during the Late Permian and/or Triassic, thus forming a continuous rift system (Ziegler, 1988, 1990).

In the *North Danish Basin*, Permo–Carboniferous wrench tectonics disrupted the Cambro-Silurian shelf sequences in the foreland of the Scandinavian and North German-Polish Caledonides. During the Late Permian, the area was peneplaned, subsided thermally, and was transgressed by the Zechstein series. At the transition to the Triassic, rifting activity commenced and persisted intermittently until the Early Cretaceous. Triassic series attain a thickness of up to 4 km in the North Danish Basin. One occurrence of Late Triassic volcanics is known from the Kattegat area (Ziegler, 1990; Mogensen, 1995).

The *Polish Trough* started to subside during the late Early Permian and progressively expanded and propagated towards the Southwest during the Late Permian and Triassic (Ziegler, 1990; Kutek, 2001). Its subsidence was partly governed by the decay of the thermal anomaly that was introduced during the Permo–Carboniferous phase of wrench tectonics and magmatism and partly by the Permo–Triassic onset of crustal extension (van Wees et al., 1998). Evidence for Late Permian and Triassic rifting activity comes from quantitative subsidence analyses (Dadlez et al., 1995) and surface and subsurface geological studies of the southeastern parts of the Polish Trough where Zechstein salts are absent. Intermittent tensional subsidence of the Polish Trough persisted until Albian times (Kutek, 2001). In the Polish Trough, there is no evidence for volcanic activity at the transition from the Permian to the Triassic.

The *North Dobrogea* orogen, which was affected by last compressional deformations during mid-Permian times (Sandulescu et al., 1995), was disrupted by rifting, starting in the latest Permian–earliest Triassic. Magmatic activity commenced during the latest Permian with the extrusion of alkaline volcanics and culminated during the Early–Middle Triassic transition (Spathian to middle Anisian) when E-MORB-type pillow basalts were extruded in the axial parts of this basin (Seghedi, 2001; Stampfli et al., 2001). Rifting activity persisted apparently into the Late Triassic (Sandulescu et al., 1995); a Late Triassic compressional phase cannot be excluded on the basis of available data (Nikishin et al., 1998b; Seghedi, 2001).

3. Stress environment and sea-level changes at the Permo–Triassic boundary

Available stratigraphic data do not permit a precise correlation of the observed tectonic and magmatic events that occurred around the Permo–Triassic boundary in Eurasia. Nevertheless, data for the Tunguska, Taymyr, Pechora, Vilyuy–Verkhoyansk, and Emeishan flood-basalt provinces and volcanism in the rifted West-Siberian basins indicate that, at the transition from the Permian to the Triassic, Eurasia was affected by a major episode of mantle-derived basaltic volcanism. Stratigraphic data show that the duration of this volcanic pulse was not longer than 5–10 million years and probably much shorter. The area that was affected by this magmatism extends from the Urals in the west to the Verkhoyansk Range in the east over a distance of more than 4000 km, and from Taymyr in the north to Transbaikal or Kazakhstan–Middle Asia in the south over more than 3000 km. To this, the Emeishan flood-basalt province of China must be added. Together, the area affected by this volcanic pulse covers about 3% of the Earth's surface.

Structural and stratigraphic data, partly supported by quantitative subsidence analyses, show that a tensional subsidence event occurred at the Permo–Triassic transition in the West European rifted basins (Brunet and Le Pichon, 1982; Brunet, 1997; Ziegler, 1990; van Wees and Stephenson, 1995; van Wees et al., 1998; Dadlez et al., 1995; Reemst, 1995), the Scythian belt (Bolotov, 1996; Nikishin et al., 1998a), in the Pechora–Barents Sea region (Nikishin et al., 1996; Korotaev, 1998), and in West Siberia (Fig. 5). Supporting palaeo-stress analyses are available for Western Europe (Hibsch et al., 1995) and Transbaikal (Delvaux et al., 1995).

The central-eastern part of the East European Craton, which subsided during Carboniferous to Permian times in conjunction with the Uralian orogeny, underwent regional exhumation at the transition from the Permian to the Triassic. This may be related to the relaxation of compressional stresses and the development of a new tensional stress regime, as suggested by the contemporaneous evolution of the Pechora region in which flood-basalts were extruded. Similarly, uplift of the Siberian Craton at the Permo–Triassic transition was associated with the extrusion of thick flood-basalts (Milanovsky, 1996). However, it is difficult to quantify the magnitude of tectonic uplift of these cratonic areas as the end of the Permian coincided with an important low-stand in sea level (Vail et al., 1977; Haq et al., 1988).

It is likely that the major end-Permian, long-term, low-stand in sea-level was punctuated by numerous short-term sea-level fluctuations (Haq et al., 1988; Ross and Ross, 1988; Embry, 1988; Yin and Tong, 1998). However, detailed analyses of sea-level changes during Late Permian–Early Triassic times, and their correlation between different basins, are affected by major uncertainties in biostratigraphic correlations (Kozur, 1998; Yin and Tong, 1998; Lozovsky and Esaulova, 1998). In this respect, it should be kept in mind that there are major uncertainties in the correlation of Late Permian–Early Triassic continental and marine strata, both at regional and global scales, due to the climatic zonation of continental biota (both latitude and elevation controlled) and the distinct provinciality of shallow marine faunas (Kozur, 1998).

After a mid-Tatarian (± 258 Ma) maximum lowstand, sea levels apparently started to rise again during the Late Tatarian (Haq et al., 1988). Detailed data from Southern China (Yin and Tong, 1998) show that the Permo–Triassic boundary strata span less than 1 million years. Marine sediments along the northeastern margin of the Siberian Craton (Kazakov, 1995) show that a gentle sea-level rise occurred during Induan times and was followed by a sea-level highstand during the early Olenekian. On the East-European craton, a regional gap in sedimentation at the Permo–Triassic boundary (Lozovsky and Esaulova, 1998) was followed by the gentle Induan transgression and the Olenekian sea-level rise. Therefore, it appears plausible that sea levels were lowest during Tatarian times, started to rise at the Permo–Triassic transition or during the Induan, and reached a relative high-stand during the Olenekian (Fig. 9).

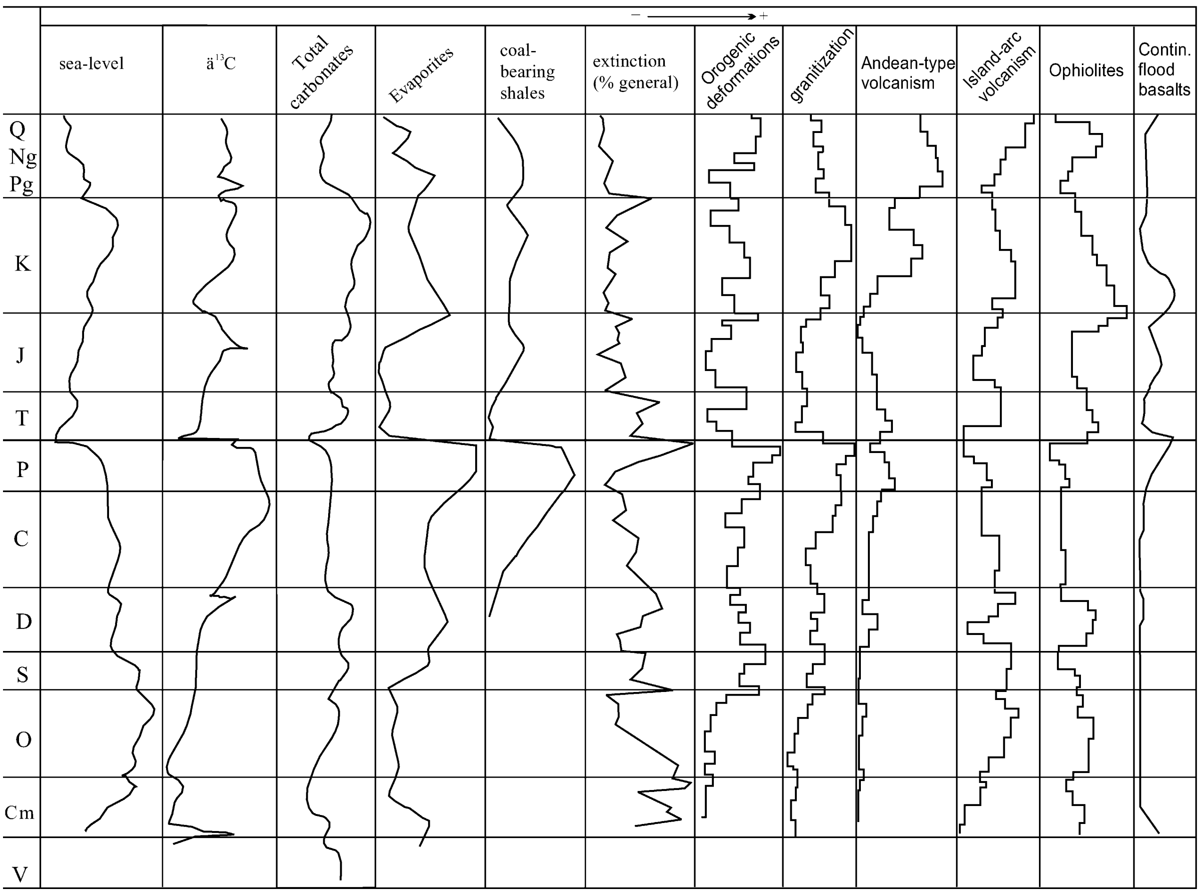


Fig. 9. Comparison of Phanerozoic tectonic, magmatic, environmental, biotic events, compiled from literature. Palaeoenvironment from Walliser (1995), palaeotectonics from Khain and Seslavinsky (1994).

It must be realized, however, that rapid short-term apparent changes in relative sea-level can also be controlled by changes in tangential intraplate stresses, causing deflection of the lithosphere (Cloetingh, 1986), as well as by loads exerted on the lithosphere by upwelling mantle convection systems and mantle plumes (Ziegler et al., 1995). Therefore, it remains to be seen whether during a global plate boundary reorganization, such as at the Permo–Triassic transition, stress-induced relative sea-level changes can, on a regional scale, seriously overprint eustatic sea-level changes (Milanovsky et al., 1992).

4. Permo–Triassic plate boundary reorganization

The Late Palaeozoic–Early Mesozoic global plate boundary reorganization, which controlled the onset of the Pangea break-up, was presumably also accompanied by a reorganization of the deep mantle convection systems (Ziegler, 1993; Nikishin and Ziegler, 1999; Ziegler et al., 2001; Ziegler and Stampfli, 2001).

With the Late Carboniferous and Early Permian consolidation of the Variscan and Appalachian–Mauretanides–Ouachita–Marathon orogens, respectively, the Hercynian suturing of Laurussia and Gondwana came to an end (Ziegler, 1989, 1990). However, during the Permian and Late Triassic, orogenic activity persisted in the Uralian system, along which Kazakhstan and Siberia were welded to the eastern margin of Laurussia (Zonenshain et al., 1993; Matte, 1995; Nikishin et al., 1996), as well as along the southern margin of Eurasia that was associated with the Palaeotethys subduction (Sengör et al., 1984; Ricou, 1995). Locking of the Hercynian sutures in the centre of Pangea was accompanied and followed by accelerated orogenic activity along the American, Antarctic, and Australian Panthalassa margins of Pangea (Proto–Cordillera–Gondwana Orogen; St. John, 1986; Ziegler, 1989, 1993; Visser and Praekelt, 1998). This reflects a first-order subduction progradation from the interior of Pangea to its peripheries. This important plate boundary reorganization was accompanied by a counter-clockwise rotation of Pangea, amounting to 20° during the Permian and a further 17° during the Triassic around a pole located in the Gulf of Mexico (Scotese and Golonka, 1993; Ziegler, 1993).

By Late Permian times, Pangea was practically ringed by subduction zones that faced Panthalassa (Fig. 3a,b). Additional subduction zones facing Palaeotethys occurred along the southern margin of Eurasia and the Chinese and Indonesian allochthonous terranes; the latter separated Palaeotethys from Panthalassa (Scotese and McKerrow, 1990; Scotese and Golonka, 1993). Already during the Late Carboniferous and Early Permian, and even more during the Late Permian and Triassic, Pangea showed signs of instability. This is evidenced by the evolution of the Norwegian–Greenland Sea rift and the so-called "Neotethys" and "Gondwana" rifts.

The Neotethys rifts preconditioned the mid-Permian separation of the ribbon-like composite Cimmerian terranes from the northeastern margin of Gondwana. Their evolution was largely governed by slab-pull forces related to subduction of Palaeotethys beneath the southern margin of Eurasia, once the Palaeotethys spreading axis had been subducted (Sengör et al., 1984; Robertson and Dixon, 1984; Robertson et al., 1996; Stampfli et al., 2001; Ziegler and Stampfli, 2001; Ziegler et al., 2001).

The Gondwana rift systems developed in the foreland of the Gondwana Orogen that fringed the southwestern and southern margin of Gondwana during Late Carboniferous to Early Triassic times. The Late Carboniferous to Mid-Triassic development of these rift systems can be related to compressional intraplate stresses emanating from the South African–Antarctic segment of the Proto–Cordillera–Gondwana Orogen that remained active until Early to Middle Triassic times (Storey et al., 1992; Guiraud and Bellion, 1995; Visser and Praekelt, 1998; Delvaux, 2001). The subsequent evolution of this rift system, which paved the way for the late Middle Jurassic separation of Western and Eastern Gondwana, was governed by processes controlling the break-up of Pangea (Ziegler et al., 2001).

Apparently, Pangea had an insulating effect on the deep mantle convection systems that were active during its Carboniferous and Early Permian suturing phases, causing the decay of old down-welling cells (Guillou and Jaupart, 1995). Moreover, detachment of subduction slab from the lithosphere, which had been active during the suturing of Pangea and some of which may have reached deep into the mantle (for modern examples see Grand et al., 1997; Bijwaard et al., 1998), caused upwelling of the mantle. Accumulation of large amounts of subducted cool oceanic lithosphere near the core–mantle boundary is thought to have a cooling effect on the outer core, causing changes in its convection pattern and the related geomagnetic field. Such a process probably underlies the development of the Late Carboniferous–Permian Reverse Superchron ("Kiaman Magnetic Interval"; Irving and Parry, 1963; Ogg, 1995; Eide and Torsvik, 1996), which commenced during the Westphalian C (± 310 Ma, Menning, 1995; Menning et al., 1997) and terminated with the lower Tatarian Illawarra reversal (265 Ma, Menning, 1995; Menning and Jin, 1998). During the Late Permian and Triassic, the modern bipolar system of deep mantle convection system is inferred to have gradually developed, one branch of which welled up under the core of Pangea and now lies beneath Africa (Cadek et al., 1995). This is compatible with the lower Tatarian resumption of frequent magnetic field reversals (Permo–Triassic Mixed Superchron, PTMS, Menning, 1995) that 15 million years later was followed by the major Siberian and Emeishan mantle-plume activity at the Permo–Triassic transition (250 Ma; Courtillot et al., 1999). The mantle that welled up and radially flowed out beneath the core of Pangea exerted drag forces on the base of its lithosphere. Constructive interference of these drag forces with plate-boundary forces presumably contributed materially to the Mesozoic break-up of Pangea along its Pan-African, Caledonian, and Hercynian sutures. The break-up of Pangea was punctuated by repeated mantle-plume activity that was not exclusively restricted to its break-up axes, as evidenced, e.g. by the Siberian flood-basalt province (Ziegler, 1990, 1993; Courtillot et al., 1999; Ziegler et al., 2001).

Dynamic processes that governed the break-up of Pangea presumably began to develop during the Late Permian and became focused on Africa during the Triassic and Jurassic. This is indicated by the dispersal of the Gondwana constituents and the radial growth of the African plate during Mesozoic times (Pavoni, 1993) and intermittent intraplate magmatism in Africa that was punctuated by several plume events (Wilson and Guiraud, 1998; Ziegler et al., 2001). The global tectonic environment at the Permo–Triassic transition has been discussed for instance by Veevers (1995), Veevers and Tewari (1995), Scotese and McKerrow (1990) and Ziegler (1990, 1993). It is striking that at the Permo–Triassic transition plume-related volcanism was concentrated within Eurasia. Moreover, only minor contemporaneous magmatism is evident in Gondwana, whereas plume-related flood-basalt extrusion was concentrated during the Jurassic within Gondwana and spread only during the Cretaceous and Tertiary to both the Gondwanan and Laurasian parts of the globe (Wilson and Guiraud, 1998; Courtillot et al., 1999; Ziegler et al., 2001). On the other hand, we note that at the Permo Triassic transition, extensional tectonics prevailed in Eurasia, as well as in large parts of Gondwana, whilst compressional stress regimes dominated the Panthalassian margins of Pangea.

5. Environmental crisis at the Permo–Triassic transition

Numerical analyses indicate that the Permo–Triassic boundary corresponds to a major change in the Phanerozoic biotic evolution of the Earth (Khain and Seslavinsky, 1994; Nikishin, 1994; Dobretsov, 1997b).

The Permo–Triassic boundary coincides with a major mass extinction of terrestrial and marine biota that must be related to fundamental changes in the Earth's environment (Sepkoski, 1990; Erwin, 1993, 1994; Stanley and Yang, 1994; Walliser, 1995; Alekseev, 1998; Kozur, 1998; Yin and Tong, 1998). Such changes are evidenced by the occurrence of important $\delta^{18}O$, $\delta^{13}C$, $\delta^{34}S$ and $^{87}Sr/^{86}Sr$ isotope anomalies that straddle the Permian-Triassic boundary (251 Ma; Faure et al., 1995; Kozur, 1998; Yin and Tong, 1998; Atudorei, 1999). Together, these anomalies reflect a period during which several severe global environmental crises occurred. These may have been caused by the massive extrusion of mantle plume-related trap basalts, for instance in the large Siberian Tunguska Province and the Emeishan Province of southwestern China, during a time span of perhaps less than 1 million years at the Permo–Triassic boundary. This subaerial intracontinental volcanism, combined with a voluminous subduction-related volcanism along the Proto–Cordilleran–Gondwana and Proto–Altaid orogens (Faure et al., 1995), presumably caused an increase in atmospheric CO_2 pressure.

However, this "greenhouse" effect was more than compensated by an aerosol-induced reduction in solar irradiation, thus resulting in a global temperature decrease (Loper et al., 1988; Kozur, 1998). Presumably, a sequence of "volcanic winters" and an overall decrease in global temperatures, combined with acid rain, can be held responsible for the observed mass extinction, a dramatic decrease in coal deposition after a Permian maximum, and important anoxic events in the ocean basins (Faure et al., 1995; Kozur, 1998; Yin and Tong, 1998).

6. Permo–Triassic plume event

Collisional amalgamation of the Pangean supercontinent during Carboniferous–Permian times involved the abandonment of the subduction systems that had governed its assembly. It caused also a global change in plate boundaries and plate kinematics, and the development of new subduction zones that almost completely surrounded Pangea during the Late Permian and Early Triassic (Fig. 3). Detachment from the lithosphere of subducting slabs that were active during the Pangea assembly, and their sinking to the core–mantle boundary, paved the way for a reorganization of the mantle convection system. Slab detachment and interaction of the upwelling asthenosphere with the lithosphere contributed to the destruction of the roots of orogenic belts, their post-orogenic uplift and collapse, as seen in the Variscan orogen (Ziegler and Stampfli, 2001; Wilson, personal communication, 2000) and in the Ural–Mongolian belt (Nikishin and Ziegler, 1999). Steepening of subduction zones was associated with back-arc extension and accounted for some rifting, e.g. in the West-Siberian Basin. During the early reorganization phases of the mantle convection system, relatively short-lived mantle plume activity may account for the extrusion of flood-basalts in unrifted areas (e.g. Tunguska Province) and variations in the level of magmatic activity in evolving rift systems (e.g. nonvolcanic Norwegian–Greenland Sea rift, highly volcanic West-Siberian rift system).

In conjunction with the assembly of Pangea, seafloor spreading rates presumably decreased on a global scale, and large parts of Panthalassa were floored by mature oceanic crust. This resulted during the Late Permian in a volume increase of the oceanic basins and a eustatic lowering of sea-levels and a corresponding regression. At the transition from the Permian to the Early Triassic, this regression was followed by a gradual rise in sea level that coincided with accelerated rifting activity, both in the Eurasia and Gondwana part of Pangea, and major plume-related magmatic activity, triggering a global environmental crisis that caused the greatest mass extinction in Phanerozoic times.

The mantle plumes, inferred to have controlled the development of the geographically widely separated Siberian and Emeishan trap basalt provinces (Fig. 3), must have impinged on the base of the lithosphere during the Late Tatarian, causing its uplift at the Permo–Triassic boundary (251 Ma). These plumes remained active for about 10 million years until the Olenekian. As such, they reflect a discrete pulse of diapiric mantle turnover during the early phases of the global convection system reorganization that ultimately led to the development of the modern bipolar mantle convection system (Cadek et al., 1995). In this context, it is noteworthy that during Early Mesozoic times, plume activity was concentrated on the evolving African and East Pacific upwelling centres. Only during the Early Cretaceous, and particularly from Senonian to Paleocene times onwards, did major plume activity spread again into the Laurasian domain (Coffin and Eldholm, 1992, 1994; Courtillot et al., 1999; Segev, 2000; Ziegler et al., 2001).

7. Discussion on Phanerozoic mantle plumes and mantle dynamics

A literature review suggests that during Phanerozoic times, intracontinental and intraoceanic flood-basalt events have occurred rather frequently with a duration of the magmatic phase in the range of 10 to 20 million years and peak activity of about 1 million years (Table 1).

Intracontinental flood-basalt provinces (or LIPs) were described for example by White and McKenzie (1989), Storey et al. (1992), Coffin and Eldholm (1992, 1994), Courtillot et al. (1999), Nikishin et al. (1996) and Fokin et al. (2001). Intraoceanic flood-basalt provinces have been discussed by Coffin and Eldholm (1992, 1994). Classical examples are the

Table 1
Duration of magmatic activity of intracontinental flood basalt provinces

Flood basalt province	Age
Columbia River Traps, USA	16–14 Ma (Courtillot et al., 1999)
Ethiopian Traps, Africa	45–35 Ma and 19–12 Ma (George et al., 1998), or ± 30 Ma (Courtillot et al., 1999)
Thulean Province, North Atlantic	64–52 Ma, peak 59–60 Ma (White, 1988)
Deccan Traps, India	76–62.5 Ma, peak 67.3 Ma (Sheth, 1999)
Rajmahal Traps, India	117–115 Ma (Kent et al., 1997; Sheth, 1999)
Franz Josef Land, Arctic	± 123–110 (Ziegler, 1988; Milanovsky, 1996)
De Long Traps, Arctic	± 112–119 (Drachev, 1999)
Levant, E-Mediterranean	140–120 Ma (Wilson and Guiraud, 1998)
Parana–Etendeka Traps	137–125 Ma (Wilson, 1992; Coffin and Eldholm, 1994), peak 133–131 Ma (Courtillot et al., 1999)
St. Helena hot spot	145–120 Ma (Wilson, 1992; Wilson and Guiraud, 1998)
Karoo Traps, Africa, and Antarctica	198–173 Ma (Cox, 1992), peak 184 Ma (Courtillot et al., 1999)
Central Atlantic Province	210–196 Ma, peak 202–200 (Wilson, 1997; Marzoli et al., 1999; McHone, 2000)
Siberian Superprovince, Russia	267–245 Ma, peak 251–250 Ma (this paper)
Emeishan Province, China	251–250 Ma (Chung and Jahn, 1995), or ± 258 Ma (Courtillot et al., 1999)
East-European Craton	Mid–Late Devonian (Nikishin et al., 1996; Wilson and Lyashkevich, 1996; Fokin et al., 2001)
Vilyuy basalts, Siberian Craton	Mid–Late Devonian (Gayduk, 1988; Milanovsky, 1996)
Volyn Traps, East-European Craton	Early Vendian (Nikishin et al., 1996; Milanovsky, 1996)

West Pacific Ontong Java, Manihiki, Nauru, East Mariana and Pifagetta oceanic plateaux that developed nearly synchronously at 125–115 Ma. Together, these form a uniquely large intraplate magmatic province (Larson, 1991; Coffin and Eldholm, 1994) that has been linked with a Western-Pacific deep mantle upwelling system (Cadek et al., 1995).

Generally, flood-basalt volcanism is related by these authors to mantle plumes that rise from the deep mantle and impact on the base of the lithosphere. On the other hand, Smith and Lewis (1999) reject the mantle plume concept and relate major flood-basalt provinces to the presence of volatile-rich source (wet spots) in a "marble-cake" upper mantle, resulting from the remixing of subducted lithospheric material into the depleted mantle MORB-source. Regarding their model, it should be noted that seismic mantle tomography images subducted slabs that either flatten out at the base of the upper mantle above the 660-km discontinuity (e.g. Western Mediterranean, Banda arc, Izu-Bonin trench) or penetrate this discontinuity and reach down to the core–mantle boundary (e.g. Central America, Japan, Mongol–Okhotsk: Grand et al., 1997; Bijwaard et al., 1998; van der Voo et al., 1999). Similarly, tomography images plume-like thermal anomalies that rise from the core–mantle boundary into the upper mantle (e.g. Iceland, Central Europe, Northeast Atlantic; Bijwaard and Spakman, 1999; Goes et al., 1999; Hoernle et al., 1995). In the light of this and the geotectonic setting of major flood-basalt provinces, a combination of the two models may be applicable in which mantle plumes provide the heat source for partial melting of the upper mantle.

According to the duration of magmatism, we recognize "long-lived" and "short-lived" plumes (Table 1). Moreover, we recognize two different mantle dynamic scenarios for plume-related magmatism.

7.1. Long-lived and short-lived mantle plumes

Classical long-lived plumes are the Rajmahal–Kerguelen, Deccan, Tristan da Cunha, St. Helena, Iceland, and Hawaiian plumes (Coffin and Eldholm, 1992, 1994; Storey et al., 1992). The Rajmahal–Kerguelen plume caused the eruption of the Rajmahal Traps in India close to 117 Ma, the Kerguelen–Broken Ridge oceanic plateau (close to 88–114 Ma), then the Ninetyeast Ridge (close to 82–38 Ma), and new magmatism within the Kerguelen Plateau (± 40–0 Ma; Kent et al., 1997). Correspondingly, its plume-related magmatism lasted for 117 million years. The Deccan plume generated the Deccan–Seychelles Traps at 65 Ma, then the Maldives–Chagos and Mascarene oceanic ridges, and finally the Reunion volcanic area. The duration of this magmatic activity spans 65 million years and may be related rather to fracture propagation than to the drift of the lithosphere over a stationary plume (Sheth, 1999). The Tristan da Cunha plume caused the eruption of the Parana–Etendeka flood-basalts (137–125 Ma), thereafter the intraoceanic Walvis Ridge and Rio Grande Ridge, and finally the Tristan da Cunha oceanic island (Wilson, 1992). This plume was active during the last 137 million years. The St. Helena plume lit up at about 145 Ma, controlled the development of the Helena sea-mount chain and is at present still active (Wilson, 1992). The Iceland plume was responsible for the development of the North Atlantic Thulean igneous province (64–52 Ma), and, after crustal separation between Europe and Greenland, for the development of the Faeroe–Greenland Ridge and the present volcanic activity of Iceland (Ziegler, 1988, 1990; Larsen et al., 1999). Thus, this plume was active during the last 64 million years. The Hawaiian plume was active during the last 75 million years or longer (Coffin and Eldholm, 1994).

To a certain extent, long-lived plumes can be laterally entrained by the convective flow of the lower mantle and the asthenosphere and by drag forces of the lithosphere, as indicated by seismic tomography and/or by the shift of their eruption centres during the opening of oceanic basins (e.g. Iceland plume: Bijwaard et al., 1998; Bijwaard and Spakman, 1999; Central European plume: Goes et al., 1999; Tristan da Cunha plume: Wilson, 1992). Similarly, the head of the East Atlantic Canary–Madeira plume, which lit up at the transition from the Cretaceous to the Cenozoic, appears to have been entrained laterally by asthenospheric flow into the area of western and central Europe (Hoernle et al., 1995). This casts doubt on the validity of the hot-spot reference frame (Morgan, 1971; Smith and Lewis, 1999).

Classical short-lived plumes are the Permo–Triassic Siberian and Emeishan plumes. Activity of the Siberian plume (or super-plume) gave rise to the development of the huge Eurasian flood-basalt provinces (Tunguska, Taymyr, Kuznetsk, Verkhoyansk–Vilyuy, Pechora) and was associated with the development of the large West Siberian, South Kara Sea, and the Pyasina–Khatanga rift systems. Flood-basalt extrusion reached a peak during 1 million years whereas the duration of rifting and magmatism lasted about 5–10 million years. Development of this super-plume did not culminate in opening of a new ocean basin, nor was it followed by long-lived intraplate volcanism of the Hawaiian type. Similarly, the Emeishan plume was clearly short-lived.

Activity of the immense Central Atlantic plume commenced around 210 Ma and terminated about 14 million years later and, thus, was also short lived; its main magmatic phase spanned the time interval 202–200 Ma (Wilson, 1997; Leitch et al., 1998; Marzoli et al., 1999; McHone, 2000). This magmatic activity was preceded and followed by some 15 million years of nonvolcanic rifting that culminated in crustal separation and the opening of the Central Atlantic, starting around 182 Ma (Ziegler et al., 2001).

The largest igneous province in the West Pacific Ocean, including the Ontong Java, Manihiki, Nauru, East Mariana, and Pifagetta plateaux, developed during about 10 million years, after which plume-related magmatism terminated (Coffin and Eldholm, 1994). Activity of this super-plume was apparently not associated with the activation of a new sea-floor spreading axis.

We conclude that "long-lived" plumes can be active for 100–150 million years, whereas the activity of "short-lived" ones is in the range of 1 to 10 million years. The activity of "long-lived" plumes relies on a continuous supply of hot mantle material through a stable "plume-tail" from a reservoir that is continuously replenished. By contrast, "short-lived" plumes lack a stable "plume tail" and must be related to the diapiric ascent of a discrete mass of hot mantle material from a reservoir that is rather quickly depleted.

7.2. Mantle plume model

The primary reservoir of hot mantle material that can rise diapirically, forming mantle plumes, is thought to be the thermal boundary layer at the core–mantle transition zone (D″ layer: Larson, 1991; Storey et al., 1992) and possibly also the lowermost parts of the lower mantle (Kellogg et al., 1999). When this layer reaches a certain thickness, it apparently becomes unstable and ejects hot material in the form of mantle plumes (Prevot and Perrin, 1992). These plumes are generally inferred to rise diapirically through the lower mantle, entraining some of its material, until they reach a density equilibrium with the surrounding mantle. In this respect, the 660- and 410-km discontinuities and the base of the lithosphere, all of which are associated with major density and viscosity changes (e.g. Dziewonski and Anderson, 1981; Shearer and Flanagan, 1999), are levels at which plume heads, depending on their temperature dominated density, may stop their rise, spread out laterally and be injected into the upper mantle (Yuen et al., 1998; Leitch et al., 1998). This concept is compatible with the tomographic images of the Iceland plume (Bijwaard and Spakman, 1999) and the Central European plume (Goes et al., 1999). Plume heads spreading out laterally either at the 660- or 410-km discontinuity probably act as major heat sources that can trigger the upwelling of a system of secondary plumes. Depending on the temperature of the primary plume material, high-pressure/high-temperature partial melting of the upper mantle may commence near the 660 km discontinuity (Ogawa, 2000) or at least from depths of 400 km upwards, particularly in the presence of "wet spots" (Smith and Lewis, 1999). This may cause the diapiric ascent of large volumes of lower density partial melts through (multiple?) secondary plumes to the base of the lithosphere with which they interact. Partial melts derived from the mantle–lithosphere are then stirred into partial melts derived from the upper mantle, possibly entraining lower mantle and original plume material. The resulting heterogeneous magmas rise through the lithosphere, and are extruded as flood-basalts that can be characterized by a wide range of geochemical and isotopic signatures (Smith and Lewis, 1999).

Flood-basalt provinces related to such plumes can have a radius of 1000 to 2000 km (Ziegler, 1988, 1990; White, 1992; Wilson, 1993a; Sheth, 1999; Marzoli et al., 1999) and are characterized by the extrusion of large volumes of basalts in a relatively short time. Extruded magmas show predominately the geochemical and isotopic signature of the mantle source component, which is distinct from that of an upper mantle source, mid-ocean ridge basalts and the mantle–lithosphere (e.g. Deccan: Sheth, 1999; Rajmahal–Kerguelen: Wilson, 1993a; Kent et al., 1997; Parana–Etendeka: Peate, 1997; Central Atlantic: Wilson, 1997; Marzoli et al., 1999; McHone, 2000).

More vigorous plumes (hotter, greater buoyancy) are likely to rise through the 660-km discontinuity directly to the base of the lithosphere where their heads will spread out. Such plumes that are likely to break quickly through the lithosphere are associated with huge volumes of flood-basalts that bear, at least in part, the signature of Ocean Island Basalts (St. Helena, Hawaii plume: Wilson, 1992, 1993a).

We regard these two models as end-members of a wide spectrum of combinations occurring in the real world (e.g. Iceland plume: Holm et al., 1992). Significantly, long lasting, vigorous plumes are almost exclusively located in relatively young oceanic basins, the opening of which was associated with the extrusion of flood-basalts (e.g. St. Helena and Tristan da Cunha plumes). This indicates that after an initial stage of plume-head spreading at the 660- or 410-km discontinuities, or at the base of the lithosphere, the plume was able to break through the lithosphere at the onset of sea-floor spreading (e.g. Iceland plume: Bijwaard and Spakman, 1999).

Upwelling deep mantle convection cells, as imaged by mantle tomography beneath Africa and the West Pacific (Cadek et al., 1995), possibly exert a decompressional effect on the D″ layer. This might induce more frequent plume activity, as evidenced by the Triassic–Jurassic concentration of plume activity at the perimeters of the African mantle upwelling area. Furthermore, such upwelling cells may facilitate heat transfer to the base of the upper mantle, activating perhaps less vigorous plumes rising from the 660-km discontinuity (e.g. Mesozoic and Cenozoic magmatic record of Africa; Wilson and Guiraud, 1998). Moreover, depending on the thickness of the continental crust and lithosphere, plume-derived melts can be contaminated to various degrees by mantle–lithospheric and crustal material (Arndt et al., 1993; Wilson, 1993a), which further complicates the interpretation of geochemical data.

7.3. Mantle plumes and rifting

Although the majority of continental flood-basalt provinces are temporally and spatially associated with "successful" rifts that culminated in crustal separation and the opening of new oceanic basins (Coffin and

Eldholm, 1992, 1994; Wilson, 1992; Segev, 2000), neither the vast Siberian nor the Columbia River province (Oldow et al., 1989) can be linked with continental break-up mechanisms. Similarly, the well-documented Hawaiian plume (Winterer et al., 1989) is also not associated with rifting. Moreover, we note that in successful rifts, large-scale plume-related magmatism can variably occur either almost at the same time as rifting commences (South Atlantic, Labrador Sea), or after 15 million years of rifting (Central Atlantic), some 70 million years after the onset of crustal extension (Indian Ocean) or as much as 270 million years after intermittent crustal extension (Norwegian–Greenland Sea). Similarly, plume-related magmatism can accompany the crustal separation phase (South Atlantic, Norwegian-Greenland Sea), or terminate 10–15 million years before crustal separation has been achieved (Central Atlantic, Karoo) or even commence after crustal separation has occurred (Canada Basin). Moreover, there are successful rifts that are characterized by a very low level or the total absence of syn-rift magmatic activity (e.g. North Atlantic, West Australian margin, Canada Basin; Ziegler, 1988, 1990, 1993; Ziegler et al., 2001).

Thus, mantle plumes on their own cannot be the primary cause of continental break-up, as advocated by Courtillot et al. (1999), although they contribute materially to weakening of the lithosphere. Therefore, far-field stresses, related to plate boundary forces, the interaction of lithospheric plates and drag forces exerted on the base of the lithosphere by the convecting asthenosphere, are seen as the main forces controlling the development and propagation of rifts (Ziegler, 1993; Ziegler et al., 2001).

The rifting stage of "successful" rifts can last from as little as 6 million years (Algero–Provençal Basin, West Mediterranean) to as much as 280 million years (Norwegian–Greenland Sea; Ziegler, 1994; Ziegler et al., 2001). Therefore, in long-duration rifts, thermal anomalies induced by early phases of lithospheric extension start to decay before rifting activity has terminated. In view of this, the sudden flare-up of flood-basalt volcanism after a protracted period of nonvolcanic rifting (e.g. Central and North Atlantic volcanic provinces) can only be related to the impact of a mantle plume. Nonvolcanic successful rifts appear to occur in areas in which the asthenosphere was characterized by normal potential temperatures (Wilson, 1993b). On the other hand, highly volcanic rifts and nonrifted flood-basalt provinces must tap mantle regions characterized by elevated potential temperatures and/or "wet spots", allowing for the generation of large volumes of partial melts (Wilson, 1993b; Ziegler, 1994). Such thermal anomalies can be provided by mantle plumes and/or by upwelling cells of the deep mantle convection system. Plume heads spreading out at the 660- or 410-km discontinuity probably induce partial melting of the upper mantle, particularly if it contains subducted lithospheric material (Smith and Lewis, 1999). Resulting partial melts will then rise to the base of the lithosphere, spread out laterally and ascend into zones of rift- or otherwise induced lithospheric thinning and cause thermal thinning of the lithosphere. Preexisting and active fractures will allow for the rapid ascent of composite partial melts to the surface and their extrusion as flood-basalts. In view of the examples discussed above, we have serious doubts about the plume-less world model advocated by Sheth (1999) and Smith and Lewis (1999).

7.4. Plate boundary and mantle convection reorganization and mantle plumes

Data presented in this paper suggest that there is a relationship between global plate kinematic reorganizations, either following a major continental collision or associated with the break-up of a mega-continent, the reorganization of the global mantle convection system and the rise of mantle plumes. For example, Caledonian suturing of Baltica and Laurentia was followed by a reorganization of plate boundaries and kinematics and a Mid–Late Devonian cycle of rifting and multi-plume activity that affected the East-European and Siberian cratons (Ziegler, 1989, 1990; Nikishin et al., 1996; Wilson and Lyashkevich, 1996). Similarly, Permo–Carboniferous suturing of Pangea was followed by a major plate boundary reorganization, Permo–Carboniferous wrench faulting and magmatism in Western and Central Europe (Ziegler, 1990) that was accompanied by the impingement of a plume on the lithosphere (Wilson, personal communication, 2000), and somewhat later by Permo–Triassic rifting and the impingement of the Siberian and Emeishan plumes on the Asian lithosphere. Furthermore, also the Mesozoic break-up of Pangea, which involved multiple plate boundary and plate kinematics reorganizations, was punctuated by repeated plume activity, including the late Early Cretaceous super-plume event (Ziegler and Stampfli, 2001; Ziegler et al., 2001).

Storey et al. (1992) discussed the relationship between Early–Mid Triassic abandonment of the Gondwana-orogen-subduction zone and the flare-up of the Early Jurassic Karoo–Ferrar plume. Similarly, we recognize a possible relationship between the Mid–Late Devonian mantle plume activity on the East European Craton, the abandonment of the Caledonian subduction system and changes in the Uralian and Scythian subduction systems (Nikishin et al., 1996). However, we see no relationship between these changes in subduction systems and the Mid–Late Devonian Vilyuy flood-basalt province of Siberia (Gayduk, 1988). On the other hand, the Permo–Triassic Siberian flood-basalt province developed shortly after the Permian Uralian orogeny. At this point, we would like to stress that some flood-basalt provinces are located at distances of 1000 to 2500 km to the rear of previously active subduction systems. For instance, the centre of the Tunguska flood-basalt province was located 2300 km to the rear of the long-lived Mongol–Okhotsk (Stanovoy) subduction zone along a southern margin of the Siberian craton (in recent position), whereas the centre of the Parana–Etendeka and the Karoo flood-basalt province were located some 1700 km to the rear of the subduction zone that flanked the southern margin of Gondwana. Considering down-flow of the mantle at deep-reaching subduction zones, some flood-basalt provinces could perhaps be regarded as some sort of back-arc magmatic provinces. In this respect, we assume that subducted cool lithospheric material that has descended to the core/mantle boundary impedes the ejection of plume material from the D″ layer and the lowermost mantle in that given area, causing the ejection of plume material at some distance from the touch-down zone (Kellogg et al., 1999). Nevertheless, we feel that in general large-scale mantle plumes are features that are independent of subduction systems.

On the other hand, lithospheric slabs associated with long-standing subduction zones, along which old oceanic lithosphere was introduced deep into the mantle (Bijwaard and Spakman, 1998), present barriers to the mantle convection system. Their post-orogenic detachment from the lithosphere probably paves the way for a reorganization of the whole mantle convection system. This is exemplified by the development of the short-lived Siberian super-plume at the Permo–Triassic transition and the closely associated West-Siberian and Kara Sea rift system. The Uralian orogeny climaxed during the Kazanian (272–267 Ma) and was followed by rapid uplift of the orogen by some 2 km (Surkov, 1995). Development of the West-Siberian Permo–Triassic rift system and the onset of flood-basalt extrusion was essentially synchronous with the collapse of the Uralian orogen. Therefore, it appears likely that detachment of the Uralian subduction slab induced mantle upwelling and provided a trap for mantle plume material that had risen to the base of the thick Siberian lithosphere further to the east from where it spread out laterally into the area of the West Siberian Basin. The Siberian flood-basalts show evidence of strong crustal contamination (Arndt et al., 1993). It is unknown whether this plume has risen from the core–mantle boundary and directly impacted on the base of the lithosphere, or whether it had spread out either at the 660- or 410-km discontinuity and triggered a system of secondary plumes (e.g. Tunguska, Taymyr, Kuznetsk, and Verkhoyansk–Vilyuy plumes).

Larson (1991) and Coffin and Eldholm (1994) suggested that plume-related magmatism has occurred irregularly during the Earth's history, and provided evidence for an Early–Middle Cretaceous cycle of super-plume activity (Wilson, 1993a; Segev, 2000). We propose that a super-plume event, affecting simultaneously more than one continent and ocean(?), occurred also at the Permo–Triassic transition. However, the Early–Middle Cretaceous and Permo–Triassic super-plumes may be related to different global mantle dynamic processes. The Permo–Triassic super-plume affected the Asian lithosphere just after the Pangea amalgamation during the early phases of a global mantle flow reorganization and 15 million years after the end of the Late Carboniferous–Permian Reverse Magnetic Superchron. On the other hand, the Early–Middle Cretaceous super-plume cycle was activated during the rapid dispersion of the Pangea constituents (Scotese and Golonka, 1993), accelerating global sea-floor spreading and subduction rates and eustatically rising sea-levels (Engebretson et al., 1992; Coffin and Eldholm, 1994), possibly reflecting an advanced development stage of the modern bipolar mantle convection system (Cadek et al., 1995). This super-plume cycle affected both the Gondwanan and Laurasian part of the globe and was accompanied by the development of the Cretaceous Long Normal Magnetic Polarity Superchron (125–84 Ma; Wilson, 1992, 1993a), which probably reflects a rapid thinning of the D″ layer and heat loss across the core–mantle boundary, thus stopping the reversal processes of the Earth's magnetic field (Larson, 1991; Larson and Olson, 1991). About 41 million years later, the D″ layer apparently began to be replenished as suggested by the resumption and gradually increasing frequency of magnetic reversals at 84 Ma. Between 67 and 64 Ma, the Deccan, the Northeast Atlantic and Iceland plumes impacted on the lithosphere (Ziegler et al., 2001).

In view of the above, the global scenarios of the Permo–Triassic and Mid-Cretaceous super-plumes differ fundamentally. Whether magmatic rocks associated with these two types of super-plumes differ in geochemical and isotopic composition is unknown and requires further research. Nevertheless, we propose that the time immediately following the assembly of a Pangea-type megacontinent and times of advanced Pangea dispersion are the most likely periods of super-plume development. Assembly and break-up of Pangea-type mega-continents has repeatedly occurred during the Archean, Proterozoic, and Phanerozoic history of the Earth (Khain, 1992).

As mantle-plumes are a mechanism for transporting excess heat from the interior of the Earth to its surface, their episodic development probably indicates that at certain times mantle convection, on its own, is not an efficient enough mechanism to provide for the required heat transfer. During such times, the D″ layer (Loper, 1984) thickens and becomes dynamically unstable, allowing for the ejection and diapiric ascent of material from it, forming mantle plumes that entrain lower mantle and eventually asthenospheric material (Loper et al., 1988). Times of D″ layer instability probably correspond to periods of global mantle convection reorganization following the assembly of a Pangea-type mega-continent, as well as to times of dispersal of such a mega-continent and the insertion of new subduction slabs deeply into the mantle. The first scenario probably applies to the Permo–Triassic super-plume event. The second one probably applies to the Early–Middle Cretaceous super-plume event that followed a sharp acceleration of global subduction rates, as well as the terminal Cretaceous–Paleocene plume event, which is associated with a second build-up of subduction rates (Engebretson et al., 1992; Grand et al., 1997). On the other hand, accumulation of large amounts of relatively cool subducted lithosphere at the core–mantle transition zone (Late Carboniferous–Permian), as well as a major mantle turnover (late Early–Middle Cretaceous), presumably had a cooling effect on the outer core, its convection pattern and the frequency of magnetic reversals (Larson, 1991; Larson and Olson, 1991).

8. Conclusions

(1) The Permo-Triassic transition, which coincided with a polyphase maximum low-stand in sea-level, followed by a sea-level rise, was accompanied by the inception of new and/or the reactivation of pre-existing rifts in West-Siberia, Western Europe, along the Tethys Belt and in central Gondwana. This was paralleled by the extrusion of voluminous flood-basalts in Eastern and Western Siberia and China.

(2) The Permo–Triassic flood-basalt event can be related to the ascent of short-lived mantle plumes to the base of the Asian lithosphere 15 million years after termination of the Late Carboniferous–Permian Reverse Superchron. The latter may be related to the accumulation of large amounts of subducted oceanic lithosphere in the core–mantle transition zone during the assembly of Pangea. This flood-basalt event contributed to the development of a major environmental crisis, giving rise to one of the most important biotic mass-extinction events in Phanerozoic times.

(3) The build-up of regional extensional stresses in Eurasia and Gondwana at the Permo–Triassic transition is related to a global reorganization of lithospheric plate boundaries and plate kinematics that followed the end of the Palaeozoic suturing of Pangea and marked the beginning of its disintegration.

(4) This plate reorganization was presumably paralleled by fundamental changes in the mantle convection system leading to the gradual development of the modern bipolar African and West Pacific upwelling system.

(5) During the Permo–Triassic, mantle plume activity was concentrated on the Laurasian part of the globe, whereas during the latest Triassic and Jurassic, mantle plume activity was concentrated on Gondwana. During the Early Cretaceous and Cenozoic, mantle plume activity affected both the Gondwanan and Laurasian parts of the globe. This illustrates that plume activity can affect simultaneously multiple lithospheric plates.

(6) The late Early to Middle Cretaceous superplume event coincided with the onset of the Cretaceous Long Normal Polarity Superchron that may have been induced by a global mantle turnover. Subsequently, plume activity resumed at the Cretaceous–Tertiary boundary, some 20 million years after the Campanian resumption of increasingly frequent magnetic polarity reversals.

(7) Based on the geological record, we recognize short-lived and long-lived plumes with a duration of magmatic activity of some 10–20 million years and 100–150 million years, respectively.

(8) Times immediately following the assembly of a Pangea-type megacontinent, involving the detachment of long-standing subduction slabs from the lithosphere, and times of advanced Pangea dispersion, involving the insertion of new subduction slabs deep into the mantle, are periods of changes in the global mantle convection system and the development of super-plumes.

(9) This suggests that global plate kinematics, driven by mantle convection, have a bearing on the development of mantle plumes and, to a degree, also on the pattern of plume-related magmatism.

(10) Mantle plumes are not the main driving mechanism of rifting, although they contribute to rifting by weakening the lithosphere.

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