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Extensional basins of the former Soviet Union — structure, basin formation mechanisms and subsidence history

L.I. Lobkovsky [a,*], S. Cloetingh [b], A.M. Nikishin [c], Yu. A. Volozh [a], A.C. Lankreijer [b], S.L. Belyakov [d], V.G. Groshev [e], P.A. Fokin [c], E.E. Milanovsky [c], L.A. Pevzner [f], V.I. Gorbachev [f], M.A. Korneev [g]

[a] *Centre for Sedimentary Basin Research, Russian Academy of Natural Sciences, Moscow, Russia*
[b] *Institute of Earth Sciences, Vrije Universiteit, Amsterdam, Netherlands*
[c] *Geological Faculty, Moscow State University, Moscow, Russia*
[d] *Geological Institute, Russian Academy of Sciences, Moscow, Russia*
[e] *SovFransGeo, Moscow, Russia*
[f] *State Scientific Industrial Enterprise on Superdeep Drilling 'NEDRA', Yaroslavl, Russia*
[g] *Institute of Geology and Exploitation of Combustible Fuels, Moscow, Russia*

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Abstract

We review the structure and evolution of a number of Riphean–Phanerozoic rifts and extensional basins within the territory of the former Soviet Union (FSU). Horst-and-graben formation in strong crustal and subcrustal lithosphere layers can explain the multi-trough character of rift systems observed in the Russian platform, the Vilyuy rift, the West Siberian rift system, the Pechora–Kolva rift system and the Laptev Sea rift. Many features in the evolution of these rifted basins are incompatible with predictions of classical stretching models. Basin subsidence often occurs in the absence of any noticeable stretching and over time scales much longer than predicted by models of thermal subsidence. Other observations include a time gap between rifting and the onset of post-rift basin subsidence of tens to hundreds Ma and a correlation in timing of subsidence phases of rifted basins and platforms with opening and closure events of adjacent ocean basins. These observations point to an important role for mechanisms such as eclogite formation within or beneath the lithosphere as well as intraplate compression and stress-induced lithospheric deflection.

Keywords: extensional basins; subsidence modelling; lithospheric tectonics; rheology; stress fields

1. Introduction

In the last decade considerable attention has been focused to the modelling of processes associated with lithosphere extension and rifted basin evolution (e.g., Ziegler, 1992, 1996). Most models attempt to explain the structure and development of continental rift zones and rifted continental margins in terms of lithosphere stretching, caused by either far-field horizontal forces applied to the lithosphere or deep mantle material uplift and its subsequent horizontal flow (Turcotte and Emerman, 1983). The wide range of actually observed kinematic patterns of lithosphere extension has lead to the development

* Corresponding author.

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of a large number of modifications of the uniform stretching or pure shear model (McKenzie, 1978). These models invoke non-uniform or discontinuous depth-dependent stretching (Royden and Keen, 1980; Beaumont et al., 1982), continuous depth-dependent stretching (Rowley and Sahagian, 1986), simple shear (Wernicke, 1985; Davis et al., 1986), linked tectonics (Gibbs, 1987) as well as combinations of the simple and pure shear model (Kusznir et al., 1987; Kusznir and Park, 1987; Reston, 1990; Van Wees et al., 1992; Kusznir and Ziegler, 1992). Although successful in explaining some first order features in the large-scale evolution of basins, models such as the stretching model are facing a number of fundamental shortcomings. Well known are the discrepancies between observed values of extension and thinning of the continental crust (e.g., Moretti and Pinet, 1987; Sibuet et al., 1990) and the occurrence of rift-flank uplifts (Moretti and Froidevaux, 1986; Braun and Beaumont, 1989; Kooi and Cloetingh, 1992; Van der Beek et al., 1994) which are not explained by simple stretching models. A growing body of observations strongly supports an important component of non-thermal subsidence during the post-rift phase of the evolution of rifted margins and intracratonic basins (Stephenson et al., 1989; Ziegler, 1990; Artyushkov and Baer, 1990; Leighton and Kolata, 1990; Cloetingh and Kooi, 1992b).

Analysis of different geophysical and geological data from a number of rifts and sedimentary basins has shown that none of the proposed models can successfully explain all features observed (Sibuet et al., 1990; Cloetingh et al., 1993, 1994). It is likely that some features of the evolution of rift zones and sedimentary basins can be described by a combination of several known mechanisms, as, for example, combination of simple shear in the crust, pure shear in the mantle and depth-dependent necking. On the other hand, it also appears that major aspects of the timing and nature of the subsidence record are governed by other unknown processes (e.g., Ziegler, 1992, 1996).

In the present paper we review the structure and evolution of a number of Riphean–Phanerozoic rifted basins located within the territory of the former Soviet Union (FSU). We focus on the style of lithosphere deformation during rifting and the duration and amount of post-rift subsidence. This study illustrates the need to collect more data from wells, seismic lines and outcrops to constrain the various models (e.g., Roure et al., 1996; Cloetingh and Lobkovsky, 1996).

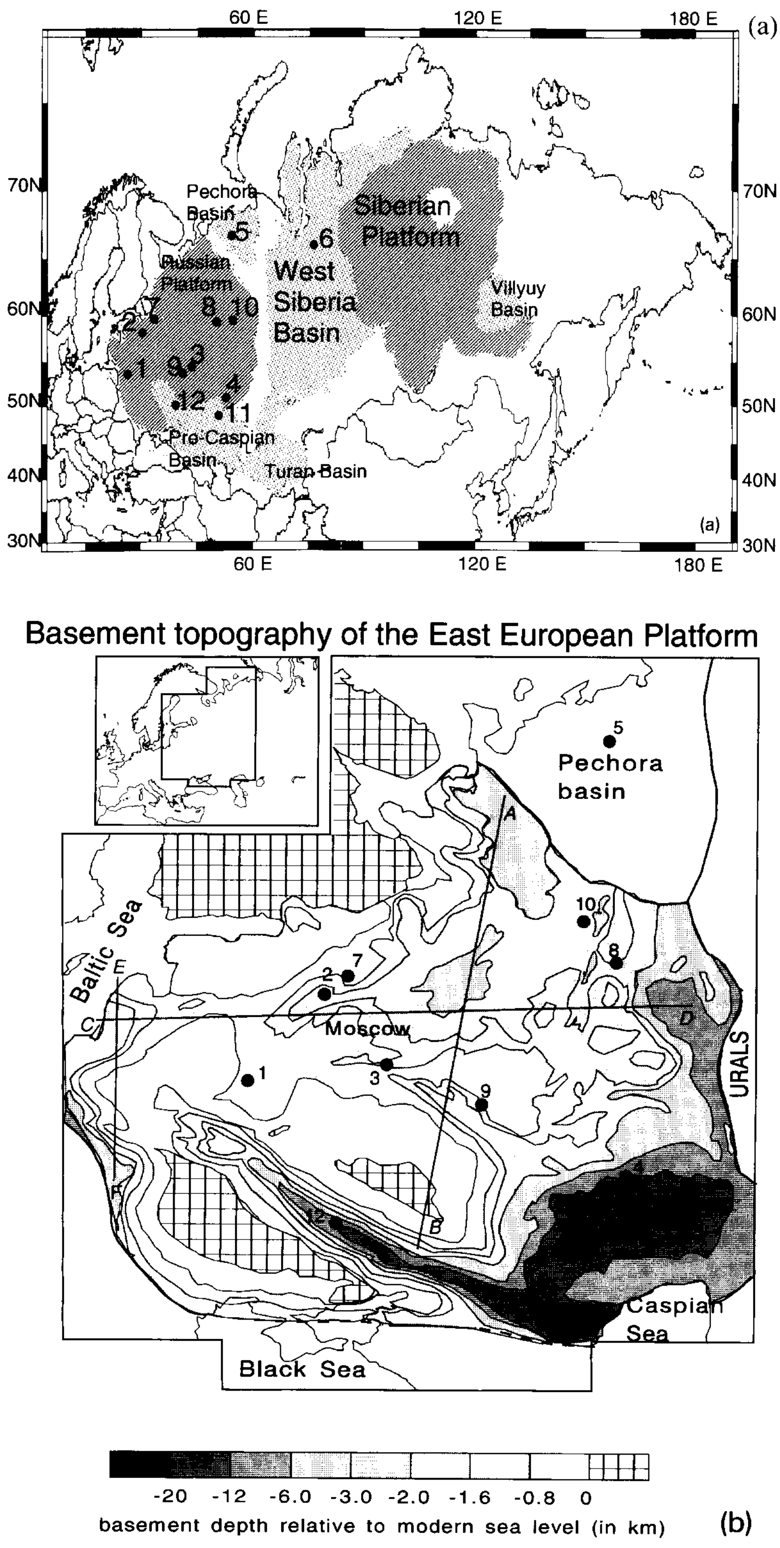
2. Structural style and geometrical constraints on rifting

The territory of the FSU contains a large number of rifted basins of different ages and tectonic settings, each with their own characteristic structure and evolution which are examined here on the base of stratigraphic data from deep and superdeep wells (Fig. 1). Below we consider the structure and geometrical features of the space–time distribution of the FSU rifts (Fig. 2a–e). We also discuss possible kinematic and mechanical models for lithosphere extension for a number of rifts of different ages.

2.1. Structural setting and spacing of subbasins

Tectonic maps of extensional basin formation in different time slices (Fig. 2a–e) demonstrate the multi-trough character of some large rifted basins in the FSU with subparallel rift depressions separated by tectonic uplifts. Examples include the Vilyuy rift zone of East Siberia (Figs. 2c and 5), the West Siberian rift system (Fig. 2d), the Pechora–Kolva rift system (Figs. 2c and 12), the Turan rift system (Fig. 2d), as well as the basins of the Trans-Baikal region (Fig. 2d), and the Laptev Sea rifts (Fig. 2e). The spacing between individual rifts or basins in these multi-trough systems is of the order of several tens up to several hundreds of kilometers. The Laptev Sea rift system is characterized by a spacing of 100 km (Fig. 2e), whereas a spacing of 200 km is

Fig. 1. (a) Map showing the location of individual sedimentary basins and platforms in the FSU investigated in the paper. The numbers refer to the deep wells used for subsidence analyses: *1* = Orsha well; *2* = Valday well; *3* = Pavlovo–Posadskaya well; *4* = Utvinskaya well; *5* = Kolvinskaya well; *6* = Tyumenskaya well; *7* = Pestovskaya well; *8* = Glazovskaya well; *9* = Issinskaya well; *10* = Oparinskaya well; *11* = Central Pre-Caspian well; *12* = Vostochno–Poltavskaya well. (b) Map of basement topography of the East European platform showing the location of deep wells and cross-sections analyzed in this paper (see Figs. 18 and 19).



(a)
60 E
120 E
180 E
70N
60N
50N
40N
30N
Pechora Basin
Russian Platform
West Siberia Basin
Siberian Platform
Villyuy Basin
Pre-Caspian Basin
Turan Basin
Basement topography of the East European Platform
Pechora basin
Baltic Sea
Moscow
URALS
Caspian Sea
Black Sea
-20 -12 -6.0 -3.0 -2.0 -1.6 -0.8 0
basement depth relative to modern sea level (in km)
(b)

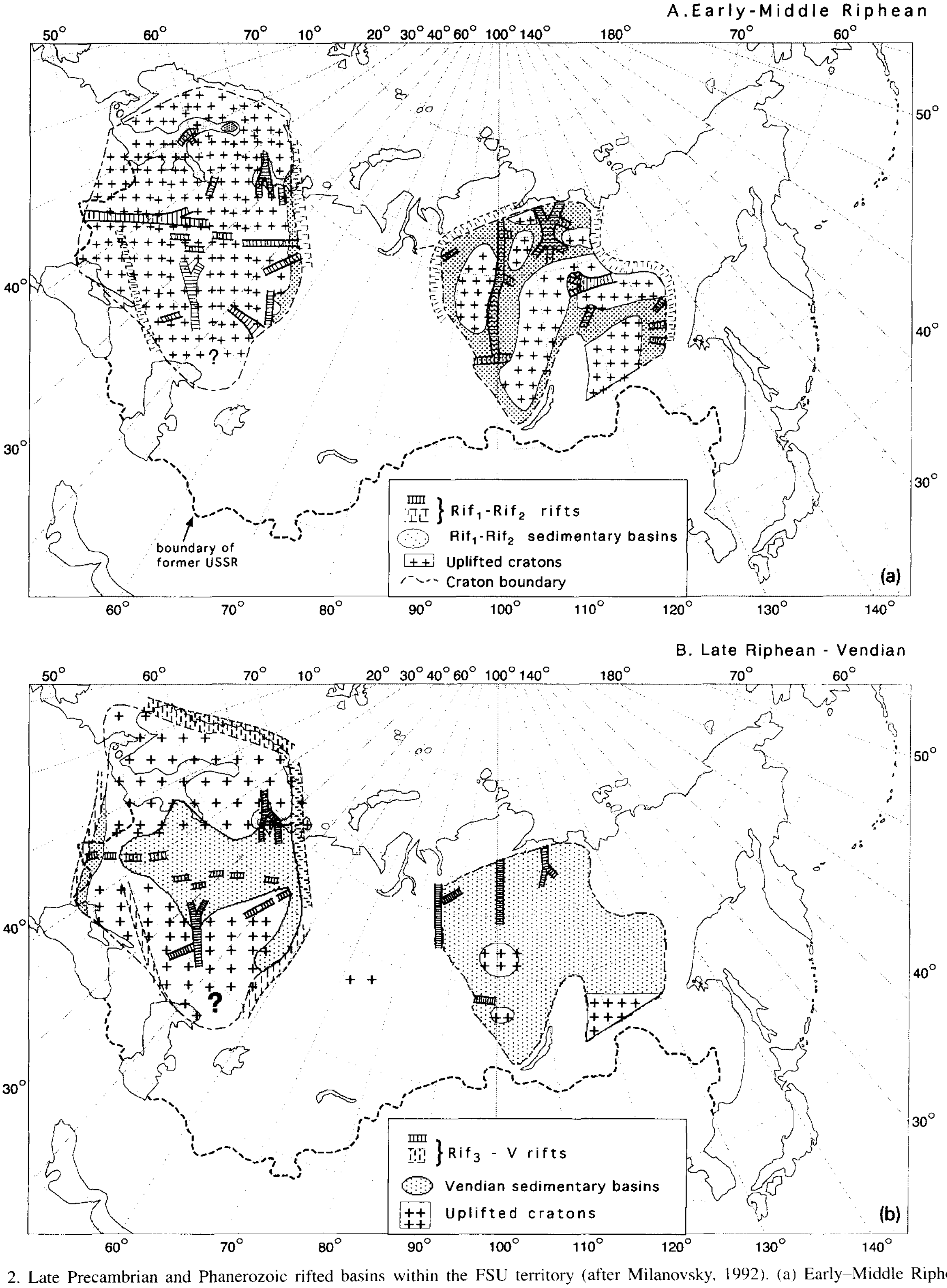
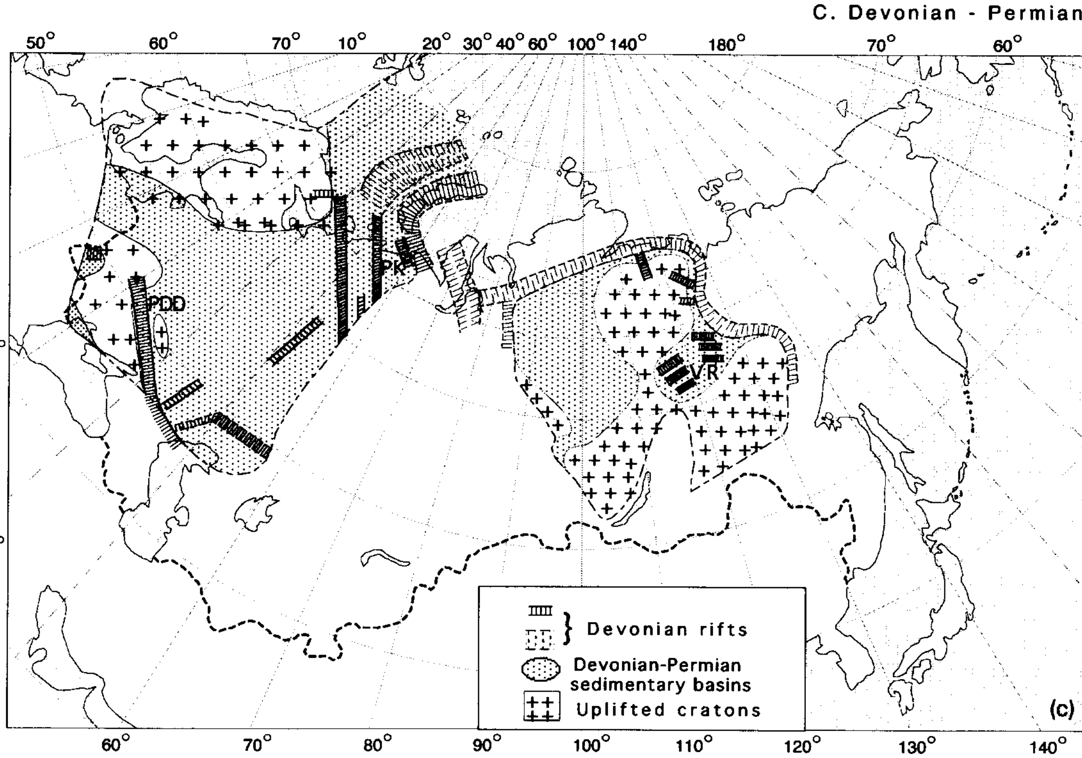


Fig. 2. Late Precambrian and Phanerozoic rifted basins within the FSU territory (after Milanovsky, 1992). (a) Early–Middle Riphean (1.6–1.0 Ga) time slice. (b) Late Riphean–Vendian (1.0–0.57 Ga) time slice.



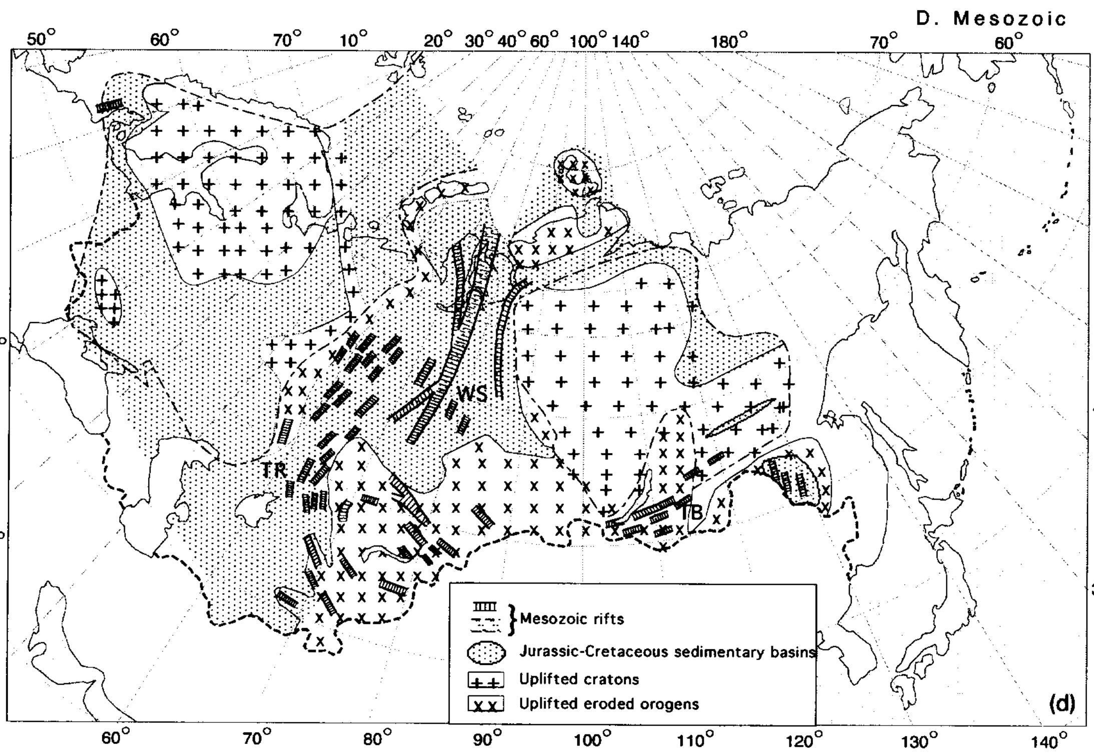


Fig. 2 (continued). (c) Devonian–Permian time slice (*PDD* = Pripyat–Dnieper–Donets rift; *VR* = Vilyuy rift; *PK* = Pechora–Kolva rift system; *TB* = Tungus Basin). (d) Mesozoic time slice (*WS* = West Siberian rift system; *TBS* = Trans-Baikal system; *TR* = Turan system).

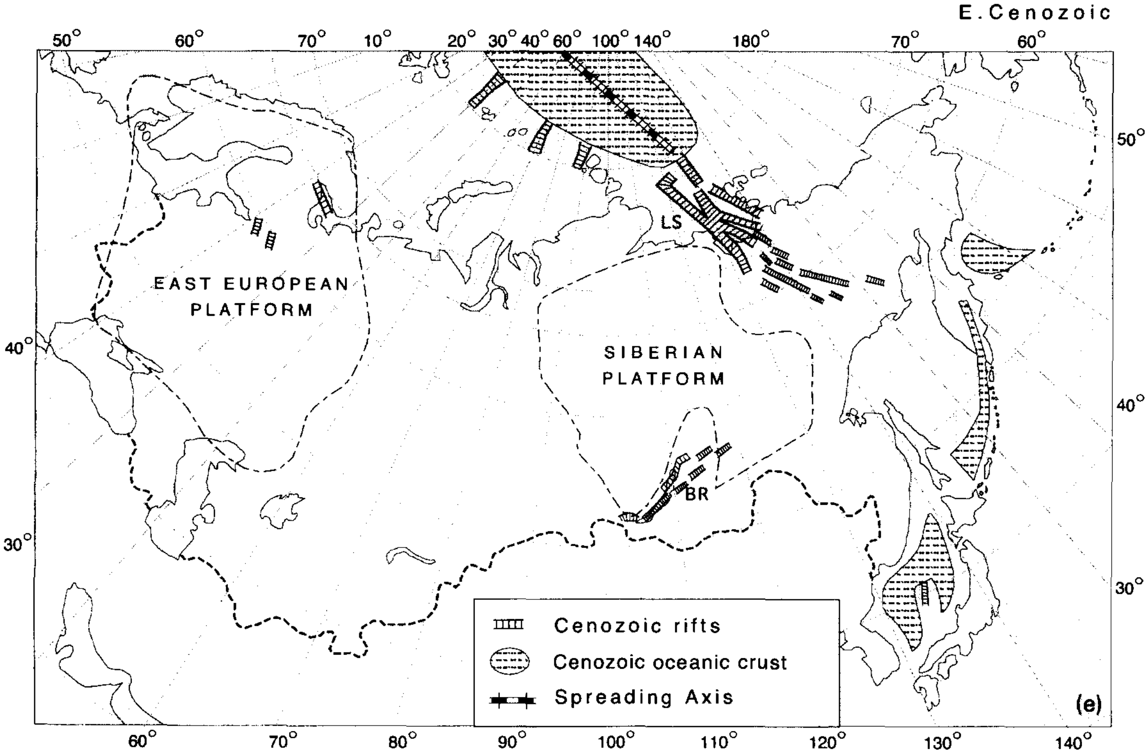


Fig. 2 (continued). (e) Cenozoic time slice (*BR* = Baikal rift; *LS* = Laptev Sea system).

observed for the Vilyuy rift system (Figs. 2c and 5). Similar rift structures, with a spacing of 80 km have been reported for the Celtic Sea Western Approaches area of northwestern Europe (Ziegler, 1990; Sibuet et al., 1990).

The Late Cenozoic Baikal rift zone (Fig. 2e) is a prominent example of a single rift system. This rift extends over a distance of 1500 km from the East Sayan mountains in the southwest to the Kodar range in the northeast (Logatchev and Zorin, 1992; Mats, 1993). The rift consists of a series of halfgrabens with a width of 30–80 km and up to 6–8 km sediments (Sherman, 1992; Hutchinson et al., 1992; Burov et al., 1994). The Baikal lake basin itself is generally considered as a typical halfgraben (Fig. 3a), bounded only on its western side by normal fault. A recent and detailed interpretation of seismic profiles by Kazmin and Golmshtok (1995) indicates that the graben asymmetry is typical for the Late Oligocene–Early Pliocene period, followed by normal fault development also along the eastern side of the lake. The Middle Pliocene period corresponds to marked changes in tectonic regime with an intensification of the rates of vertical movements (Logatchev and Zorin, 1987) and a modification of the stress regime (Delvaux et al., 1996). The changes in tectonic regime in the middle Pliocene also changed the kinematics of graben formation. Since the middle Pliocene, the basins evolve to more symmetrical grabens, with normal faults on both sides.

The Devonian Pripyat–Dnieper–Donets (PDD) rift is located in the southern part of the East European platform (Fig. 2c). A noticeable feature pointed out by many authors (e.g., Milanovsky, 1987; Ljashkevich, 1987; Gavrish, 1989; Garetsky and Klushin, 1989; Chekunov et al., 1992; Stephenson et al., 1993) is its unusually large width (100–150 km). Garetsky and Klushin (1989) have shown that the Western Pripyat basin is bounded by systems of deep listric faults (Fig. 3b). The Pripyat graben with a length of 200 km terminates abruptly in the west. In the east the structure is bordered by the transversal Bragin rise. Extension in the graben does not exceed several percents (Garetsky and Klushin,

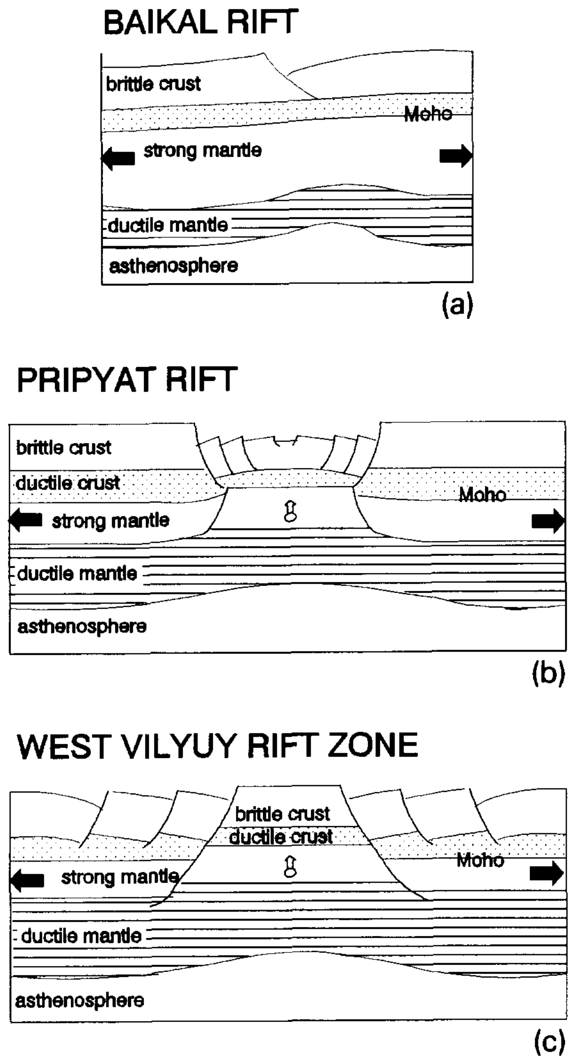


Fig. 3. Three rifting types for thick and intermediate lithosphere. (a) Halfgraben. (b) Symmetrical graben formation. (c) Formation of a central horst block in the lithosphere.

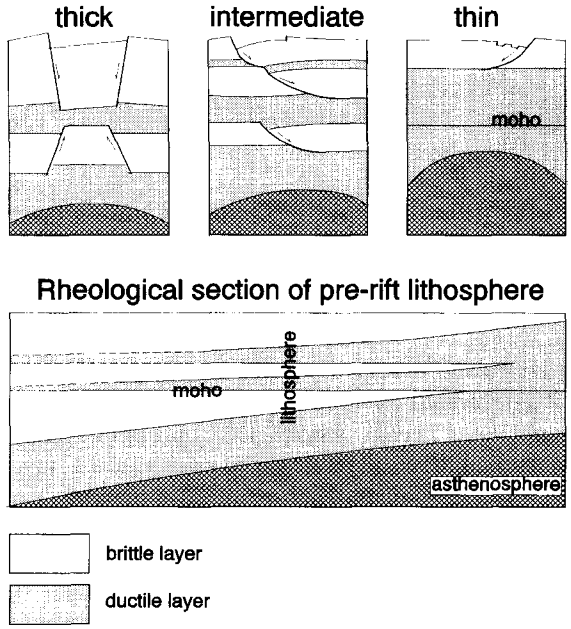


Fig. 4. Variations in rheological stratification for thick, intermediate and thin lithosphere and its response to extension. Modified after Nikishin (1987).

1989). Similar to the Baikal rift, halfgrabens of 40–60 km width form the principal structural building blocks during rifting (Fig. 4).

The Early Devonian–early Frasnian Pechora–Kolva rift system (Belyakov, 1994) consists of several narrow (10–40 km) halfgrabens (see Fig. 12). This system extends from the Urals to the Barents sea area for 1000 km across the Timan–Pechora basin (Fig. 2c). The syn-rift sequence consists of shallow water sediments with basaltic intrusions and volcanic rocks up to 2–3.5 km thick.

The Devonian Vilyuy rift system is located in the eastern part of the Siberian platform (Fig. 2c). The rift system (see Gaiduk, 1988 for a description) consists of several depressions of Frasnian–Tournaisian age, separated by longitudinal and transversal rises. The observed presence of a 500 km wide dike belt indicates that the rift evolved in a regime of roughly symmetrical pure shear pre-rift extension. The 70–90 km wide Suntar horst underwent a syn-rift uplift of about 2 km (Figs. 3c and 6). The size of the horst suggests a whole-lithosphere control (Fig. 3c). The Kempendyai, Ygattin and Sarsan basins are deep asymmetrical halfgrabens strongly resembling the structure of late Cenozoic rifts such as the Baikal, Tanganyika and Malawi rifts (Ebinger, 1989; Burov et al., 1994; Van Wees and Cloetingh, 1994; Van der Beek, 1995). The arcuate shape of their bounding faults is also similar to structures found in the Pripyat (Stephenson et al., 1993), the Baikal (Hutchinson et al., 1992) and the East African rifts (Rosendahl, 1987). The Pripyat rift initially formed on thick (over 100 km) and cold lithosphere with a well pronounced upper crustal layer (Stephenson et al., 1993) and a thinned lower crustal ductile layer, which could have contributed to the formation of whole-crustal faults (Fig. 4).

2.2. Role of bulk thermo-mechanical properties of the lithosphere

The geometry of the FSU rifted basins discussed above obviously reflects a control by a lithospheric rheology characterized by abrupt alternations of brittle–ductile layers (Kirby, 1983; Ranalli and Murphy, 1987) rather than being the expression of long wavelength changes in a viscous asthenosphere rheology (Moretti and Froidevaux, 1986). Bott (1976) applied the wedge subsidence concept of Vening Meinesz (1950) to the brittle upper crustal layer. This model was modified by Lobkovsky (1989) to incorporate the presence of brittle subcrustal lithosphere with compensatory flows occurring within the ductile lower crust and lower lithosphere and the asthenosphere. A model of extension of a broken elastic layer predicts (Bott, 1976) that the width of the horst/graben wedge is between $\pi\alpha/4$ and $\pi\alpha/2$, where α denotes the flexural parameter of the layer; $\alpha^4 = ET_e^3/3g(\rho_a - \rho_c)(1 - \nu^2)$. For an effective elastic thickness (T_e) of the lithosphere of 40 km and adopting values of $E = 10$ GPa for Youngs modulus, $\nu = 0.25$ for Poissons ratio, $g = 10\ \mathrm{m\,s^{-2}}$ for the gravitational acceleration and $0.5\ \mathrm{g\,cm^{-3}}$ for the density contrast ($\rho_a - \rho_c$) between asthenosphere and crust, a width of the horst/graben wedge between 90 and 175 km is predicted by the model (Fig. 4). An effective elastic thickness of 40 km corresponds to a thick brittle subcrustal layer in cold continental lithosphere (e.g., Cloetingh and Banda, 1992; Burov and Diament, 1995). For thinner and warmer lithosphere, the thickness of the brittle subcrustal layer is of the order of 10 km (Cloetingh and Banda, 1992), implying a width of the horst/graben wedge between 60 and 125 km. In extremely thin and hot lithosphere, the brittle subcrustal layer is usually absent, as demonstrated for parts of the Pannonian Basin (Cloetingh et al., 1995; Horvath and Cloetingh, 1996). In this case horst/graben structures in the upper crustal layer are predicted with characteristic widths for the subsiding wedge of the order of 30–60 km (Fig. 4).

The model of horst/graben formation in a subcrustal brittle layer is consistent with the characteristic lateral spacing of rifts and post-rift basins and the observed abrupt change in configuration of rift structures in most of the examples discussed here for the FSU. According to the model, isostatic rise of an upward narrowing subcrustal lithosphere wedge occurs under influence of applied extensional forces, squeezing the viscous material of the lower crustal layer away from the rift axis. This in turn would cause thinning of the crust (neck formation) as well as extension induced and isostatic subsidence of the upper brittle crust (Fig. 4). In this mechanism the material flow that leads to the thinning of the lower ductile crust is not caused by an external tensional force operating on the crust but by the squeezing effect of the rising mantle wedge block (Lobkovsky, 1989; Lobkovsky and Kerchman, 1991). This concept can provide an explanation for observed discrepancies between the amount of upper crust extension as determined by measuring the heave on faults mapped from reflection-seismic data and estimates of stretching factors derived from subsidence analysis and the crustal configuration (Artyushkov and Sobolev, 1982; Ziegler, 1990). The model invokes the interaction of two strong layers in the upper crust and brittle subcrustal lithosphere with ductile layers in the lower crust and lower viscous lithosphere as well as mantle asthenosphere. Furthermore, subsidence of the rift neck due to sedimentary loads and uplift of rift shoulders in response to erosion can lead to flow in the low-viscosity lower crust. This flow, directed outwards from the basin center might facilitate uplift of rift shoulders, affecting predictions inferred from conventional backstripping models (Burov and Cloetingh, 1996).

The model of horst/graben formation within a subcrustal strong lithosphere layer is consistent with mantle reflections observed in deep seismic profiles (e.g., Matthews and BIRPS group, 1987), sometimes interpreted as mantle faults or shear zones, detaching upwards into the base of the crust (Klemperer, 1988; Reston, 1990; Blundell, 1990).

Commonly employed elasto-visco-plastic lithosphere rheologies (Vilotte et al., 1987; Chery et al., 1990) for intermediate P–T regimes in the lithosphere, fail to predict the localization of deformation in subcrustal lithosphere in the form of fault-like narrow shear zones, cutting the lithosphere along, for example, steep angle surfaces. However, such observations can probably be explained if more complex non-associative plastic flow law for mantle rocks are adopted (Nikitin and Ryzhak, 1977; Nikolaevskii, 1983) or when softening effects of the mantle superplasticity are taken into account (Ranalli and Murphy, 1987; Bassi et al., 1993).

The rifts in the FSU, located on lithosphere with different thermo-tectonic ages, show a clear relationship between the style of near-surface tectonics and the bulk rheological properties of the deeper lithosphere (for a general discussion see Cloetingh et al., 1995). The present-day Baikal rift and the Devonian rifts of the East European and Siberian platforms, such as the Vilyuy rift system, are examples of rifting occurring in thick cold lithosphere (see Figs. 3 and 4). An example of a Mesozoic rift system formed on thin hot lithosphere is found southward of the Siberian platform. Here a collisional fold belt was formed in late Palaeozoic and early Mesozoic times, with extensive development of granitoid plutonism. A subsequent phase of Late Jurassic and Early Cretaceous extension in relatively thin and hot lithosphere resulted in the formation of the Trans-Baikal rift system, consisting of a series of 10–30 km wide rift basins (Fig. 2d; Milanovsky, 1989). Recent work (Ernikov, 1994; Delvaux et al., 1995) shows that the late Palaeozoic and Mesozoic tectonic activity of this area is related to the progressive closure of the Mongol–Okhostk ocean.

The first phase (Triassic–Jurassic) took place in an extensional setting, possibly related to the development of metamorphic core complexes, followed by the collision of Mongolia–China with Siberia in the Early Cretaceous, which caused basin inversion under a N–S-oriented compression. Recent field studies, also point to an important component of compression in the evolution of these basins [see Cobbold et al. (1993), Burov et al. (1993) and Nikishin et al. (1993) for a possible analogy in the Tien Shan basins].

In the following we review the post-rift subsidence history of a number of extensional basins in the FSU, focusing on several fundamental questions. (1) Is there any relationship between the style of lithosphere fault formation during the rifting stage and the nature of post-rift subsidence? (2) Is the post-rift history of rift zones formed on cold and hot lithosphere similar? (3) Does the stretching model describe post-rift basin evolution adequately? (4) What is the role of other tectonic mechanisms for post-rift basin subsidence and basin deformation?

In many cases, rift zones formed on thick lithosphere undergo major post-rift subsidence. Rift zones formed on thin lithosphere are frequently subject to post-rift compression and general uplift as observed for large parts of the Trans-Baikalian rift system. This difference can sometimes be explained by the reduced strength of thinned lithosphere, leading to a more pronounced response to regional compression in terms of thrusting and uplift (e.g., Stephenson et al., 1990; Nikishin et al., 1993; Burov et al., 1993). Rifts on thick old lithosphere are in many cases inverted several tens of millions of years after termination of rifting [e.g., Ougarta trough of northern Africa (Ziegler, 1989; Ziegler et al., 1995)].

3. Post-rift subsidence history of some extensional basins in the FSU

Most of the above mentioned rift zones located in the FSU territory are associated with major sedimentary basins. Several superdeep wells (Fig. 1) have been drilled in these basins during the last 20 years. New data obtained from these wells have allowed in some cases to reinterpret existing data leading to better constraints on models for basin structure and evolution.

The results of quantitative subsidence analysis of data from a number of superdeep wells discussed below demonstrate that several main phases of post-rift subsidence can be distinguished during late Precambrian, early Palaeozoic, late Palaeozoic, Mesozoic and Cenozoic times. These periods followed long phases of continental rifting during Riphean–Vendian, Devonian, Triassic and Palaeogene times.

The backstripping procedure (e.g., Bond and Kominz, 1984) removes the effect of sediment loading from the basement subsidence, thus allowing quantification of tectonic basement subsidence. The amount of decompaction is calculated using empirical porosity/depth relations for the specific lithology of each layer (Bond and Kominz, 1984). Local isostatic behaviour of the lithosphere is assumed. This affects the inferred amount of the tectonic subsidence but not the subsidence pattern.

A review of the post-rift subsidence history of a number of extensional basins in the FSU where scientific drilling was performed is given below with a special emphasis on new data for their structure and evolution. The basins discussed can be divided into two major groups: the first group are the intracratonic rift basins, belonging to the East European and East Siberian ancient platforms; the other group consists of basins located on young platforms (West Siberia) or within marginal areas of ancient platforms (Timan–Pechora and Pre-Caspian basins).

4. Intracratonic basins

4.1. Late Precambrian basins of the East European platform

During late Precambrian times (1600–570 Ma) the East European platform underwent four main rifting episodes (Fig. 2a, b): early Riphean (1600–1350 Ma), mid-Riphean (1350–1000 Ma), late Riphean (1000–680 Ma) and early Vendian (680–630 Ma) (Milanovsky, 1987). The post-rift phase of the basins began after the late Vendian (approximately 630 Ma) (Fig. 2b). It appears that the rifting phase is not always immediately followed by a phase of post-rift subsidence (Nikishin et al., 1996). Some of the Riphean rifts were inverted, therefore leading to the absence of a post-rift subsidence (Milanovsky, 1992). Only after the fourth (early Vendian) rifting stage, a phase of post-rift subsidence started. This occurred simultaneously with the opening of the Iapetus Ocean and the Central Asian Ocean, located to the west and to the east of the platform, respectively. The post-rift subsidence was irregular in space and time. Major subsidence occurred in rift zones which died during mid-Riphean times (1000 Ma), whereas subsidence was minor for rifts terminated during the late Riphean–early Vendian (about 680–630 Ma). The post-rift subsidence ceased during the Late Silurian–Early Devonian simultaneously with the Caledonian orogeny in Western Europe. The time history of the Riphean–Vendian sedimentary basins of the East European platform demonstrates that a simple scenario of sedimentary basin formation in terms of stretching followed by post-rift thermal subsidence (McKenzie, 1978) requires substantial modification.

4.2. Late Precambrian basins of the East Siberian platform

At least three rifting phases occurred on the East Siberian platform during Riphean times (Fig. 2a, b): early Riphean (1600–1350 Ma); mid-Riphean (1350–1000 Ma); and late Riphean (1000–800 Ma) (Milanovsky, 1987; Shpunt, 1988). The first two epochs coincide with phases of ocean basin opening in the vicinity of the Yenisyey Mountain Ridge and Trans-Baikalian area; the third rifting phase occurred simultaneously with an early stage of ocean basin opening in the Sayan area. These three rifting stages are probably followed by a phase of post-rift subsidence. However, the Riphean stratigraphy of the East Siberian platform is not worked out well enough to reconstruct the evolution of the rifted basins. Almost the whole East Siberian platform was covered by marine sediments partly overlain by basalts during Vendian–Sturtian times (800–570 Ma) (Milanovsky, 1987). This phase of platform subsidence corresponds to a phase of ocean basin formation in the Altai–Sayan region (Zonenshain et al., 1990). The platform subsidence terminated in the Late Silurian–Early Devonian synchronous with the Late Caledonian orogeny in the Altai–Sayan region (Milanovsky, 1987).

4.3. Devonian–late Palaeozoic basins of the East European platform

Continental rifting became very prominent on the East European platform during the Devonian (Fig. 2c). The Pripyat–Dnieper–Donets rifts as well as a rift belt parallel to the Urals–Nova Zemlya system were formed in this time slice. A number of stages can be separated in the evolution of the Pripyat–Dnieper–Donets rift (Bronguleev, 1981; Milanovsky, 1987; Gavrish, 1989; Garetsky, 1990; Stephenson et al., 1993). During mid-Devonian times a shallow continental Pripyat–Dnieper–Donets rift was formed accumulating continental and shallow-water sediments up to 100–200 m thick, with possible minor volcanic activity. The Pripyat–Dnieper–Donets rift could have been formed by rift propagation from a spreading ridge of the North Caucasian ocean (Palaeotethys ocean) located to the south of the East European platform during Palaeozoic times.

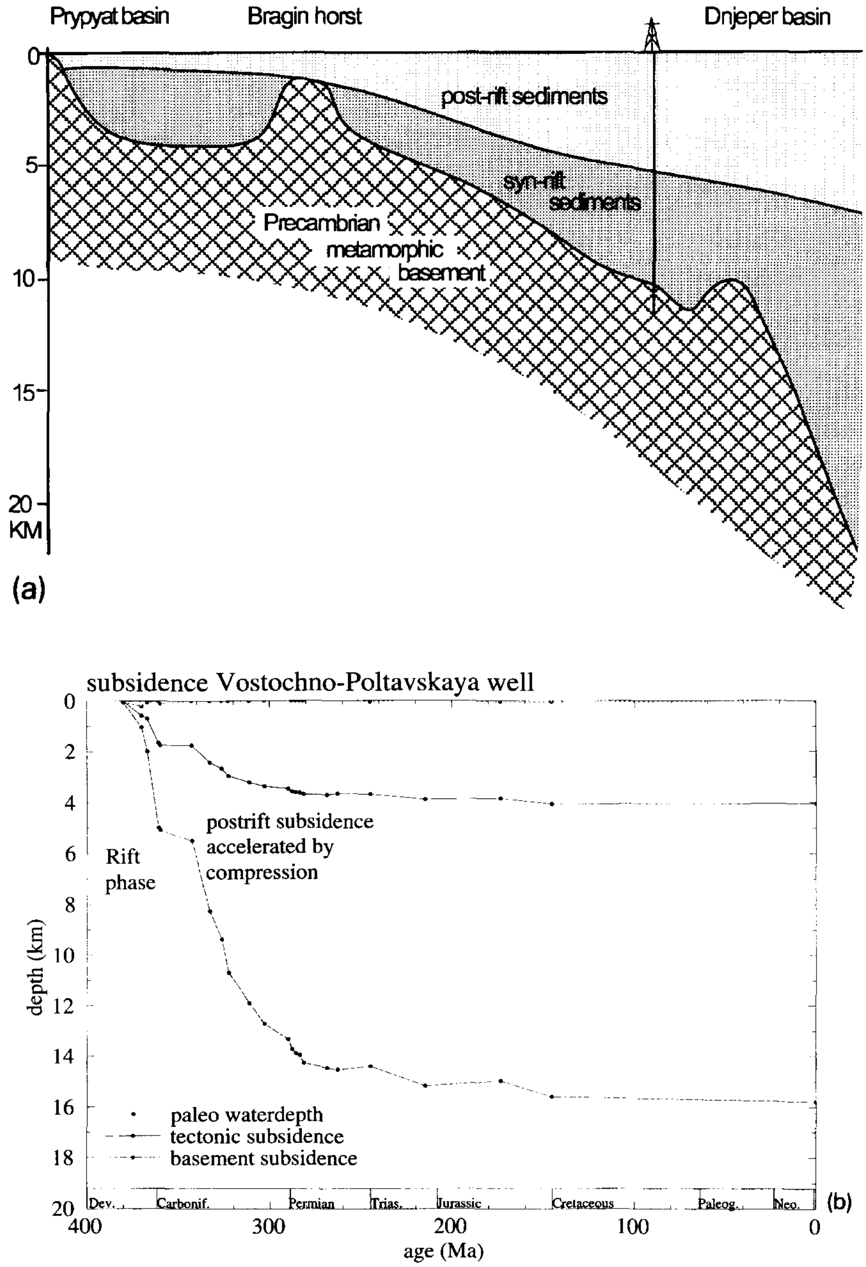


Fig. 5. (a) Cross-section along the Pripyat–Dnieper–Donets rift with the location of the Vostochno–Poltavskaya well. *1* = post-rift sediments; *2* = syn-rift sediments; *3* = Precambrian basement. (b) Subsidence history for the Vostochno–Poltavskaya well (no. 12 in Fig. 1), central part of the Dnieper basin. The time of the main rifting phase was Late Devonian, with a long post-rift subsidence phase during Carboniferous–Mesozoic–Cenozoic times.

The Late Devonian (Frasnian and Famennian stages) was the time of the principal phase of faulting and rifting, with deposition of up to 3–4 km sediments in the Pripyat graben, and up to 3–4 km and more than 6 km in Dnieper and Donets grabens, respectively (Fig. 5a, b). Rifting was associated with several phases of plateau basalt volcanism and doming. The Carboniferous–Early Permian post-rift subsidence was accompanied by the accumulation of shallow-water and continental deposits, with less than 1 km of post-rift subsidence in the Pripyat graben, about 4 km in the Dnieper graben and approximately 12–15 km in the Donets basin (Fig. 5a, b). A major amplification of thermal subsidence by sediment loading can be observed in the Donbass region. As pointed out by Nikishin et al. (1996), the late Visean–Early Permian acceleration of post-rift subsidence is probably controlled by compressional stresses induced by collisional tectonics in the Caucasus–Dobrogea orogen. Mid-Permian inversion in the Donets basin did, however, not affect the Pripyat and Dnieper basins which continued to subside. In Late Permian–Cenozoic times a slow subsidence of the Pripyat and Dnieper basins took place of up to 1.6–1.8 km.

Rifting also affected the eastern part of the East European platform during mid-Late Devonian times (Milanovsky, 1987). It started in the Givetian (Vyatka and Don–Medvedits palaeorifts) and continued in the Late Devonian by the formation of new systems of narrow grabens (Fig. 2c). In contrast to the Dnieper–Donets rift, these rifts do not show evidence of regional doming (Milanovsky, 1987). In the middle Frasnian the Volga–Ural region underwent rapid subsidence of a few hundreds meters with deposition of bituminous shales. In Bashkirian–Permian times, the Volga–Ural area underwent a phase of foreland subsidence and sedimentary infilling. For the area of the Volga–Ural basin associated with rapid subsidence in excess of sedimentation the crustal thickness is about 35 km, whereas a crustal thickness of 40 km has been reported for the surrounding areas of the platform (Bronguleev, 1978, 1981).

4.4. *Devonian–late Palaeozoic/Mesozoic basins of the East Siberian platform*

The Vilyuy basin is the largest middle–late Palaeozoic sedimentary basin of the Siberian platform (Fig. 2c). Devonian rifts underlay the Vilyuy basin (Fig. 6). The presence of three large subparallel rifts is a characteristic feature of the 500 km wide rift system. In the eastern part, the rift system is linked with the Mesozoic Verkhoyansk fold belt (Milanovsky, 1987). The Vilyuy rift belt formed probably during the Devonian the continuation of the ocean spreading zone within the Verkhoyansk–Kolyma region. An extensive geological and geophysical data set, including well data, seismic pro-

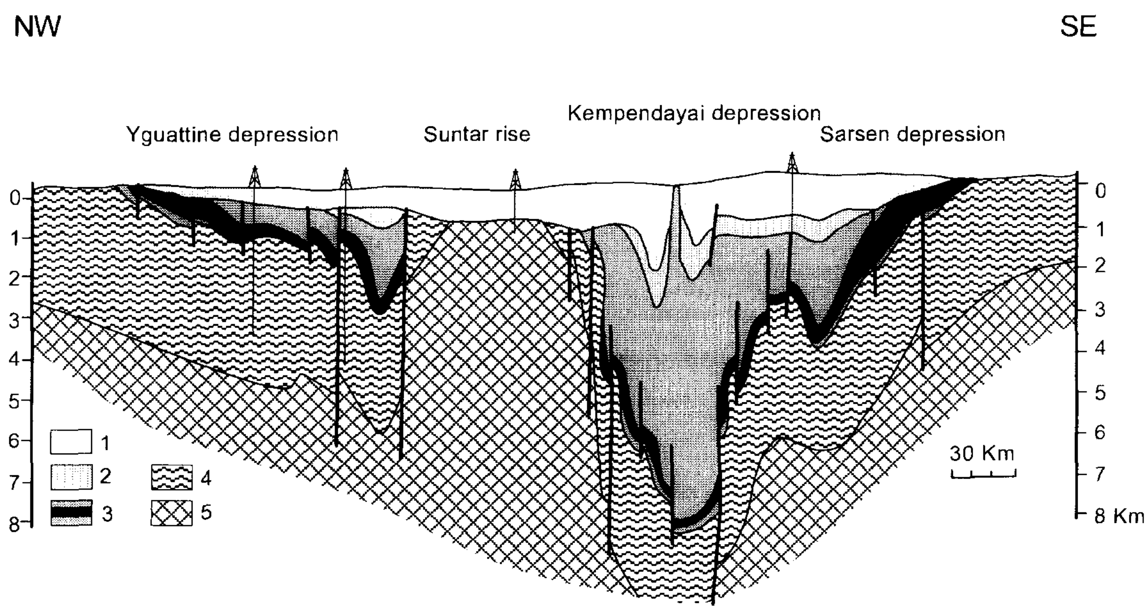


Fig. 6. Cross-section of the Western part of the Vilyuy basin with location of boreholes (Gaiduk, 1988). *1* = Jurassic–Lower Cretaceous post-rift cover; *2* = rift complex; *3* = early Frasnian basalts; *4* = pre-rift sediments (lower Palaeozoic); *5* = Precambrian basement.

files and outcrop data (Gaiduk, 1988), suggests that the following stages can be distinguished in the development of the rift zone: (1) the mid-Devonian (Eifelian and Givetian) formation of a wide sedimentary basin with continental sediments of several hundreds meters thickness; (2) a Late Devonian (Frasnian and Famennian) main phase of rifting, accompanied by extrusion of plateau basalts, flank uplift and syn-rift sediment accumulation up to 2–7 km; (3) a Carboniferous–mid-Jurassic phase, characterized by a post-rift subsidence of less than 1 km and sedimentation over a wide area, with an eastward increasing subsidence towards the passive margin; (4) a Late Jurassic–Cretaceous phase, characterized by the formation of a foreland basin in front of the Verkhoyansk orogen in the eastern part of the Vilyuy basin, with the accumulation of sedimentary thicknesses up to 4–5 km.

5. Basins on young platforms and on the periphery of the cratons

5.1. The Pre-Caspian basin

The Pre-Caspian basin forms an elliptic depression with a diameter of about 600–900 km and a sedimentary thickness of up to 20–22 km (Fig. 1b). During the 1960–1970's two superdeep wells (Aralsorskaya and Biikjalskaya) were drilled and in the 80's another three superdeep wells (Derculskaya, Utvinskaya and Koskulskaya) with a projected depth of 7 km were started in the Pre-Caspian basin for the direct investigation of deep horizons of the sub-salt sedimentary complex and basement (Figs. 1b and 7). Unfortunately, the Utvinskaya and Derculskaya drill holes were terminated at a depth of approximately 4.5 km (Khakheav, 1993). As a result, the structure of sub-salt deposits of inner regions of the basin can only be assessed by seismic data, leading

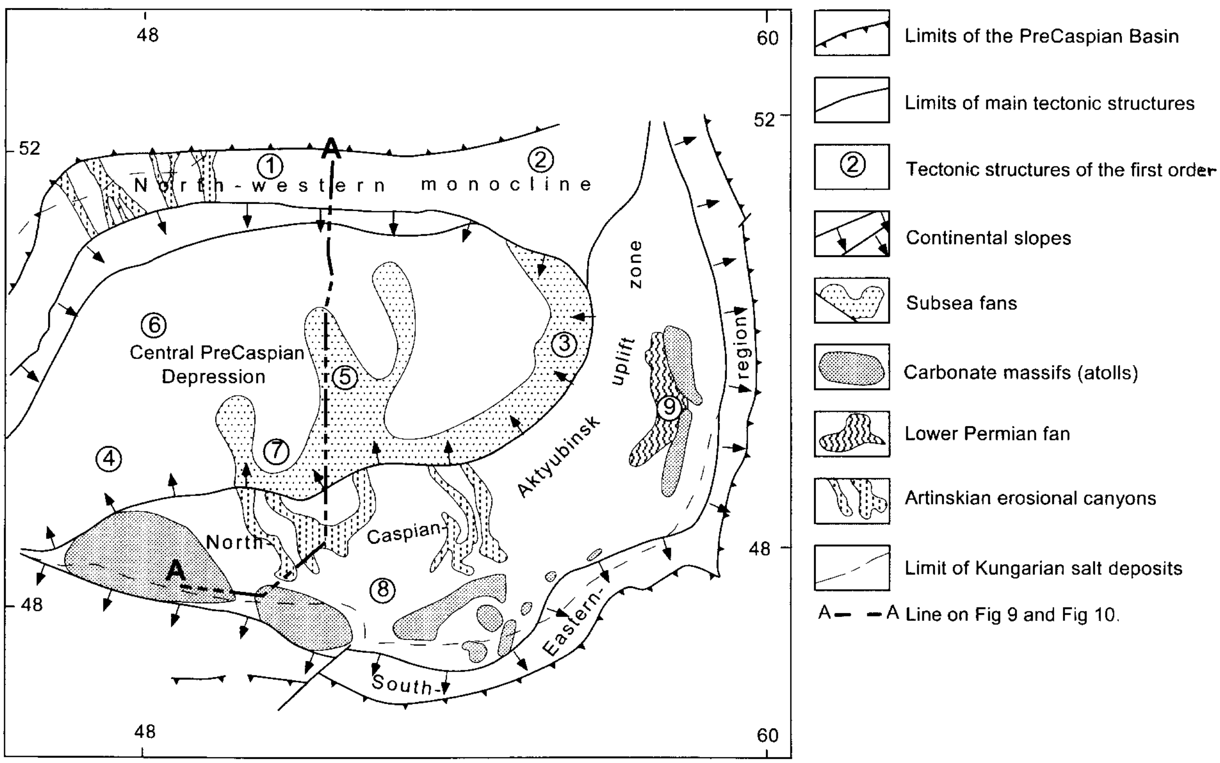


Fig. 7. Geological zonation of Pre-Caspian sub-salt deposits. Tectonic structure: *1* = Pugachov arch; *2* = Sol–Ilek uprise; *3* = Uil uprise; *4* = Sarpin trough *5* = Khobdin trough; *6* = Aralsor arch; *7* = Mezdurechensk step; *8* = East Pre-Caspian arch; *9* = West Pre-Caspian arch.

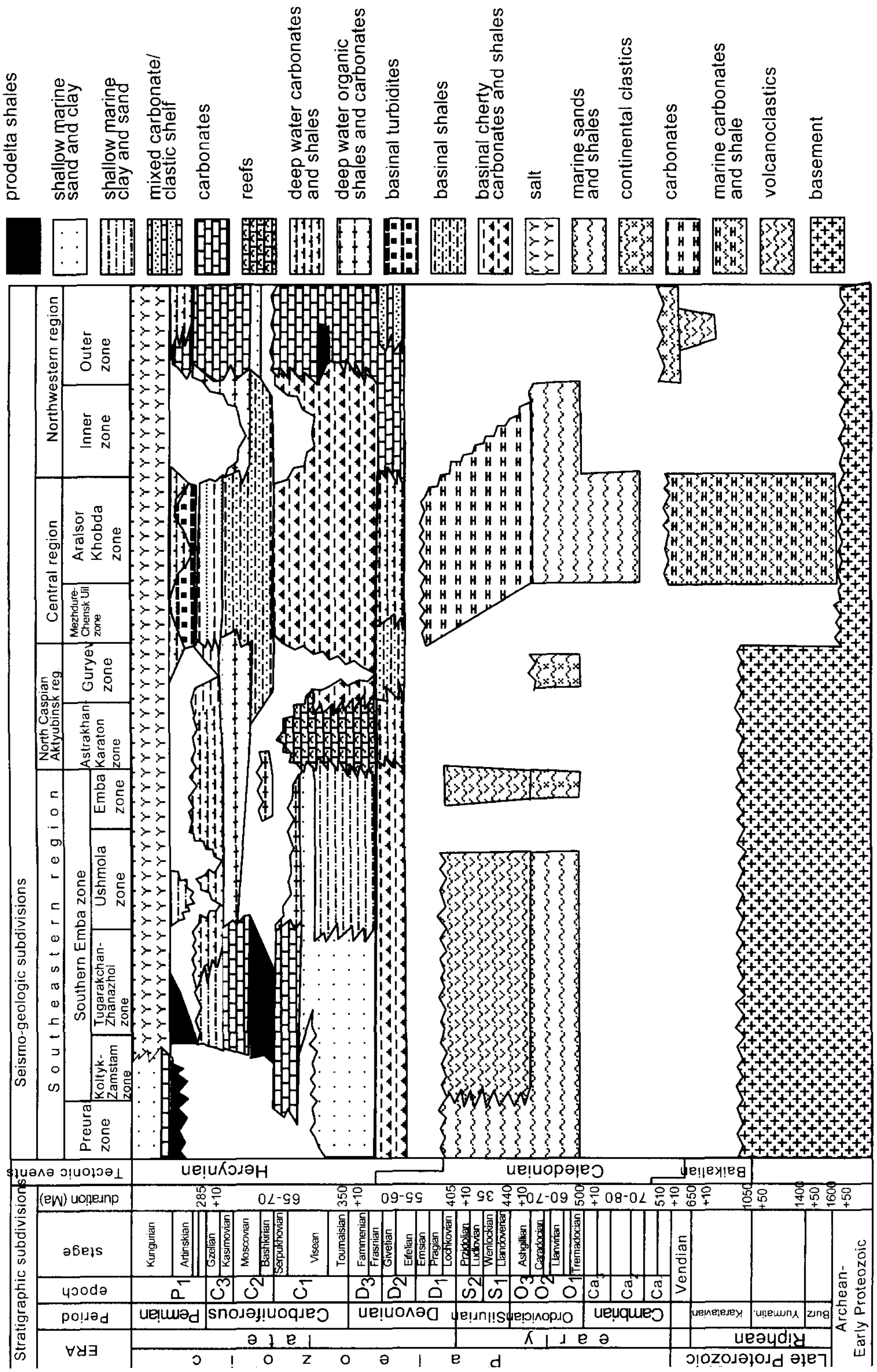


Fig. 8. Chronostratigraphic framework for the Pre-Caspian subsalt deposits (after Volozh, 1991).

to considerable uncertainty on the age and composition of the subsalt strata of the Central parts of the Pre-Caspian depression. Seismostratigraphic analysis (Volozh, 1991) of a network of SDP regional profiles has lead to the construction of a general chronostratigraphic chart of the sub-salt Pre-Caspian complex (Fig. 8). Four seismogeological layers can be recognized in the sub-salt section: Riphean, lower Palaeozoic, Devonian–Lower Carboniferous and Middle Carboniferous–Lower Permian. These divisions of the section are separated by regional stratigraphical gaps of various duration, accompanied by the development of erosional unconformities in marginal areas. Unconformity surfaces correspond to reflecting reference horizons which can be traced through the entire depression (Fig. 8).

Fig. 9 shows a seismogeological section through the Pre-Caspian basin illustrating the relationship of reflection seismically defined sedimentary sequences used for the reconstruction of its late Palaeozoic history (Fig. 10). The following main stages can be distinguished: (1) a phase of late Precambrian rifting; (2) a phase of early Palaeozoic (Cambrian–Ordovician) sediment deposition in a relatively deep water environment, followed by shallow water conditions during Late Ordovician–Early Devonian times; (3) the formation of a deep-water through with water depths of more than 2.5 km in the inner part of the Pre-Caspian basin accompanied by rapid tectonic subsidence during Middle Devonian–Early Carboniferous times; (4) Middle Carboniferous–Early Permian infilling of the deep-water through mainly by clastic sediments; (5) Kungurian deposition of salt; (6) Late Permian–Cenozoic accumulation of shallow water sediments and continental deposits. Subsidence analysis of the Utvinskaya well in the Northern part of the Pre-Caspian basin shows a general subsiding trend from the Middle Triassic onward (Fig. 11a). Subsidence analysis of the Central Pre-Caspian area carried out on the base of well data and seismic profiles shows a general subsidence from Ordovician onward (Fig. 11b).

A number of different interpretations exist of the mechanisms of the origin and evolution of the Pre-Caspian basin (Zonenshain et al., 1990; Volozh, 1991; Artyushkov, 1993). The following observations provide important constraints on the proposed models. As shown in Fig. 11b, the basin underwent four main phases of prolonged rapid subsidence. These occur during Riphean–Vendian, Middle Cambrian–Ordovician, Middle Devonian–Early Carboniferous and Middle Carboniferous–Triassic times. The timing of the Middle Cambrian–Ordovician subsidence phase correlates with the initial stage of the opening of the Urals palaeo-ocean, whereas the subsequent Middle Devonian–Early Carboniferous subsidence phase occurs at the time of back-arc basin formation in the Ural palaeo-ocean. The Middle Carboniferous–Triassic phase coincides with the occurrence of collisional tectonics in the Ural orogenic belt and Karpinsky Swell (Zonenshain et al., 1990; Nikishin et al., 1996). The above suggests that the Pre-Caspian basin was affected by tensional activity in Riphean, early Palaeozoic and Middle Devonian–Early Carboniferous times. Zonenshain et al. (1990) proposed that the Devonian rift phase progressed into the opening of a small, up to 100 km wide, ocean basin. An alternative interpretation attributes the Palaeozoic subsidence to a basalt-eclogite transformation in the lower crust (Artyushkov, 1993), or at the bottom of the thinned lithosphere during rifting (Lobkovsky et al., 1993). The Middle Carboniferous–Triassic phase of rapid subsidence could be the result of compressional tectonics, loading of the Ural and Karpinsky swell orogens and a major supply of clastic sediments (Nikishin et al., 1996). The associated increases in the level of the regional stress field could also have promoted to the effectivity of intra-lithospheric phase changes (Cloetingh and Kooi, 1992a). The rapid subsidence event occurring at the Permian–Triassic boundary is probably related to a regional tensional event.

5.2. *The Timan–Pechora basin*

Fig. 12 shows the main structural units of the Timan–Pechora basin. Within the Pechora basin the Palaeozoic–Mesozoic sedimentary cover has a thickness of 8–10 km (Bronguleev, 1978; Parasina et al., 1989; Daragan-Sukhova, 1991). The basement is Baikalian in age. The Kolvinskaya drill hole with a depth of 7 km is the deepest well in the Timan–Pechora province (Figs. 1 and 12). Terrigenous-carbonate Mesozoic and middle–upper Palaeozoic sediments were penetrated by the well (Khakheav,

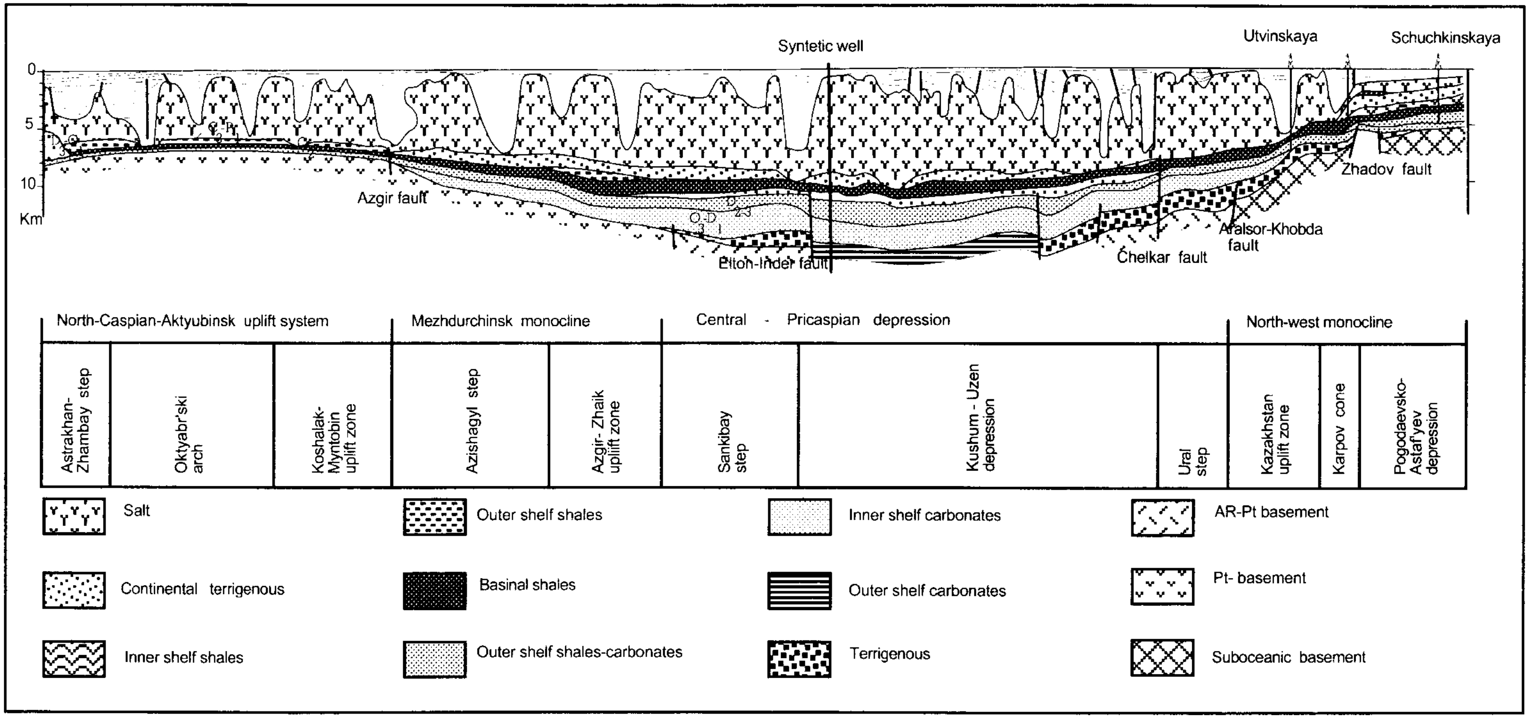


Fig. 9. Regional geologic-geophysical cross-section on the line Zhambay–Uralsk (line AA in Fig. 7) (compiled by Yu.A. Volozh, N.V. Nikolaeva and A.I. Timoshin).

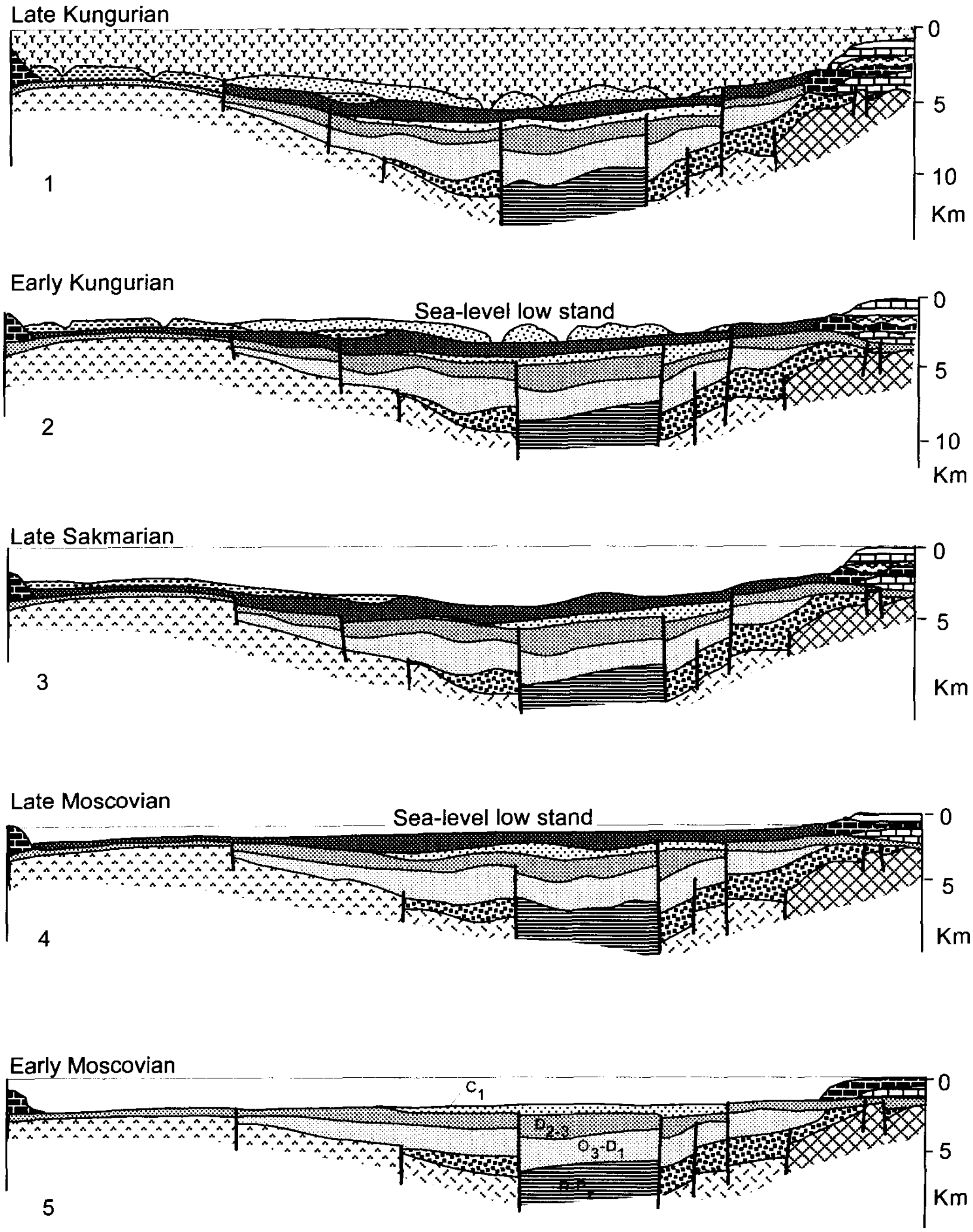


Fig. 10. Palaeoreconstruction of the Palaeozoic history of Pre-Caspian basin on line Zhambay–Uralsk (after Volozh, 1991). Legend corresponds to Fig. 9.

1993). The well reached Silurian sediments, whereas Ordovician deposits are expected to underlie the Silurian strata (Ehlakov et al., 1991a; Belyakov, 1994). Based on a seismic section crossing the Pechora–Kolva basin (Fig. 13) and well data, Belyakov (1994) has compiled the general chronostratigraphic frame-

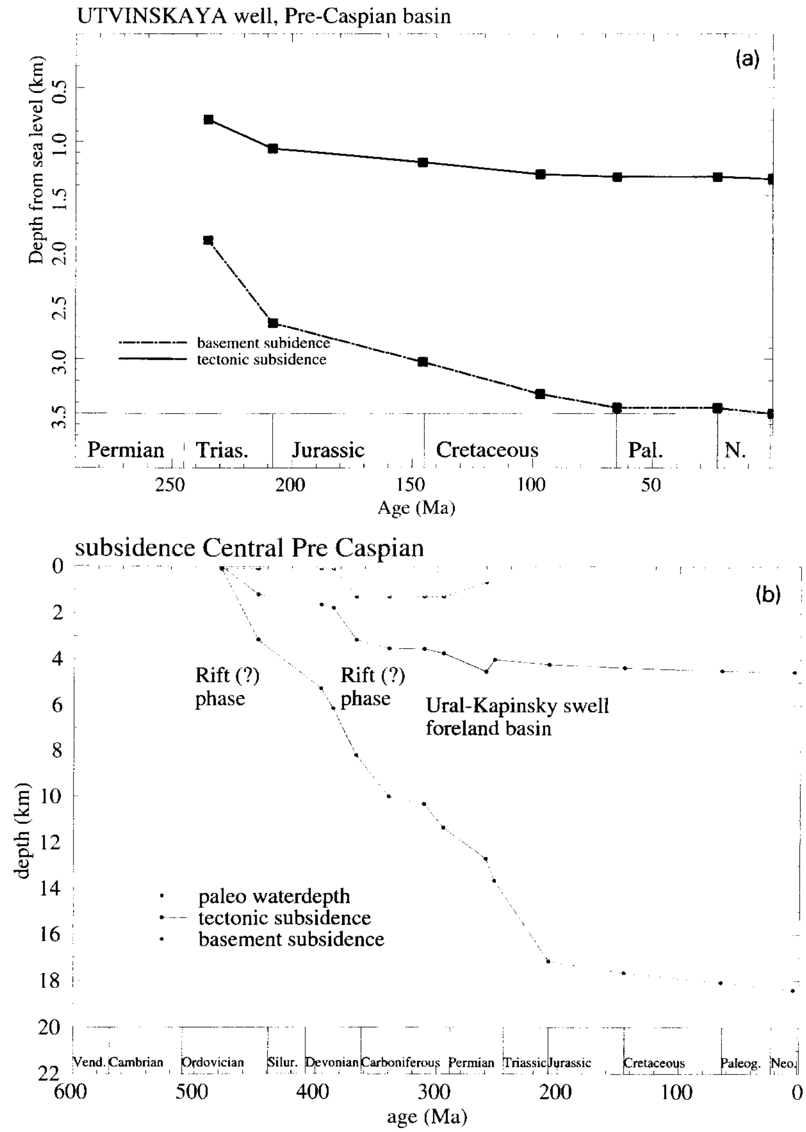


Fig. 11. (a) Subsidence history for the Utvinskaya well (no. 4 in Fig. 1), located in the north flank of the Pre-Caspian basin. Stratigraphy data of Perevozchikov et al., 1991). The well is drilled 3650 m deep and reached the Middle Triassic. Main subsidence phases: Mesozoic–Cenozoic post-rift subsidence; a possible tension phase took place in Triassic times. (b) Subsidence history for the Central part of Pre-Caspian basin (synthetic well, no. 11 in Fig. 1). The timing and nature of the main phases of subsidence: Ordovician — possible rift phase; Silurian–Early Devonian — post-rift subsidence; Middle–Late Devonian — main phase of rapid subsidence, possible rift phase; Carboniferous — post-rift subsidence; Early Permian — possible foreland subsidence in front of Urals; Kungurian-Late Permian — infilling of basin by salt and molasse; Mesozoic–Cenozoic — slow subsidence; *1* = tectonic subsidence; *2* = basement subsidence; *3* = rate of tectonic subsidence; *4* = depth of sedimentation surface.

work of the Timan–Pechora basin (Fig. 14). This has allowed the reconstruction of the evolution of the basin from Early Ordovician to late Artinskian times (Fig. 15) (Belyakov et al., 1994).

Quantitative subsidence analysis of the Kolvinskaya well (Fig. 16) demonstrates the occurrence of major phases of rapid subsidence in Early Devonian (Lockhovian), Middle Devonian and Late Devonian

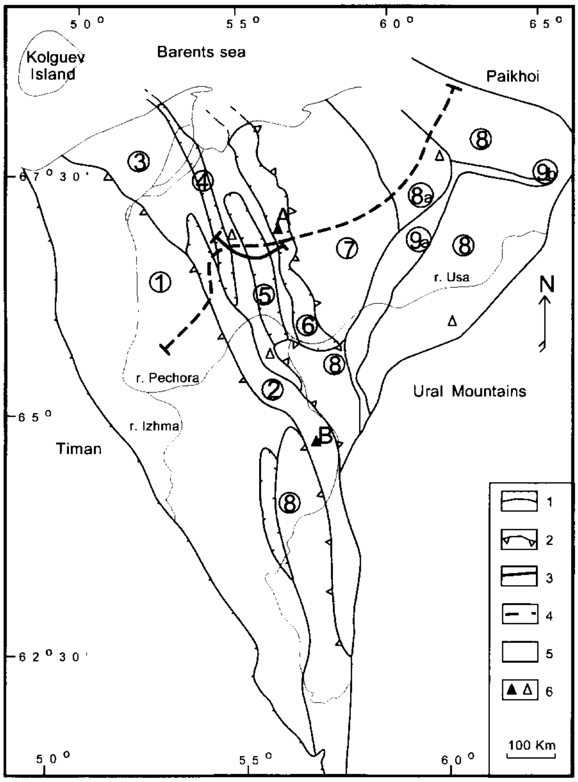


Fig. 12. Major structural units of the Timan–Pechora basin. *1* = boundaries between structural units; *2* = boundary of the Pechora–Kolva aulacogen; *3* = profile line of seismic section (see Fig. 13); *4* = profile line for geological reconstruction (see Fig. 15); *5* = areas of Devonian graben-rifts; *6* = wells, *a* = drilled, *b* = projected, by the Enterprise NEDRA (*I* = Kolvinskaya N1, *II* = Timan–Pechorskaya). Structural units: *1* = Izhma–Pechora depression; *2* = Pechora–Kozhva megaswell; *3* = Malaya Zemlya monocline; *4* = Shapkina–Yuryakhinsk swell; *5* = Laya dome (swell); *6* = Kolva swell; *7* = Khoreiversk depression; *8* = Ural foredeep depression; *8a* = Varandej–Adzvinsk structural zone; *9* = ridges (*a* = Chernyshev, *b* = Chernov).

times related to rifting. These rift phases were separated by local inversions. The phase of rapid post-rift subsidence during middle Frasnian–Famennian (Figs. 15 and 16) seems to be caused partially by a non-thermal mechanism possibly related with eclogite lens formation on the lithosphere–asthenosphere boundary (Lobkovsky et al., 1993).

As a result of Uralian compression, the Devonian grabens within the Timan–Pechora basin were partly inverted during the Carboniferous–Triassic period (Fig. 16) and in Late Triassic–Early Jurassic times some erosion occurred. This compressional deformation also initiated the formation of Lower Permian structural units and controlled the distribution and evolution of the Permian reefs and bioherms (Belyakov, 1994).

5.3. The West Siberian basin

During the Mesozoic, three main rifting phases are prominent in the Urals–Mongolian fold belt (Fig. 2d), mainly active during the Triassic, Early–Middle Jurassic, and Late Jurassic–Early Cretaceous (Milanovsky, 1992). Through time, rifting activity shifted eastward. Post-rift subsidence leading to the formation of wide basins occurred in the West Siberian, Turan and Zeya–Bureya basins (Fig. 2d), but evidence for post-rift subsidence is lacking in several other sub-basins. A number of these basins even underwent inversion due to tectonic compression (Fig. 2d).

The West Siberian basin (Fig. 2d) is the largest sedimentary basin in the territory of the FSU, with a sedimentary cover of up to 13 km thick. The basement of the West Siberian basin consists of Palaeozoic folded strata and Precambrian blocks, with different interpretations proposed for their boundaries (Surkov, 1986; Bogolepov et al., 1988; Milanovsky, 1989; Khain et al., 1991; Peterson and Clarke, 1991). A number of Vendian–Palaeozoic sedimentary basins are located under the Mesozoic cover (Surkov, 1986; Siemov, 1987; Milanovsky, 1987). The history of the West Siberian basin itself started with a phase of Late Permian rifting, following the completion of the last phase of the Uralian Carboniferous–Early Permian orogeny in the Urals–West Siberia region. The entire basin subsided in a back-arc position on top of an orogenically destabilized lithosphere (Ziegler, 1989). In West Siberia the Permo-Triassic rifting was widely manifested (Surkov, 1986), though its scale is still disputed as a result of the lack of deep bore hole evidence.

The N–S-trending Urengoi–Koltogor rift can be traced in the axial part of the basin (Fig. 2d). The graben is 50 km wide. According to seismic data a sedimentary cover thickness of 8 km is overlying crystalline crust with a thickness of 29 km (Surkov et al., 1993). The Tyumenskaya superdeep well is located in the northern part of the Urengoi–

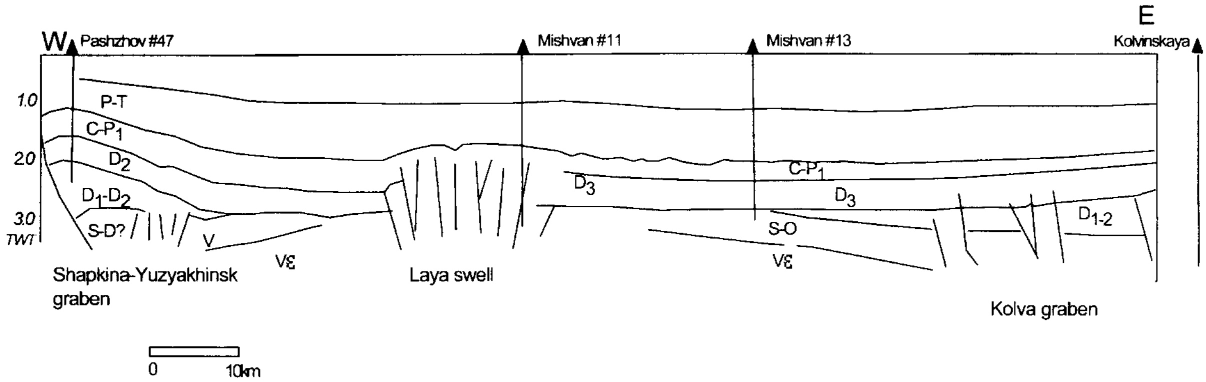


Fig. 13. Line drawing based on interpreted time section across the Pechora–Kolva aulacogen (for location profile see Fig. 12).

Koltogor graben (Figs. 1, 2d and 17a). In 1994 the well reached a depth of 7.5 km, sampling an almost complete Mesozoic section (Fig. 17b), and entered volcanic-sedimentary sequences of latest Permian age at a depth of 7.3 km (Khakheav, 1993). The most lower sequence is represented by a Late Permian–Early Triassic volcanic-sedimentary rift complex. The Middle–Late Triassic series are represented in the well by clastic sediments and conglomerates (Ehlakov et al., 1991b).

From Jurassic times onward, the entire area of West Siberia began to subside regionally (Fig. 2d). Shallow-water and continental clastics were prominent during the post-Triassic, with exception of strata deposited at the transition of Jurassic to Cretaceous. During uppermost Jurassic time the inner part of the West Siberian Basin underwent an acceleration of subsidence with deposition of approximately 40 m of black bituminous sediment under starved basin conditions (the Bazhenov formation). During the Neocomian terrigenous clinoforms prograded into the basin and by the end of the Neocomian the basin returned to shallow water conditions again (Milanovsky, 1989).

Subsidence analysis of the Tyumenskaya well (Fig. 17c) shows a Late Permian–Early Triassic rapid subsidence phase, with the occurrence of basalts and deposition of terrigenous shallow water sediments. The Middle–Late Triassic sequence possibly reflects a phase of thermal subsidence. The Triassic–Jurassic boundary (208 Ma) exhibits a mild acceleration in subsidence, which is possibly related to an extensional event. From the Jurassic to Neogene sediment loading keeps up with continuous subsidence, only disturbed by relatively small-scale events. Since the Late Oligocene–Miocene the basin probably underwent syn-compressional uplift.

We have used a simple stretching model (McKenzie, 1978) to estimate the amount of extension required to explain this subsidence pattern, assuming a phase of Early Triassic extension. Differential crustal and subcrustal extension is required to model the large amount of post-rift subsidence. Adopting a pre-extensional crustal thickness of 40 km for the orogenically destabilized crust, we obtained estimates for the amount of crustal extension $\beta = 1.15$, subcrustal extension $\delta = 1.3$ and a duration of rifting of 10 Ma. The large size of the West Siberian basin probably contributed substantially to the long duration of the post-rift subsidence since thermal relaxation of the extended lithosphere cannot take place in this case through lateral heat transport.

6. Relationship between regional subsidence of the Russian platform and orogenic activity in adjacent belts

Fig. 18 shows cross-sections of Russian platform basins along three regional profiles (see Fig. 1b). Subsidence analysis of the platform carried out for seven wells (Fig. 19), demonstrates that the deposition of the cover of the Russian platform took place during three distinct phases in late Vendian–Silurian, mid-Devonian–Triassic and Jurassic–Palaeogene times (see also Aleinikov et al., 1980; Milanovsky, 1987). These phases are separated

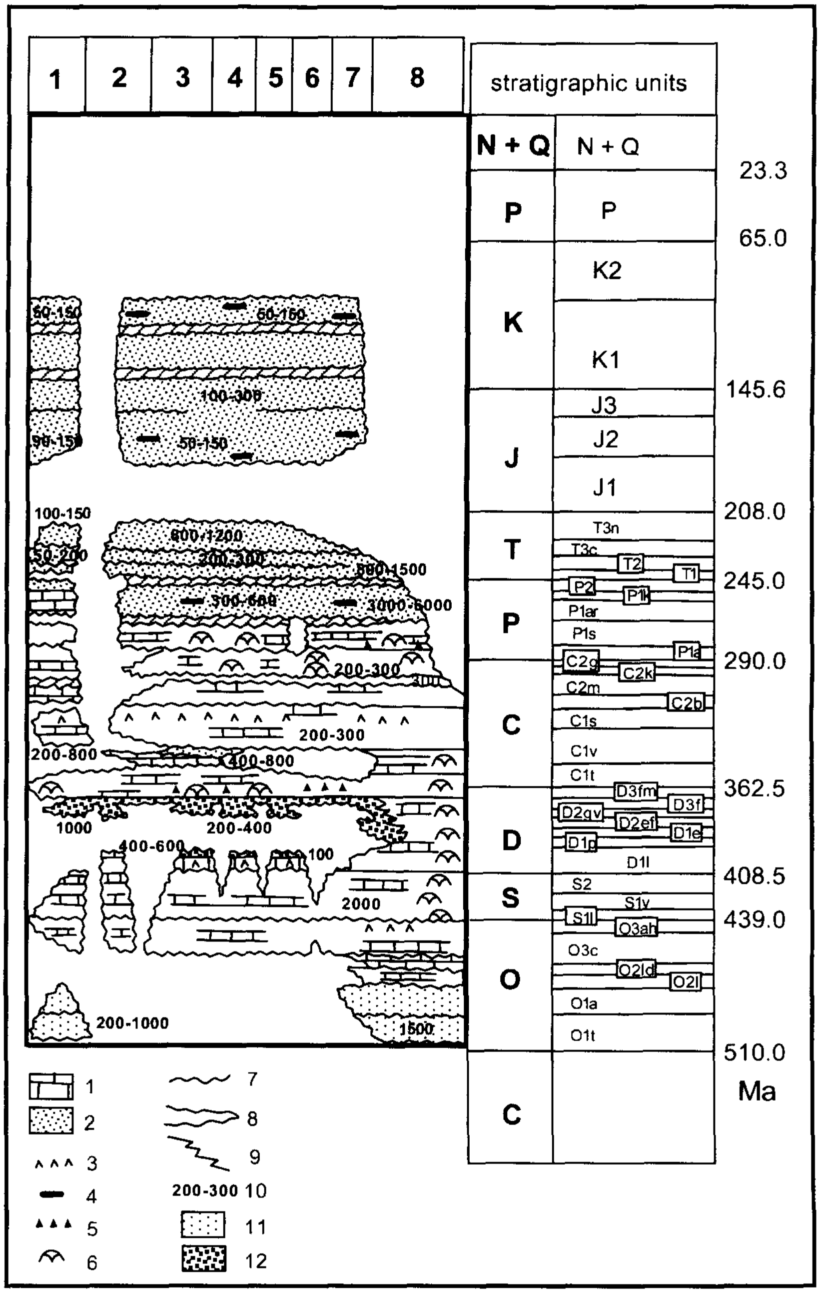


Fig. 14. Chronostratigraphic chart of the Timan–Pechora basin (after Belyakov, 1994): *1* = limestone and dolomite; *2* = terrigenous rocks; *3* = anhydrite and salt; *4* = coal; *5* = cherty-bituminous limestone (domanic facies); *6* = reef and bioherm; *7* = possible hiatuses; *8* = documented unconformities and hiatuses; *9* = facies boundaries; *10* = thickness (m); *11* = indexes of the general seismic reflection; *12* = red terrigenous rocks; *13* = basic intrusions and volcanics. Geological time scale after Harland et al. (1990). See for structural units areas *1–8* in Fig. 12.

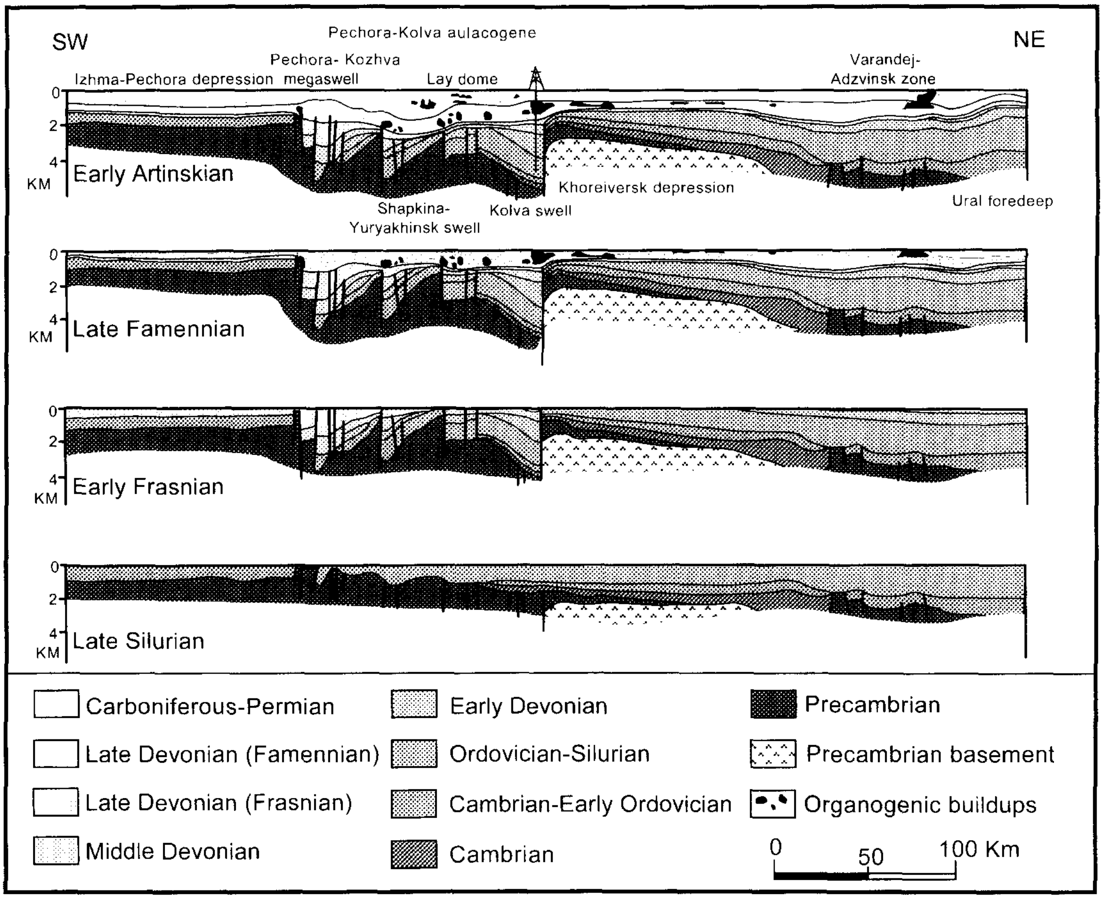


Fig. 15. Palaeoreconstruction of the Timan–Pechora basin geological evolution from Early Ordovician to late Artinskian times on the profile SW–NE (see Fig. 12) (after Belyakov et al., 1994).

by the Caledonian, Hercynian and Alpine deformation phases respectively (Fig. 20). Well data for the Moscow basin reflect these cycles in the subsidence record of the Pavlovo–Posadskaya well, Valday well and the Orsha well (Fig. 19a–c). During these three cycles of platform cover development the amplitude of regional tectonic subsidence was up to 0.5–1 km.

Regional subsidence occurred mainly in the platform areas located near the adjacent orogenic belts which were active at the same time (known in the Russian literature as the Karpinsky rule, see Milanovsky, 1987). During the Iapetus–Tornquist orogenic phase the main subsidence was concentrated on the western parts of the platform (late Vendian–Silurian). At the time of the Ural orogenic activity (Middle Devonian–Permian), subsidence was localized mainly in the eastern part of the platform, whereas during the Tethys orogenic activity (mid-Jurassic–Cenozoic) the main subsidence was concentrated in the southern part of the platform. The formation of the Tethys oceanic basin occurred in pre-Jurassic times (Zonenshain et al., 1990; Ziegler, 1990), followed by the development of a northward dipping subduction system in Jurassic–Eocene times. During Middle Jurassic–Eocene times, the southern part of the Russian platform underwent a significant subsidence.

Opening of the Ural palaeo-ocean occurred mainly in the Early Ordovician (Zonenshain et al., 1990). A stable orogenic system in association with a probably westward dipping subduction zone was created in Middle Devonian time. Simultaneously,

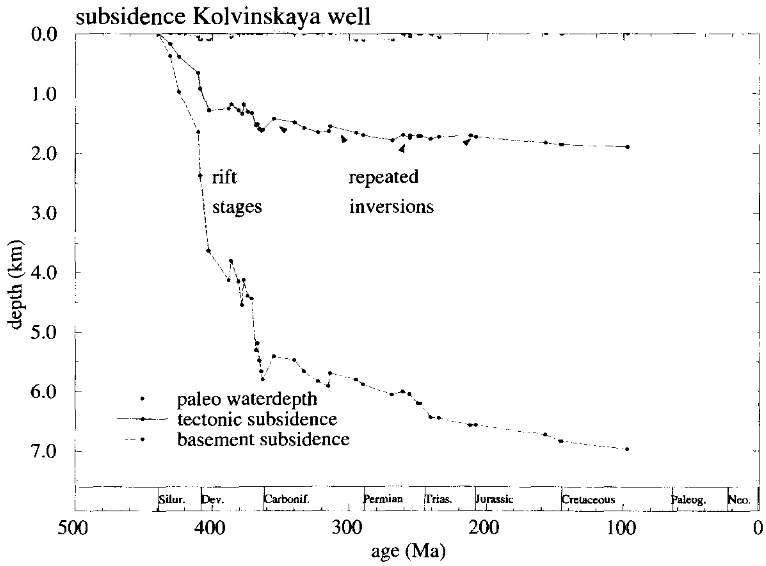


Fig. 16. Subsidence history for the Kolvinskaya well in the Pechora Basin (no. 5 in Fig. 1; see also Fig. 15). Subsidence history: Silurian — post-rift(?) subsidence; Early Devonian (Lochkovian) — rift phase; end of the Early Devonian — possible inversion phase; Middle–Late Devonian — rift phase or non-thermal subsidence; Carboniferous–Cretaceous — post-rift subsidence affected by a number of inversion events.

in Middle Devonian–Permian times the eastern part of the Russian platform underwent regional subsidence, following the Middle–Late Devonian rifting events. These findings support the existence of a causal relationship between the evolution of the subduction related orogenic belts and the platform subsidence histories. The documented long-term regional subsidence phases of the Russian platform were interrupted by numerous short-term uplift or rapid subsidence events (see Fig. 19). The timing of these short-term events is correlatable over wide areas. These features support an interpretation of the short-term anomalous uplift and subsidence events in terms of changes in intraplate stress fields (Cloetingh et al., 1985, 1989; Cloetingh and Kooi, 1992a).

7. Possible causes of non-thermal subsidence during the post-rift phase

Analysis of observed basin structures and subsidence characteristics of a number of rifted basins in the FSU show that a simple stretching model alone does not explain all features of the basin record. Post-rift subsidence appears to occur both faster and slower than predicted by the stretching model. The stretching model cannot explain observed stages of short-term rapid basin subsidence and does not account for long breaks between the rifting stage and the onset of post-rift subsidence as observed for the Russian platform. A particularly important observation is the extremely long duration of the post-rift subsidence phase in a number of these basins, compatible with similar findings for intracratonic basins in northern America (e.g., Leighton and Kolata, 1990). More stratigraphic data from wells, seismic profiles and outcrops are needed to constrain more complex basin formation models.

The analysis of rifted basins in the FSU demonstrates important differences in the transition of rifting stages to subsequent phases of post-rift evolution. The numerous Mesozoic rifts of the Urals–Mongolian belt form examples where no post-rift sedimentary basin has been formed (Fig. 2d). In the Pripyat–Dnieper–Donets basin, the Vilyuy basin and the Pechora basin a post-rift sedimentary basin begins to form just after completion of the rifting phase. In other cases, a post-rift sedimentary basin was formed tens or even hundreds million years after rifting completion. For example, four main rifting phases occurred in the East European platform during Riphean–early Vendian times followed in a number of aulacogens by post-rift subsidence begin-

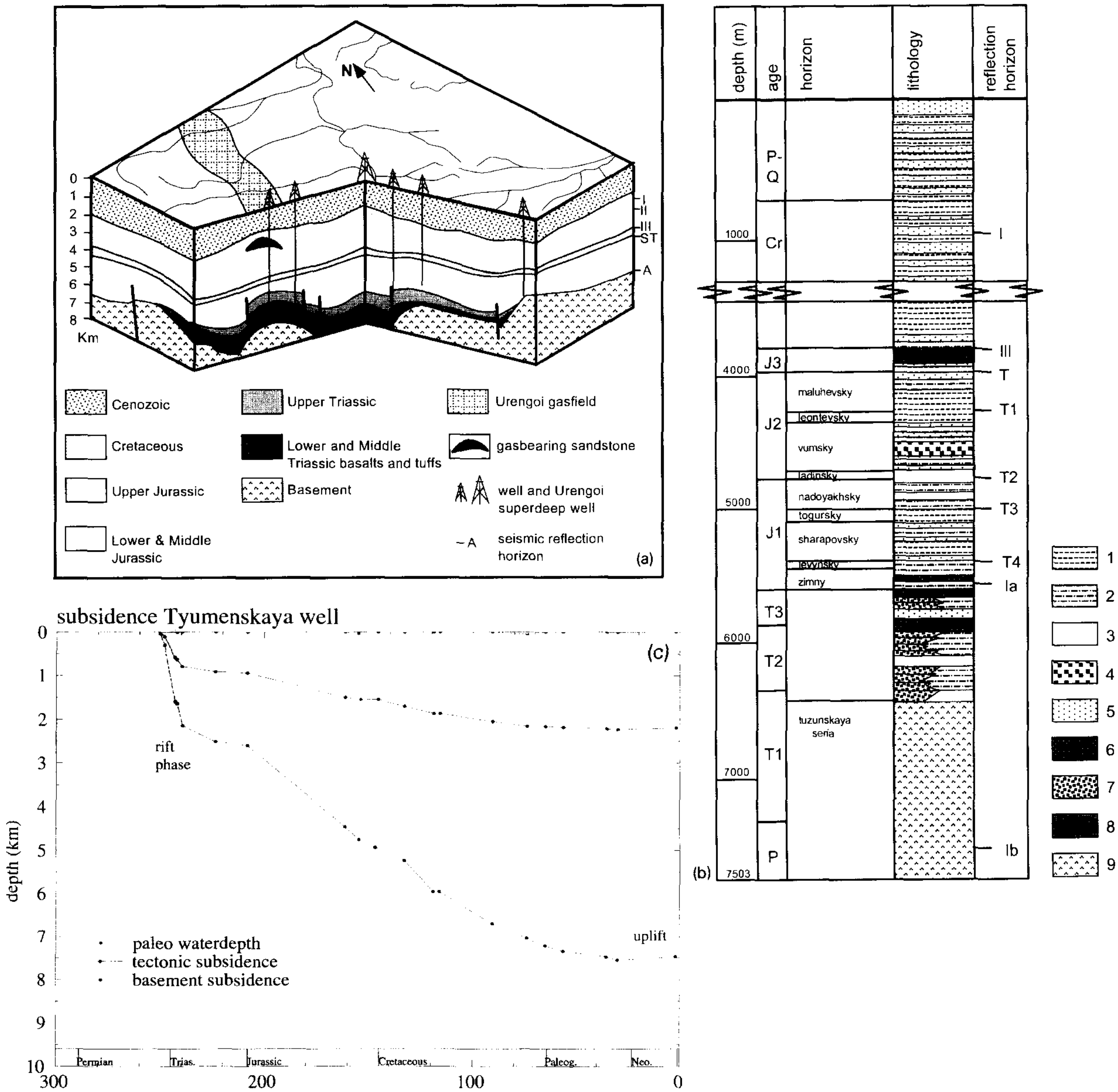


Fig. 17. (a) Structure of upper part of the crust for Tyumenskaya superdeep well region (after Surkov et al., 1993). (b) Lithological column of Tyumenskaya well: *1* = argillites; *2* = aleurolites; *3* = thinly layered argillites, aleurolites and sandstones *4* = coaly argillites; *5* = sandstones; *6* = conglomerates; *7* = gravelites; *8* = bituminous clays with silicon and pyrite concretions; *9* = basalt and tuffs. (c) Subsidence history for the Tyumenskaya well (No. 6 in Fig. 1). Timing and nature of subsidence history: Early Triassic — main rift phase; Middle–Late Triassic — post-rift subsidence with a possible inversion event at the end of Triassic times; Jurassic–Cenozoic — post-rift subsidence with a rapid subsidence event at the Jurassic–Cretaceous boundary and possible syn-compression uplift in the late Cenozoic.

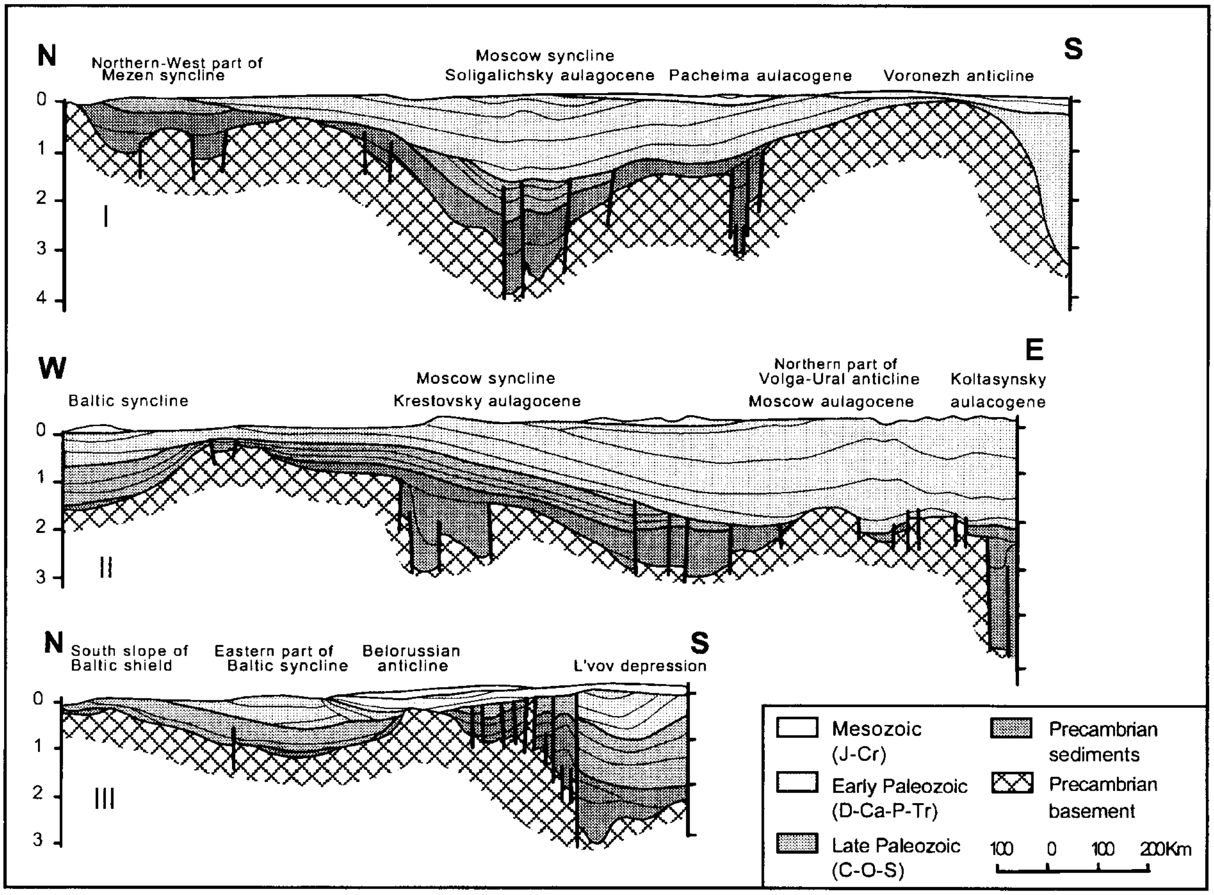
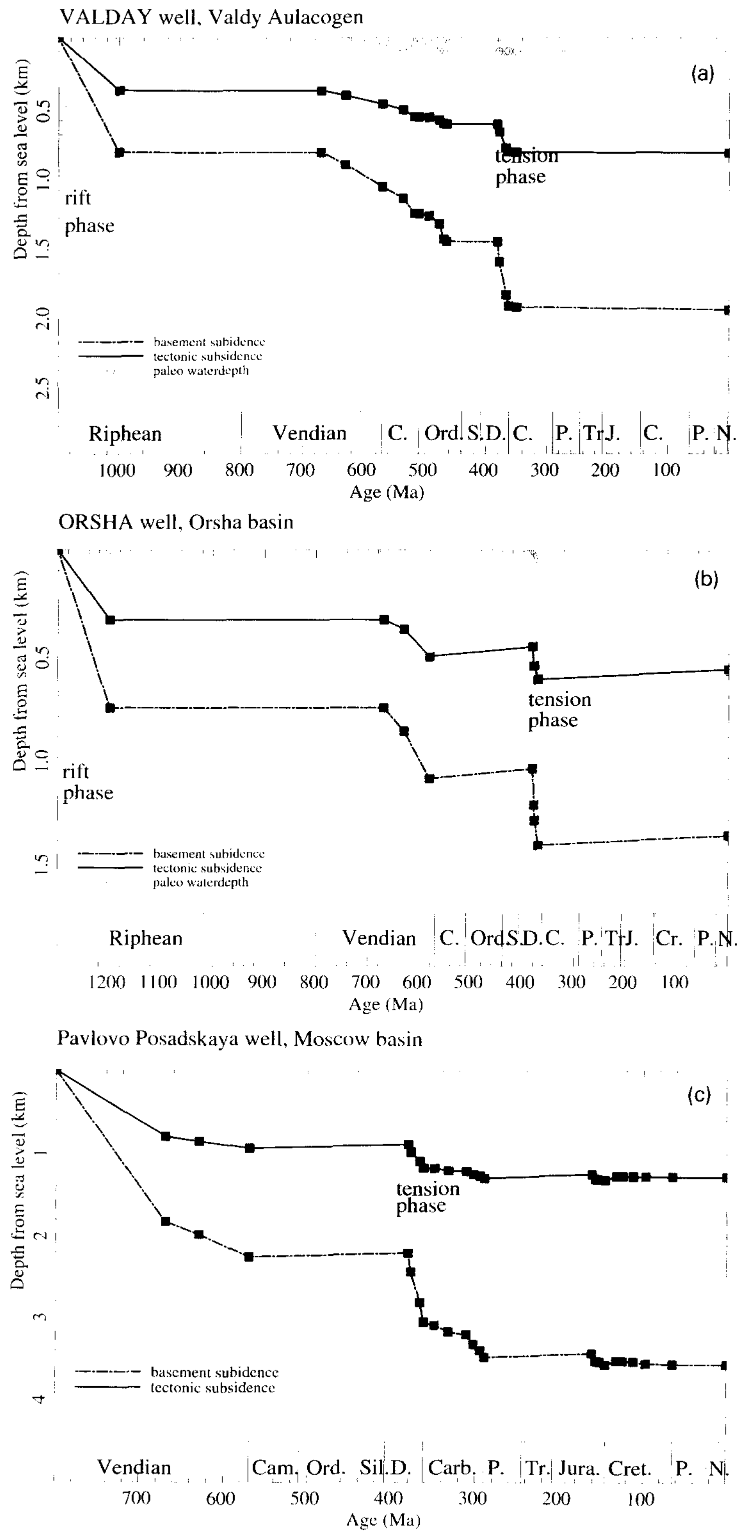


Fig. 18. North–south (lines AB and ED in Fig. 1b) and E–W (line CD in Fig. 1b) trending geological sections of the Russian platform (Milanovsky, 1987).

ning only in the mid-Vendian. The North Turan plate underwent rifting through the Triassic and Jurassic (Fig. 2d) with a phase of post-rift subsidence beginning only in the mid-Cretaceous (Milanovsky, 1989).

Post-rift subsidence of vast platform areas often coincides with the opening of an oceanic basin on the platform boundary, and is terminated or interrupted simultaneously with ocean closure and collision. The

Fig. 19. Subsidence history curves for seven wells in the Russian platform (see Fig. 1 for location of the wells): *a* = Valday well; *b* = Orsha well; *c* = Pavlovo–Posadskaya well; *d* = Pestovskaya well; *e* = Glazovskaya well; *f* = Issinskaya well; *g* = Oparinskaya well. Timing and main characteristics of Russian platform subsidence history: rift events in Riphean and early Vendian (Orsha well, Valday well, Pavlovo–Posadskaya well); molasse foreland subsidence in the late Vendian (Pestovskaya, Glazovskaya, Pavlovo–Posadskaya, Valday, Orsha wells); Early Cambrian inversion event (Pestovskaya, Glazovskaya, Orsha wells); early Palaeozoic platform subsidence (Pestovskaya, Glazovskaya, Valday wells); Late Silurian–Early Devonian inversion tectonics and uplifting (Pestovskaya, Glazovskaya, Pavlovo–Posadskaya(?), Valday, Orsha(?) wells); Middle–Late Devonian tension events (Pestovskaya, Oparinskaya, Issinskaya, Glazovskaya, Pavlovo–Posadskaya, Valday, Orsha wells); post-rift subsidence interrupted by a compressional event in the Ural belt (Carboniferous); Permian — foreland subsidence and inversion events in connection with Uralian orogeny (Pestovskaya, Oparinskaya, Issinskaya, Glazovskaya wells); Permian/Triassic boundary — weak tension event (Glazovskaya well); Triassic–Early Jurassic — uplifting of the platform(Pestovskaya, Oparinskaya, Issinskaya, Glazovskaya, Pavlovo–Posadskaya wells); Middle Jurassic–Cretaceous — platform subsidence (Pestovskaya, Oparinskaya, Issinskaya, Glazovskaya, Pavlovo–Posadskaya wells); post-Cretaceous — domination of uplift (possibly recorded in all wells).


VALDAY well, Valdy Aulacogen
(a)
Depth from sea level (km)
0.5
1.0
1.5
2.0
2.5
rift
phase
tension
phase
basement subidence
tectonic subsidence
paleo waterdepth
Riphean
Vendian
C.
Ord.
S.
D.
C.
P.
Tr.
J.
C.
P.
N.
1000
900
800
700
600
500
400
300
200
100
Age (Ma)
ORSHA well, Orsha basin
(b)
Depth from sea level (km)
0.5
1.0
1.5
rift
phase
tension
phase
basement subidence
tectonic subsidence
paleo waterdepth
Riphean
Vendian
C.
Ord.
S.
D.
C.
P.
Tr.
J.
Cr.
P.
N.
1200
1100
1000
900
800
700
600
500
400
300
200
100
Age (Ma)
Pavlovo Posadskaya well, Moscow basin
(c)
Depth from sea level (km)
1
2
3
4
tension
phase
basement subsidence
tectonic subsidence
Vendian
Cam.
Ord.
Sil.
D.
Carb.
P.
Tr.
Jura.
Cret.
P.
N.
700
600
500
400
300
200
100
Age (Ma)

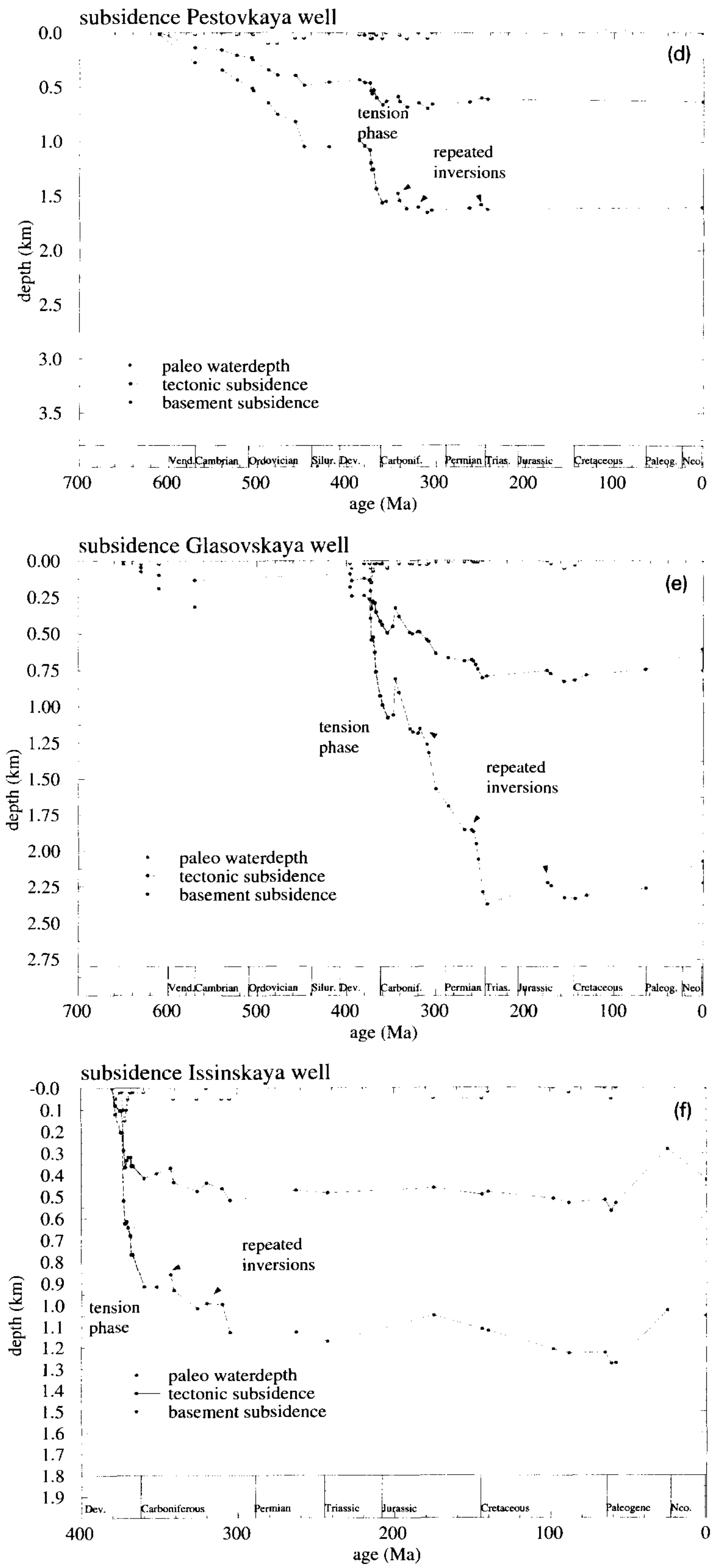


Fig. 19. Continued.

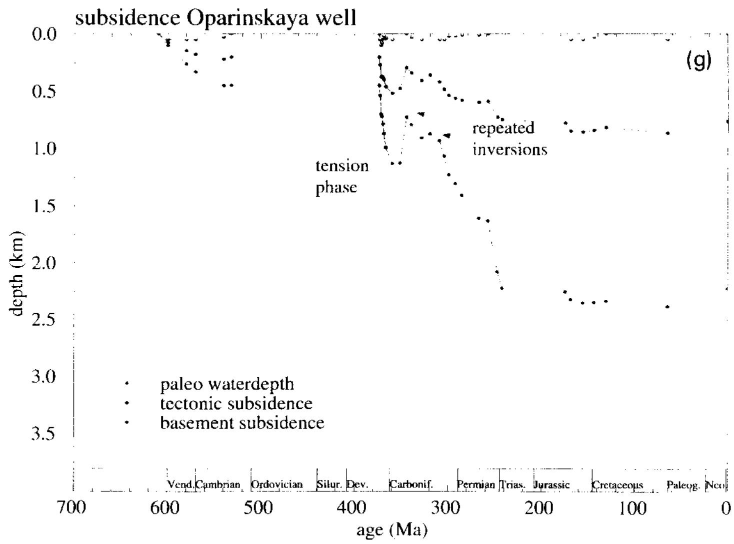


Fig. 19. Continued.

Iapetus Ocean, for example, began to open in the late Vendian; at the same time post-rift basins of the East European platform began to subside. In the Late Silurian–Early Devonian, simultaneous with the Caledonian orogeny, this subsidence stopped or was interrupted.

In the late Riphean (about 800 Ma), simultaneous with the opening of an oceanic basin at its southern and western margin, almost the whole East Siberian platform underwent rapid subsidence and deposition of marine sediments. In the Late Silurian–Early Devonian subsidence was terminated simultaneous with the Caledonian orogeny south and west of the platform.

In some cases pre-rift and syn-rift volcanism could have contributed to post-rift subsidence. Rifts characterized by large-scale volcanism ('wet' rifts) are in many cases not followed by deep post-rift subsidence and the development of wide sedimentary basins (e.g., the Palaeozoic Oslo rift). Areas of flood basalt volcanism without rifting or considerable extension (e.g., Stel et al., 1993) most times do not undergo noticeable subsidence after the completion of volcanism, but can stay in an uplifted position as observed in the flood basalt provinces in the Siberian platform.

It appears, therefore, that in addition to stretching, other mechanisms such as eclogite formation in the lower crust (Fowler and Nisbet, 1990) or beneath extended lithosphere (Lobkovsky et al., 1993), and regional compression of the lithosphere during the post-rift phase (Cloetingh, 1988) play a potentially significant role.

7.1. Eclogite lens formation in the upper mantle beneath extended lithosphere and related subsidence

The role magmatism plays in the formation of sedimentary basins is not yet sufficiently understood (e.g., Quinlan et al., 1993; Wilson, 1993). Extension of the lithosphere results in decompression of the underlying asthenosphere, causing partial melting and advection of hot asthenospheric material to the base of the lithosphere and into the space created by its extension (LePichon and Sibuet, 1981; Spohn and Schubert, 1983; Neugebauer, 1983; McKenzie and Bickle, 1988; Ziegler, 1992). Continued lithosphere thinning causes further decompression of the asthenosphere and the generation of additional partial melts. The partial melts must then ascend and segregate from its parent rock by processes that result in emplacement of the magma either at the earth's surface, within the crust or near the lithosphere–asthenosphere boundary.

Most models presented over the last few years place the depths of the phase changes at levels in the lower crust (e.g., Fowler and Nisbet, 1990). Density increases due to basalt-eclogite phase trans-

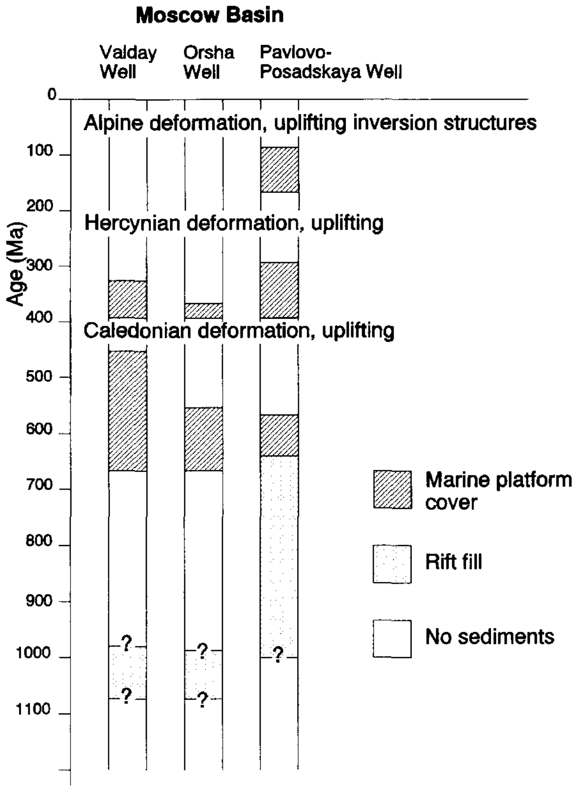


Fig. 20. Main phases of subsidence and uplift of the Russian platform documented for the Moscow basin.

formations in the lower crust have, for example, been proposed as a possible cause of sedimentary basin formation (Falvey, 1974; Haxby et al., 1976; Artyushkov and Sobolev, 1982; Stel et al., 1993).

Recent experimental data indicate (see Carswell, 1990, for a review) that the minimum pressures required for eclogite stability in rocks of lower crustal composition could be in excess of pressure estimates from linear extrapolation of the plagioclase-out reaction curves of Green and Ringwood (1967) to the low *P–T* range. This implies that large-scale eclogite formation in the lower crust for crustal thicknesses less than 40 km does not necessarily occur (Ringwood, 1975; Carswell, 1990). This is important, as currently available seismic data reviewed by Siemov (1987) show that the crust underlying most of the extensional basins in the FSU is less than 35 km thick. As shown by a number of recent studies (e.g., Sleep et al., 1980; Kooi, 1991), detailed investigations of gravity data could provide useful constraints to determine the depth range of the inferred phase changes.

An alternative scenario, placing the phase changes in the upper mantle, has been proposed by Lobkovsky et al. (1993). In this model the lithosphere–asthenosphere boundary prevents further rising of the buoyant magma, leading to a concentration as a magmatic lens in the upper part of an asthenospheric bulge (Fig. 21). In the following we explore this model, assuming that the basalt melt in the asthenospheric bulge was not transported to shallower levels at the surface but formed a magmatic lens in which crystallization took place during the process of cooling during the post-rift phase. The theoretical foundation for this model is based on the behaviour of a permeable porous medium saturated with a two-phase melt and a visco-deformed skeleton (Karakin and Lobkovsky, 1979, 1982; Scott and Stevenson, 1986; McKenzie, 1984; Richter and McKenzie, 1984).

Eclogite lens formation in the upper mantle can occur as a result of two simultaneously operating processes: (1) large-scale upper mantle heating due to active uplift of mantle material (hot spot mechanism); and (2) stretching resulting in lithospheric thinning and necking and passive uplift of the asthenosphere. This can result in the formation of an asthenospheric bulge under the rift and concentration of basalt magma due to vertical filtration near the bulge roof (Fig. 21). The subsequent evolution of the rift system depends on whether magmatic liquid remains within the lens after completion of the extension or whether it penetrates the crust and reaches the surface. The first case corresponds to magma consolidation into deep eclogite mineral facies and high density eclogite lens formation below the lithosphere leading to rifting, subsidence and the formation of a deep sedimentary basin (Lobkovsky et al., 1993; Ismail-zadeh et al., 1994). The second scenario (see Fig. 21) can lead to the formation of a flood basalt province or a volcanic rifted margin, without noticeable subsidence or even uplift (White and McKenzie, 1989; Coffin and Eldholm, 1991).

Increases in fluid flow rates induced by increases in the level of compressional intraplate stress in the lithosphere (Van Balen and Cloetingh, 1993) can enhance the reaction rates of phase changes and, hence, contribute to the effectivity of eclogitization as a mechanism for post-rift subsidence (Cloetingh and Kooi, 1992a).

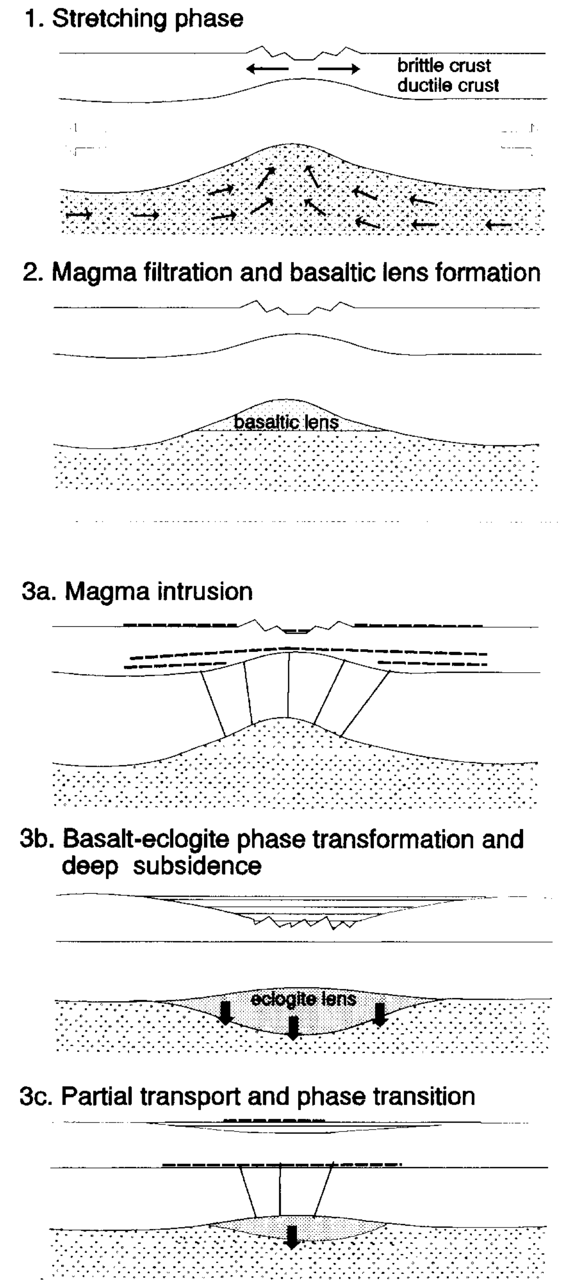


Fig. 21. Stages of post-rift evolution of the lithosphere for a model invoking an underlying magmatic lens (After Lobkovsky et al., 1993). (1) Rifting stage, formation of an asthenospheric bulge and filtration of magmatic melt. (2) Accumulation of melt in the upper part of the bulge and formation of a magmatic lens. (3a) Transport of magma from the lens to the surface, accompanied by intensive volcanic activity and intrusion of magmatic material in the lower crust. (3b) Crystallization of the magmatic lens into an eclogite body, followed by subsidence and formation of a deep sedimentary basin. (3c) Partial transport of magma from the lens and transition of the remaining part into eclogite rocks; this stage is accompanied by a moderate volcanic activity, minor subsidence and formation of a shallow sedimentary basin.

7.2. Stress-induced subsidence perturbations

Intraplate stress fluctuations in the lithosphere can also directly contribute to non-thermal subsidence during the post-rift phase of extensional basins (Cloetingh, 1988; Kooi and Cloetingh, 1989; Cloetingh et al., 1989; Ziegler et al., 1995). The Russian platform appears to be a good study area to quantify the effect of the stresses on subsidence and uplift patterns. The compressional stresses can lead to different expressions on different spatial scales: acceleration of subsidence such as observed for the North Sea area (Cloetingh et al., 1990), broad-scale uplifting such as observed for the north American craton (Ziegler et al., 1995) and inversion tectonics documented extensively in northwestern Europe (Ziegler, 1990) and the Donets Basin (Stephenson et al., 1993). The actual response of the lithosphere to intraplate stresses depends on many factors, including the load configuration and rheology of the underlying lithosphere (Cloetingh et al., 1989; Kooi and Cloetingh, 1992; Ziegler et al., 1995). For high levels of intraplate stresses, approaching estimates of lithospheric strength (Burov and Diament, 1995; Cloetingh and Burov, 1996) these stresses can induce large-scale lithosphere folds such as observed in Central Asia (Nikishin et al., 1993; Burov et al., 1993) and Arctic Canada (Stephenson et al., 1990). Rapid accelaration of the post-rift subsidence of the Dnieper basin in Visean–Early Permian times can be explained by the effect of superimposed compressional stresses propagated from the Variscan–Dobrogea–Caucasus collisional belt (Nikishin et al., 1996). Accelaration of subsidence of the Russian platform and Timan–Pechora basin in post-Devonian times and inversion events such as inferred for Carboniferous–early Mesozoic times (see Fig. 19) are also compatible with this mechanism. Further modeling studies and analysis of subsidence data is required to quantify these causal relationships.

8. Conclusions

We have demonstrated that stretching models of rifting and sedimentary basin formation cannot fully explain a number of important features of the evolution of rifted basins in the FSU. In our analysis of rifted basins in the FSU we have encountered important differences in the transition of rifting stages to subsequent phases of post-rift evolution. Post-rift sedimentary basins were sometimes formed tens or even hundreds of million years after the completion of the rifting. Post-rift subsidence of vast platform areas often begins more or less simultaneously with the opening of an oceanic basin on the platform margin and ceases or is interrupted simultaneously with the closure of the oceanic basin and the formation of a collision belt. In some cases the formation of deep sedimentary basins occurs without clear evidence for extension. A first analysis of basin structures and subsidence characteristics of rifted basins in the FSU points to an important role of non-thermal subsidence mechanisms such as phase changes and eclogite lens formation beneath thinned lithosphere. The subsidence record and basin geometries discussed in this paper provide also a number of key examples of the effects of regional stresses on basin histories in the FSU. Future work has to focus on the development of quantitative models of FSU basin evolution constrained by the full range of available geological and geophysical data.

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