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Late Precambrian to Triassic history of the East European Craton: dynamics of sedimentary basin evolution

A.M. Nikishin [a,*], P.A. Ziegler [b], R.A. Stephenson [c], S.A.P.L. Cloetingh [c], A.V. Furne [a], P.A. Fokin [a], A.V. Ershov [a], S.N. Bolotov [a], M.V. Korotaev [a], A.S. Alekseev [a], V.I. Gorbachev [d], E.V. Shipilov [e], A. Lankreijer [c], E.Yu. Bembinova [a], I.V. Shalimov [a]

[a] *Geological Faculty, Moscow State University, Leninskie Gory, 119899 Moscow, Russia*
[b] *Geological-Palaeontological Institute, University of Basel, Bernoullistr. 32, 4056 Basel, Switzerland*
[c] *Vrije Universiteit, Faculty of Earth Sciences, De Boelelaan 1085, 1081 HV Amsterdam, Netherlands*
[d] *NEDRA, Yaroslavl, Russia*
[e] *Institute of Marine Geophysics, Murmansk, Russia*

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Abstract

During its Riphean to Palaeozoic evolution, the East European Craton was affected by rift phases during Early, Middle and Late Riphean, early Vendian, early Palaeozoic, Early Devonian and Middle–Late Devonian times and again at the transition from the Carboniferous to the Permian and the Permian to the Triassic. These main rifting cycles were separated by phases of intraplate compressional tectonics at the transition from the Early to the Middle Riphean, the Middle to the Late Riphean, the Late Riphean to the Vendian, during the mid-Early Cambrian, at the transition from the Cambrian to the Ordovician, the Silurian to the Early Devonian, the Early to the Middle Devonian, the Carboniferous to Permian and the Triassic to the Jurassic. Main rift cycles are dynamically related to the separation of continental terranes from the margins of the East European Craton and the opening of Atlantic-type palaeo-oceans and/or back-arc basins. Phases of intraplate compression, causing inversion of extensional basins, coincide with the development of collisional belts along the margins of the East European Craton. The origin and evolution of sedimentary basins on the East European Craton was governed by repeatedly changing regional stress fields. Periods of stress field changes coincide with changes in the drift direction, velocity and rotation of the East European plate and its interaction with adjacent plates. Intraplate magmatism was controlled by changes in stress fields and by mantle hot-spot activity. Geodynamically speaking, different types of magmatism occurred simultaneously.

Keywords: East European Craton; palaeotectonics; sedimentary basins; palaeogeography

1. Introduction

The East European Craton (EEC) is a classical region for the study of intracratonic sedimentary basins, intraplate tectonic phenomena and for the analysis of the dynamic processes governing them. The EEC is bounded in the east by the Uralian–Novaya Zemlya–Taymyr orogen, to the south by the Scythian and Crimean–Caucasus orogens, to the southwest by the Tornquist Line, to the northwest

* Corresponding author.

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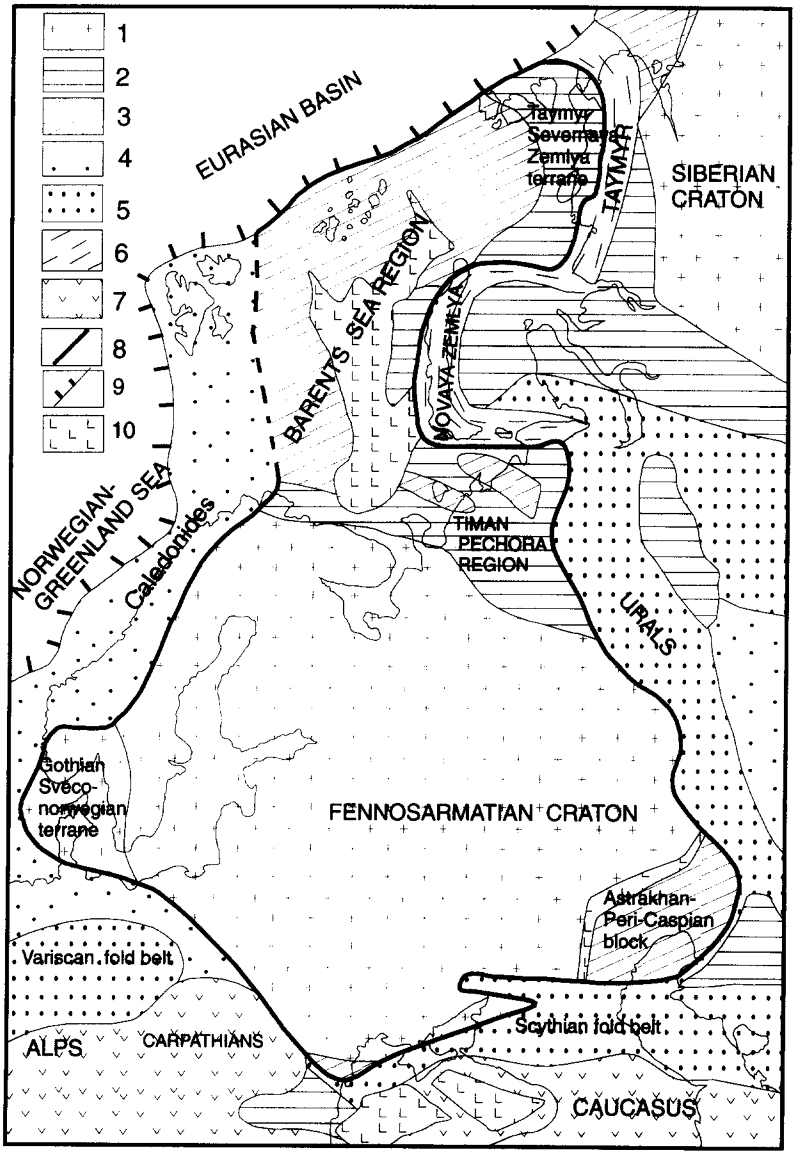


Fig. 1. Main provinces of the East European Craton. *1* = Archaean and Early Proterozoic terranes, undivided; *2* = 1.1–0.55 Ga age terranes; *3* = Precambrian terranes, undivided; *4* = early Palaeozoic fold belts; *5* = late Palaeozoic fold belts; *6* = Mesozoic fold belts; *7* = Alpine fold belts; *8* = outlines of East-European Craton; *9* = passive continental margins; *10* = area with very thin continental or oceanic crust.

by the Scandinavian Caledonides and to the north by the oceanic Eurasian Basin (Fig. 1). The EEC is characterized by a Precambrian crystalline basement outcropping in the Fennoscandian (Baltic) and Ukrainian shields and the Voronezh Massif; elsewhere, this basement is covered by sediments with thicknesses up to 23 km. The Late Precambrian and Phanerozoic evolution of the EEC has been investigated by many researchers, including Shatsky (1964), Vinogradov (1969), Bogdanov (1976), Bronguleev (1978), Bogdanov and Khain (1981), Bronguleev (1985), Milanovsky (1987), Ziegler (1989, 1990) and Khain and Seslavinsky (1991).

This paper reviews, on the basis of a set of interpretative palaeogeographic–palaeotectonic maps and supporting basin analysis studies, the Late Precambrian and Palaeozoic evolution of the EEC. The dynamics of the evolution of sedimentary basins are analyzed in the framework of processes affecting the margins of the craton, resulting from its interaction with adjacent plates. This paper provides the reader with an introduction to our current understanding of the geology and geodynamics of Eastern Europe.

2. Basement provinces

The Precambrian EEC consists of several basement provinces which differ in the age of their consolidation (Fig. 1). The main constituents are: the Early Precambrian Fennosarmatian Craton (terminology of H. Stille; Milanovsky, 1987); the Gothian Sveconorwegian terrane (1.75–1.55 Ga), which was welded to Fennosarmatia during the Dalslandian orogeny (1.2–0.9 Ga; Gorbatchev and Bogdanova, 1993); the Late Precambrian Pechora–Barents Sea orogenic belt (Pechora–Barentsia), which was accreted to the northeastern margin of Fennosarmatia during Vendian–Early Cambrian times; the Late Precambrian Taymyr–Severnaya Zemlya terrane, which was accreted to the evolving EEC probably at the end of the Cambrian (Dedeev and Zaporozhtseva, 1985; Getsen, 1987; Uflyand et al., 1991; Zonenshain et al., 1993); and the Astrakhan–Peri-Caspian (Nevolin, 1988) and Pre-Dobrogea blocks of possibly Late Precambrian age (Garetsky, 1990). The palaeo-continent incorporating all these older than Early Riphean blocks is here referred to as Baltica. Fennosarmatia is the core of the Baltica palaeo-continent, the outlines of which changed significantly through time.

The Fennosarmatian craton consists of Archaean domains and Early Proterozoic orogenic belts (Bogdanov and Khain, 1981; Bogdanova, 1986; Milanovsky, 1987; Park, 1991; Nikishin, 1992; Gorbatchev and Bogdanova, 1993; Abbot and Nikishin, 1996). Prior to the Riphean, the EEC was affected by major orogenic cycles during which several continental terranes were amalgamated to form a huge composite continental craton; this process terminated with the development of the Andean-type Trans-Scandinavian magmatic belt (1.82–1.65 Ga; Park, 1991; Gorbatchev and Bogdanova, 1993). Orogenic and magmatic activity terminated during the earliest Riphean. To the east of the Trans-Scandinavian magmatic belt, a large anorogenic rapakivi-granite province developed between 1.65 and 1.55 Ga (Gorbatchev and Bogdanova, 1993). The rapakivi granites of the Ukrainian Shield are 0.1–0.2 Ga older (Shcherbak, 1991; Gorbatchev and Bogdanova, 1993).

The Pechora–Barents Sea basement province is poorly known. The few wells which reached basement yield ages ranging from 1.0 to 0.55 Ga (Dedeev and Zaporozhtseva, 1985; Kogaro et al., 1992; Senin, 1993; Shipilov, 1993; Puchkov, 1993). The composite Pechora–Barentsia terrane was probably welded to Fennosarmatia at the end of the Precambrian (Dedeev and Zaporozhtseva, 1985; Zonenshain et al., 1993; Puchkov, 1993).

The Late Precambrian Northern Taymyr–Severnaya Zemlya block contains remnants of Cambrian flysch, unconformably covered by Ordovician platform series (Milanovsky, 1987). Reconstruction of the early Palaeozoic history of this region is problematic; however, it is possible that the Northern Taymyr–Severnaya Zemlya block was accreted to the Pechora–Barentsia terrane at the transition from the Cambrian to the Ordovician.

No reliable age dates are available for the possibly Late Precambrian (Dalslandian?) Astrakhan and Pre-Dobrogea blocks.

3. Working methods

The sedimentary cover of the EEC records its Late Precambrian and Phanerozoic evolution (Fig. 2). Geophysical and well data provide information on the geometry and deformation of sedimentary basins and permit analysis of their subsidence and uplift histories and of the dynamic processes governing their evolution. Determination of the timing of the different tectonic events, which controlled basin subsidence and destruction, depends entirely on the accuracy of dating the different sedimentary sequences.

Riphean and Vendian sediments record a Riphean–early Vendian rifting (aulacogen) and a late Vendian–early Palaeozoic platform stage (Shatsky, 1964; Bogdanov, 1976; Milanovsky, 1987). Although wells provide ample control on the thickness and lithological composition of the Riphean sediments, palaeontological control is poor and often conflicts with the available K/Ar dates. As correlations between basins are not very reliable, the stratigraphic chart

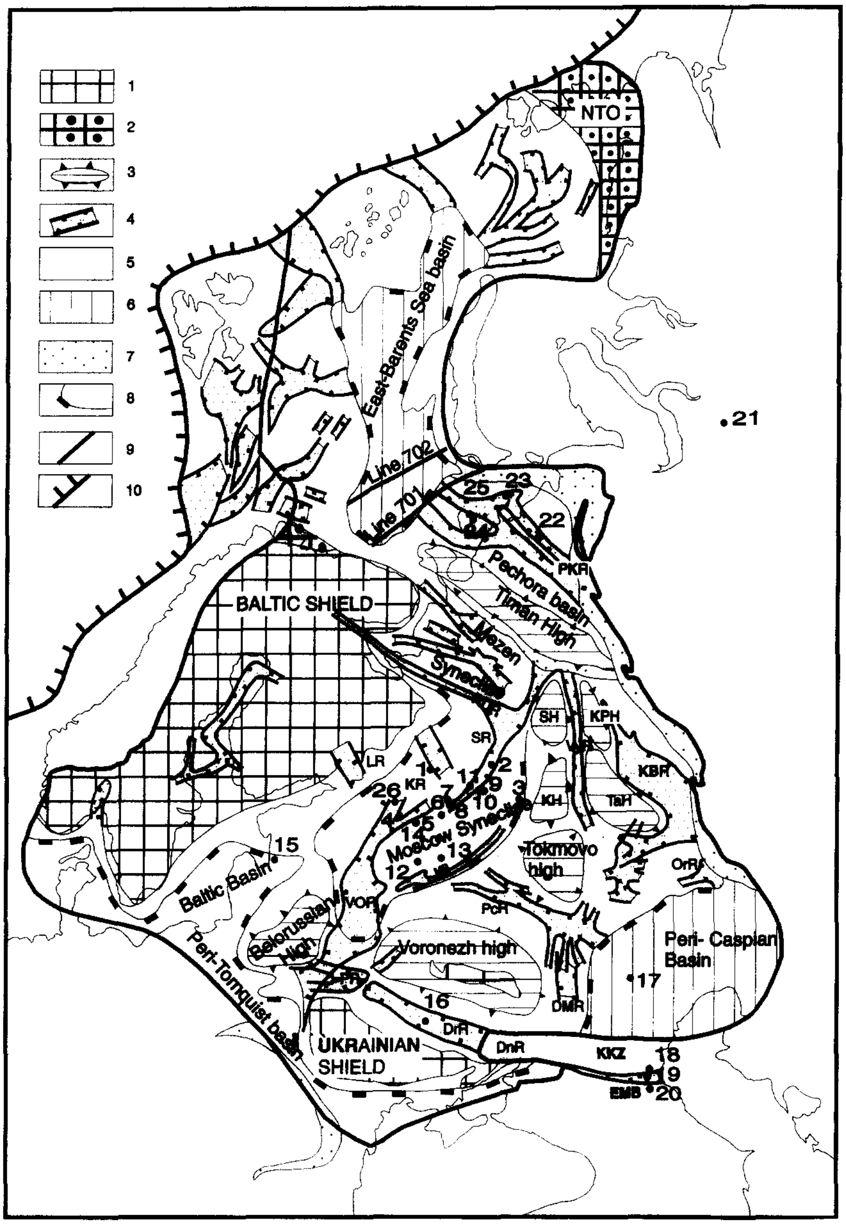


Fig. 2. Main sedimentary basins of the East European Craton and location of wells and seismic profiles used for 1-D and 2-D modelling. *1* = Early Precambrian basement (shield); *2* = Late Precambrian basement (0.8–0.55 Ga); *3* = intraplatform highs; *4* = rifted basins; *5* = platform areas; *6* = very deep sedimentary basin (up to 20–23 km); *7* = foreland basins (Fore-Ural basin, Fore-Timan basin, Fore-Carpathian basin); *8* = outlines of some sedimentary basins; *9* = boundary of EEC; *10* = passive margin of the Atlantic–Arctic Ocean. Names of structures: *DMR* = Don-Medveditsa rifted basin; *DnR* = Donets rifted basin; *DrR* = Dniepr rifted basin; *EMB* = East-Manych basin; *KBR* = Kama-Belaya rifted basin; *KDR* = Kandalaksha–Dvina rifted basin; *KH* = Kotelnich High; *KKZ* = Karpinsky Kryazh zone; *KPH* = Komi–Permyak High; *KR* = Krestets rifted basin; *LR* = Ladoga rifted basin; *MR* = Moscow rifted basin; *NTO* = North-Taymyr orogen; *OrR* = Orenburg rifted basin; *PcR* = Pachelma rifted basin; *PKR* = Pechora–Kolva rifted basin; *PR* = Pripyat rifted basin; *SH* = Sysola High; *SR* = Soligalich rifted basin; *TaH* = Tatarian High; *VOR* = Volyn–Orsha rifted basin; *VyR* = Vyatka rifted basin. Solid lines labelled 701 and 702: location of profiles given in Fig. 11. Black dots and numbers: location of wells used for subsidence analyses in Figs. 5 and 6.

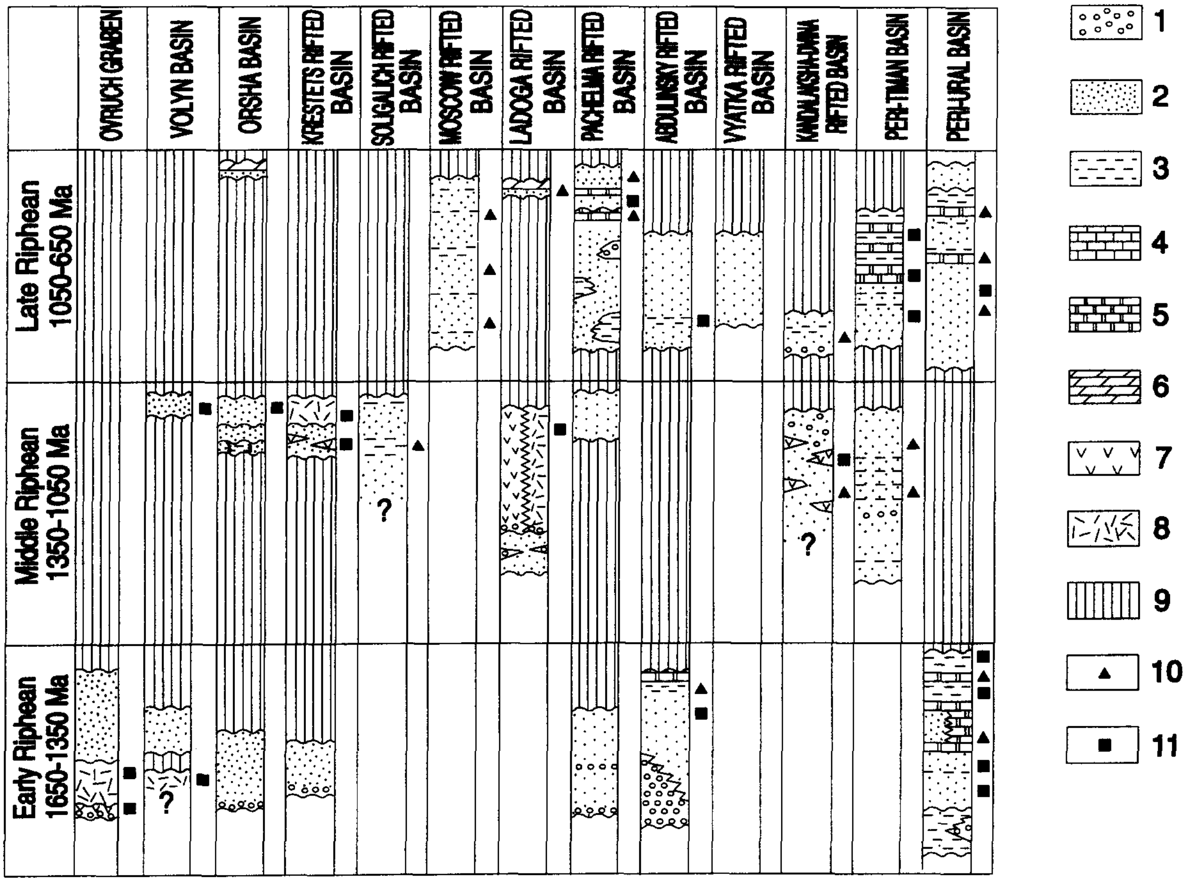


Fig. 3. Stratigraphic scheme of Riphean sediments: *1* = conglomerates and conglomeratic sands; *2* = sandstones and siltstones; *3* = mudstones; *4* = limestones; *5* = dolomites; *6* = marls; *7* = basalts; *8* = tuffs; *9* = hiatus; *10* = fossils; *11* = isotope dating.

summarising the Riphean sedimentary record of the Russian Platform (Fig. 3) must be considered as tentative. For the same reason, it is difficult to develop Riphean palaeogeographic–palaeotectonic maps for the entire craton. In contrast, the Vendian and Palaeozoic stratigraphy is quite well constrained by biozonations; therefore, basin-to-basin correlations, summarised in Fig. 4, are more reliable. For the Precambrian, the time scale of Semikhatov et al. (1991) and for the Phanerozoic, the Harland et al. (1990) scale were adopted.

Our analysis of the Riphean to early Mesozoic evolution of the EEC is based on regional compilations and syntheses of stratigraphical and structural data, an inventory of magmatic activity, and the construction of a set of regional, interpretative palaeogeographic–palaeotectonic maps. Special attention was paid to the distribution and facies development of mapping intervals in terms of depositional versus erosional basin edges, transport directions of clastics, monitoring the uplift of adjacent cratonic highs or orogenic belts, and to intrabasinal syn-depositional tectonic and volcanic activity. The evolution of selected basins (cf. Fig. 2) was analyzed by calculating tectonic subsidence curves on the basis of well data, applying conventional backstripping techniques (e.g., Watts, 1992); examples of such subsidence curves are shown in Fig. 5. Subsidence curves were converted into plots of subsidence and uplift rate (Fig. 6) permitting ready comparison of the different basins. Selected basin transects, controlled by reflection-seismic profiles, were analyzed and palinspastically restored (cf. Nikishin et al., 1995).

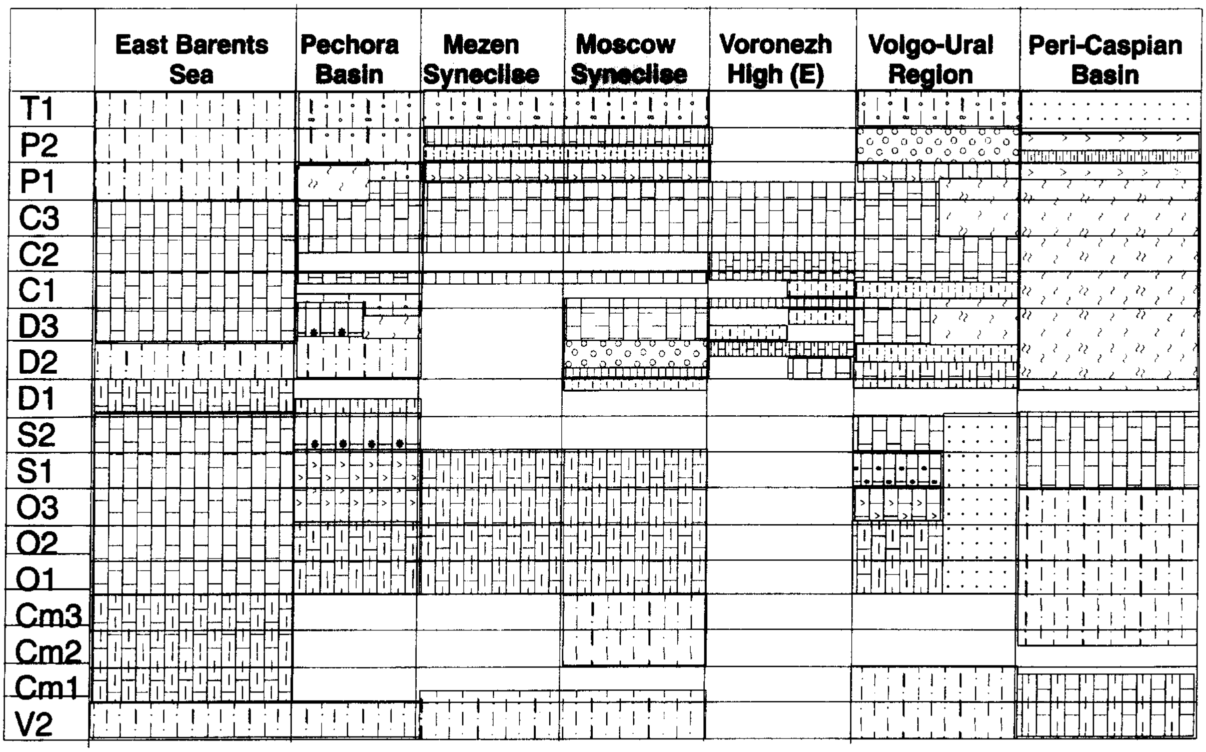


Fig. 4. Late Vendian–Palaeozoic stratigraphic record of the East European Craton. For legend, see Fig. 7.

Palaeogeographic–palaeotectonic maps were constructed for a sequence of time slices, corresponding to the main steps in the evolution of the craton. These time-slices are large and variable. Preliminary versions of these maps, showing the present distribution of sediments pertaining to the respective time interval, as well as hypothetical depositional basin outlines, are given in Fig. 7 in a present-day geographic framework. In discussions of these maps all direction indications pertain to the present coordinates of the EEC.

4. Riphean to Early Triassic evolution of the EEC

A stage-by-stage discussion of the Riphean to Triassic evolution of the EEC is given below. For each of the main stages one or several schematic maps are presented. More detailed maps are in preparation and will be published at a later date. Here, the main conclusions and outstanding problems of ongoing research are highlighted.

4.1. Early Riphean (1650–1350 Ma)

In pre-Riphean times, the EEC was affected by a sequence of major orogenic cycles during which continental and orogenic terranes were amalgamated to form a large continental craton referred to as 'Baltica'.

During and just before the Early Riphean a large rapakivi-granite province developed east of the Trans-Scandinavian magmatic belt (Fig. 7a). Intrusion of the more than 20 constituent plutonic massifs was accompanied by the subsidence of volcano-tectonic depressions above them and by the injection of major dike swarms. The duration of rapakivi-granite plutonism spanned about 200 m.y. and terminated at ca. 1.5 Ga (Milanovsky, 1987; Svetov et al., 1990; Shcherbak, 1991; Haapala and Ramo, 1992; Gorbatchev and Bogdanova, 1993).

During the middle Early Riphean, the southeastern parts of Baltica were affected by a major rift cycle, the timing of which is, however, poorly constrained by radiometric data. Major rifts include the

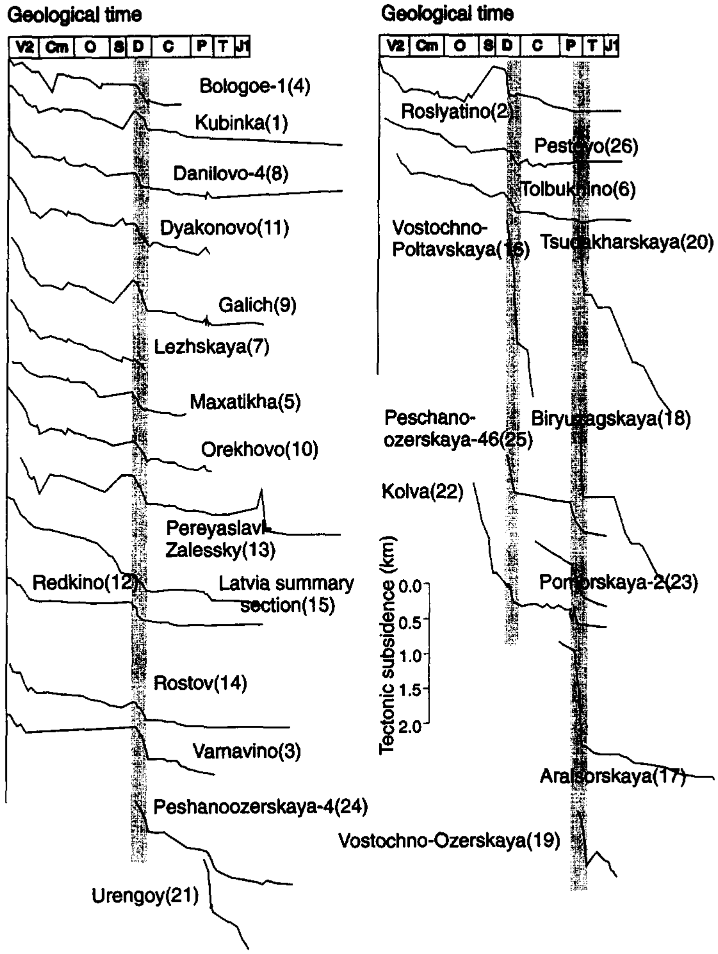


Fig. 5. Tectonic subsidence for some deep wells. Location of wells is shown in Fig. 2. Time scale after Harland et al. (1990). Eustatic sea-level changes were disregarded.

deep Peri-Ural Basin and its branches, the Pachelma, Orenburg and Abdulino basins. Sediments, attaining thicknesses of more than 6 km, are developed mainly in a continental clastics facies in the craton-ward parts of these basins and grade laterally into shallow-marine series and carbonates. Sedimentation was accompanied by the intrusion of tholeiitic basaltic magmas and a basic plutonism (Mirchink, 1977; Lagutenkova and Chepikova, 1982; Lozin, 1994). As the tectonic setting of these rifts resembled that of an Atlantic-type rift system, it is very likely that the middle Early Riphean rifting cycle culminated in the separation of a continental terrane from the south-eastern margin of Baltica and the opening of a new oceanic basin.

A regional hiatus between the Early and Middle Riphean series (Fig. 3) indicates that towards the end of the Early Riphean the entire craton was uplifted, perhaps in response to compressional deformation of its western and eastern margins. Occurrences of metamorphic rocks and granitic plutons, yielding radiometric ages in the 1.4–1.2 Ga range, have been reported from eastern Poland, the Trans-Scandinavian magmatic belt, and the Urals (Pozaryski, 1977; Semikhatov et al., 1991; Milanovsky et al., 1994).

4.2. *Middle Riphean (1350–1050 Ma)*

A new rifting cycle apparently commenced during the mid-Middle Riphean (Fig. 7b), giving rise

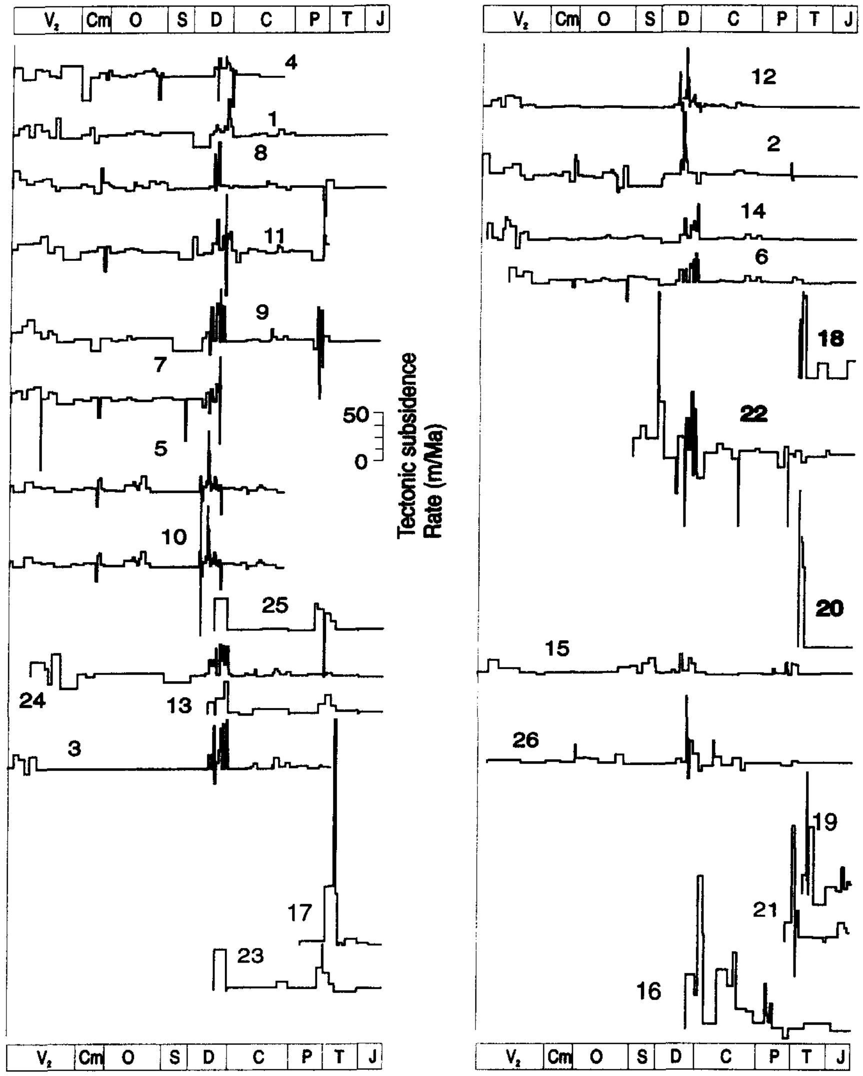


Fig. 6. Comparison of tectonic subsidence rates for selected wells from the East European Craton. Note correlation between main subsidence phases. For well locations see Fig. 2. Names of wells are listed on Fig. 5.

to the subsidence of the deep Tornquist (Denmark–Poland–Dobrogea) and the Peri-Timan–Peri-Ural rifts along the present-day western and eastern margins of Baltica, respectively. At the same time the interior of Baltica was transected by the Mid-Russian, Kandalaksha-Dvina, and Gulf of Bothnia system of aulacogens and the Ladoga rift. Sediments accumulating in the peri-cratonic rift basins consist of continental to shallow-water terrigenous clastics which grade laterally into very thick carbonates. In contrast, the interior rifts are characterized by continental deposits. Subsidence of the Ladoga, Kandalaksha and Krestets basins was accompanied by volcanism and the intrusion of plutons. Contemporaneous dike swarms occur on the Baltic Shield. There is also some evidence for basic plutonism in the Peri-Uralian basins (Lagutenkova and Chepikova, 1982).

In view of the sedimentary record of these basins, it is postulated that during the Middle Riphean rifting cycle, probably as part of a general plate reorganisation, continental fragments were separated from the western and eastern margins of Baltica, resulting in the development of new passive margins and abortion of the intracratonic rifts.

At the transition from the Middle to the Late Riphean, possibly the entire craton was affected by regional compression, as evident by the inversion of the Middle Riphean rifts and the development of the Dalslandian orogen along the western margin of Baltica. This reflects a further major change in the megatectonic setting of Baltica. Traces of the Dalslandian orogen are recognised in western Norway, southern Sweden and possibly also in the Dobrogea (Garetsky, 1990). Angular unconformities between Middle and Late Riphean series are also evident in the Uralian domain from which low-grade metamorphic and plutonic rocks have been reported (Keller and Chumakov, 1983; Semikhatov et al., 1991). Also in the interior rifts, a major break in sedimentation straddles the Middle–Late Riphean boundary (Fig. 3).

These deformations are indicative of a major change in the megatectonic setting of Baltica. During the Dalslandian orogenic cycle, which spanned 1.2–0.9 Ga and is the time equivalent to the Grenvillian orogeny of North America (Gorbatschev, 1985; Gorbatchev and Bogdanova, 1993), Baltica was welded along Grenvillian–Dalslandian sutures in the west to Laurentia–Greenland and in the south to Amazonia and thus was incorporated in the Late Riphean Pangea (Hoffman, 1991; Dalziel et al., 1992).

4.3. Late Riphean (1050–650 Ma)

Instability of the post-Grenvillian Pangea is reflected in Baltica by the Late Riphean rifting cycle which affected mainly the eastern margin and the eastern-central parts of the EEC. Main elements of this rift system are the pericratonic Peri-Timan and Peri-Ural basins and the interior Pachelma aulacogen, the Moscow rift belt and the Kandalaksha aulacogen (Fig. 7c). Contemporaneous dike swarms occur on the Baltic Shield (Svetov et al., 1990; Scheglov et al., 1993). Significantly, the early Riphean Abdulino and Pachelma and the Middle Riphean Kandalaksha and Ladoga aulacogens were tensionally reactivated during the Late Riphean rifting cycle, whereas other older rifts remained quiescent.

As the Peri-Timan and Peri-Ural basins have the characteristics of rifted passive margins, it is assumed that during the Late Riphean yet another crustal block was separated from the eastern margin of Baltica. Whereas the interior rifts are characterised by predominantly continental sediments, marine strata, dominated by carbonates, characterise the peri-cratonic rifts (Getsen, 1991; Milanovsky et al., 1994).

A regional pre-Vendian unconformity is evident on the entire Baltica Craton (Fig. 3). As this end-Riphean hiatus is not associated with compressional deformations and magmatism, it is likely to be of a eustatic origin related to early Vendian glaciation (see below).

Fig. 8 summarises the Riphean tectonic history and drift patterns of the EEC. Riphean aulacogens contain a considerable amount of volcanic rocks. There are numerous Riphean dike swarms (Milanovsky et al., 1994; Grachev et al., 1994). Geochemical data show that Riphean rift-related volcanic rocks are basalts, trending towards alkali-basalts, typical for intraplate magmatism (Grachev et al., 1994).

4.4. Vendian (650–540 Ma)

During the Vendian, rifting activity along the western and southwestern margin of Baltica testifies to increasing instability of the Late Riphean Pangea (Hoffman, 1991; Andréasson, 1994).

On the EEC, the lower parts of the Vendian are subdivided into the Laplandian glacial and the Volyn flood basalt stages. The upper parts of the Vendian are represented by extensive continental and shallow-marine sediments; these cover much of the craton and form part of a single depositional sequence which extends into the earliest Cambrian.

In large parts of Baltica, the oldest Vendian sediments were deposited during the Laplandian stage which coincides with a major global glaciation (Hambrey and Harland, 1981). Laplandian glaciers covered much of Baltica, as indicated by remnants of glacial deposits found along its margins and also in its interior (Fig. 7d). These often attain sizeable thicknesses (Sokolov and Ivanovsky, 1985; Sokolov and Fedonkin, 1985), evidencing significant glacial erosion, perhaps of the order of hundreds of metres and up to 1 km. Glacial erosion deeply truncated the Riphean and possibly also earliest Vendian sediments which had been deposited in the interior

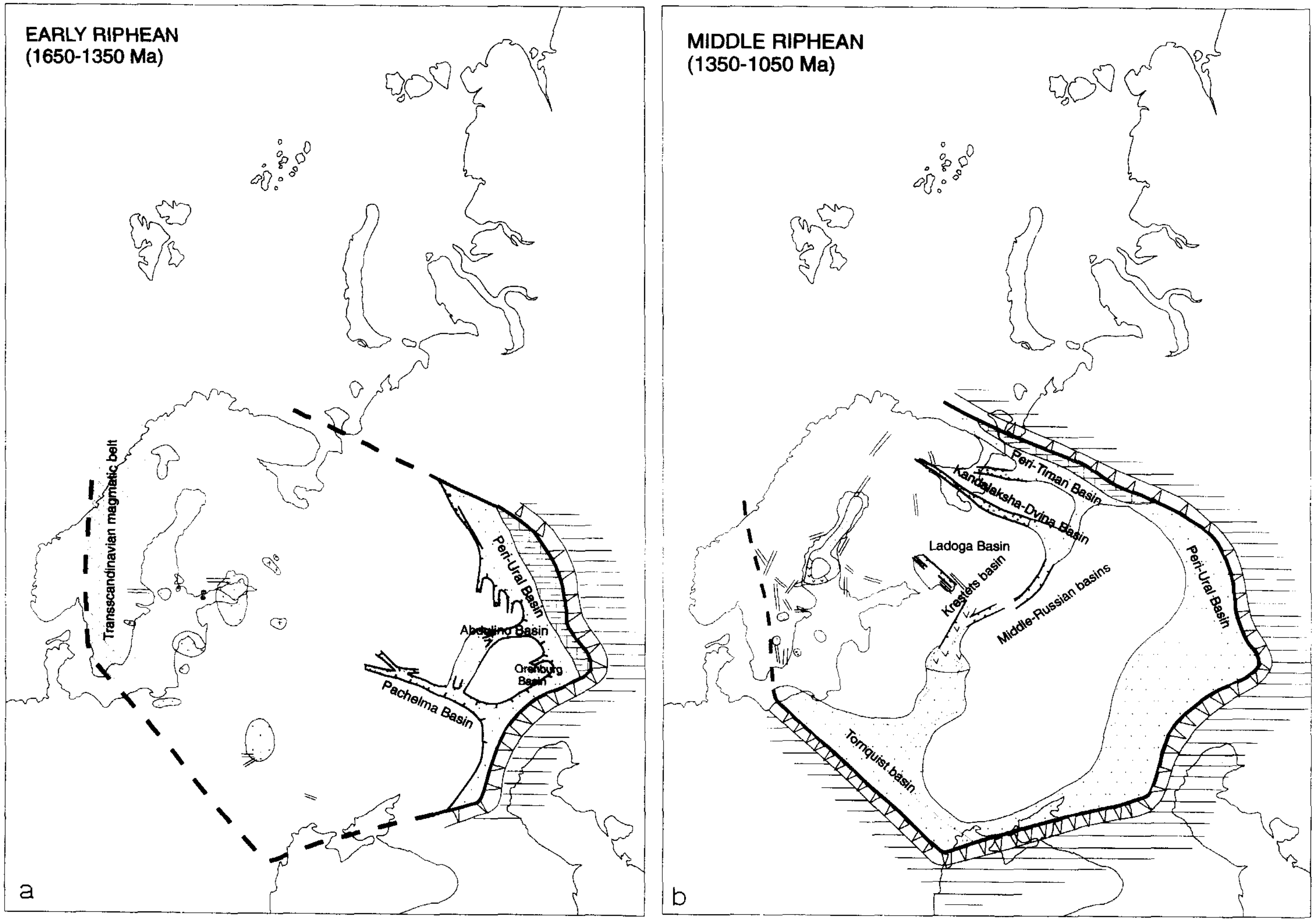
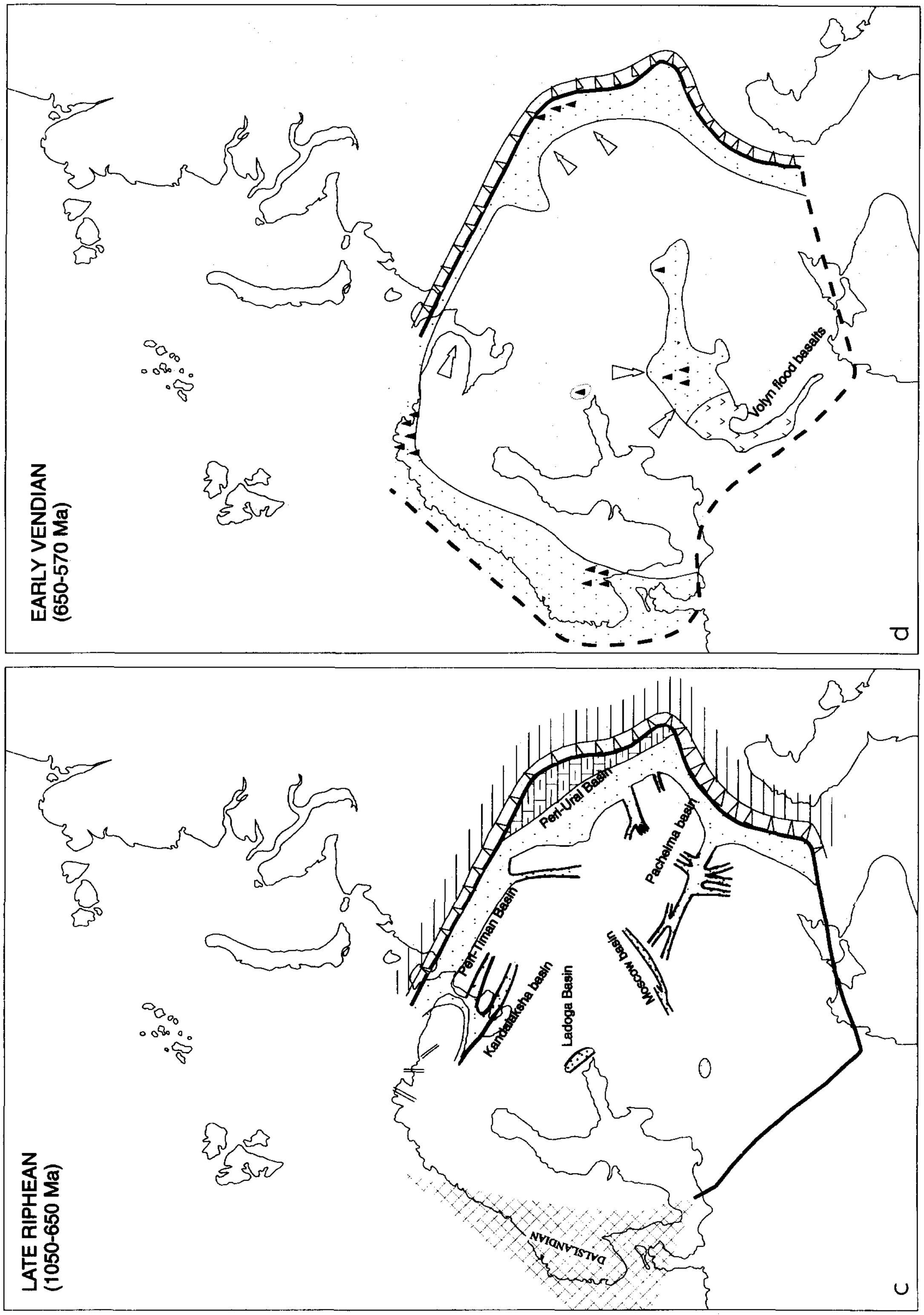


Fig. 7. Palaeotectonic/palaeogeographic maps of the East European Craton. Legend: *1* = continental sands; 2 = continental sands and shales; *3* = tillites (Early Vendian); *4* = alluvial-deltaic and shallow-marine, mainly sands; *5* = alluvial-deltaic and shallow-marine sands and shales; *6* = shallow-marine sands and shales; *7* = alluvial-deltaic and shallow-marine sands and shales (for Precambrian and earliest Cambrian only); *8* = shallow-marine sands, shales and carbonates; *9* = shallow-marine carbonates and shales; *10* = mainly carbonates; *11* = carbonates, mainly coral and/or algal; *12* = carbonates and evaporites; *13* = mainly evaporites; *14* = deeper-marine carbonates, clays and siliceous shales; *15* = deeper-marine clays and siliceous shales; *16* = deeper-marine clastics and/or carbonates; *17* = turbiditic series, flysch; *18* = plateau basalts; *19* = acid volcanites and clastics; *20* = granite intrusions (for Early Riphean); *21* = oceanic basin; *22* = active fold belts; *23* = inactive fold belts; *24* = cratonic highs; *25* = boundaries of the craton and main tectonic units; *26* = major active faults; *27* = spreading axes; *28* = subduction zones; *29* = inversion axes; *30* = dyke systems (Precambrian); *31* = continental slope; *32* = rifts; *33* = highly stretched continental or oceanic crust; *34* = active major thrusts; *35* = boundaries of lithological zones; *36* = erosional edge of mapping interval; *37* = directions of clastic influx; *38* = orogenic volcanism; *39* = basaltic volcanism; *40* = unknown continental terrane.



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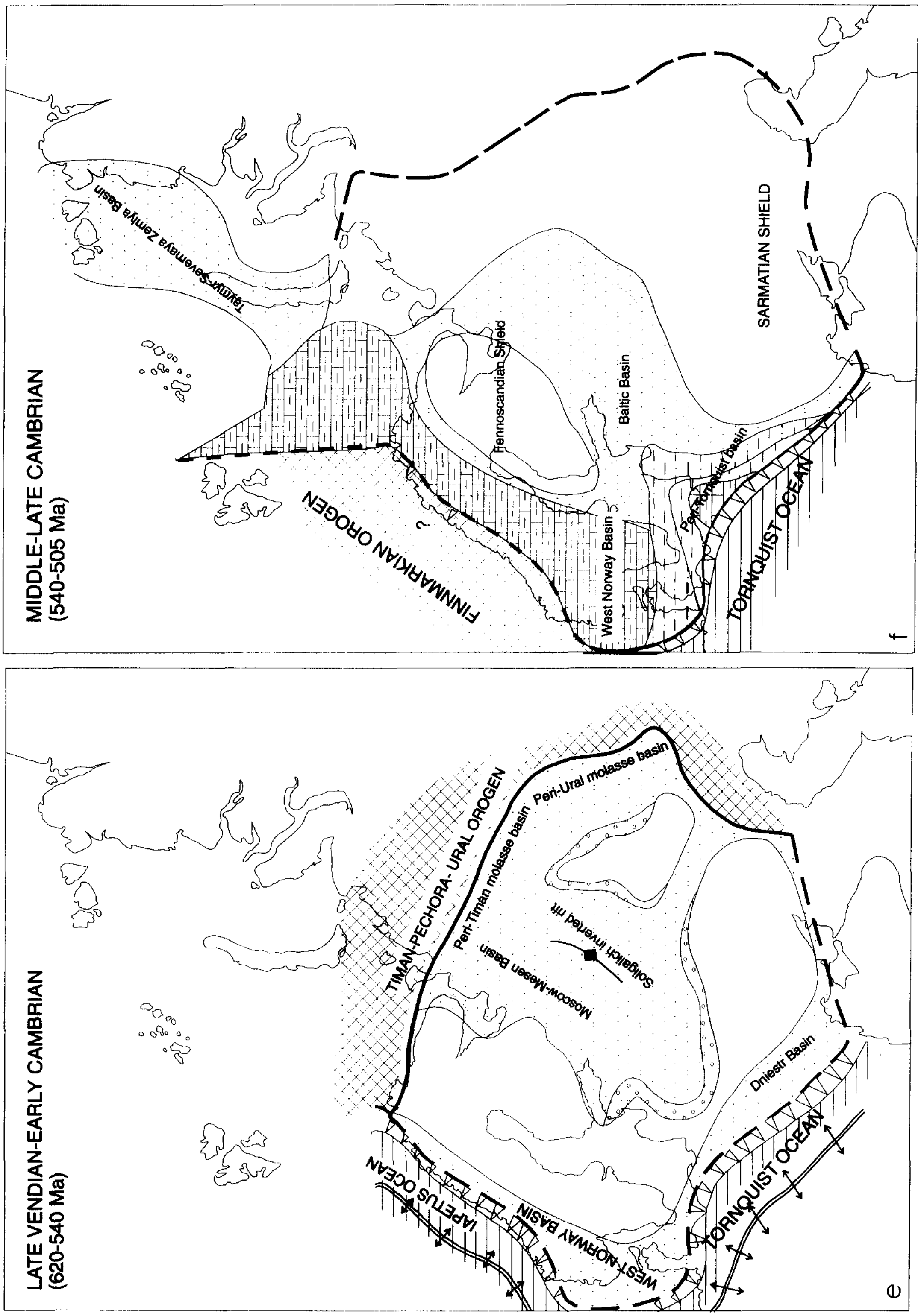

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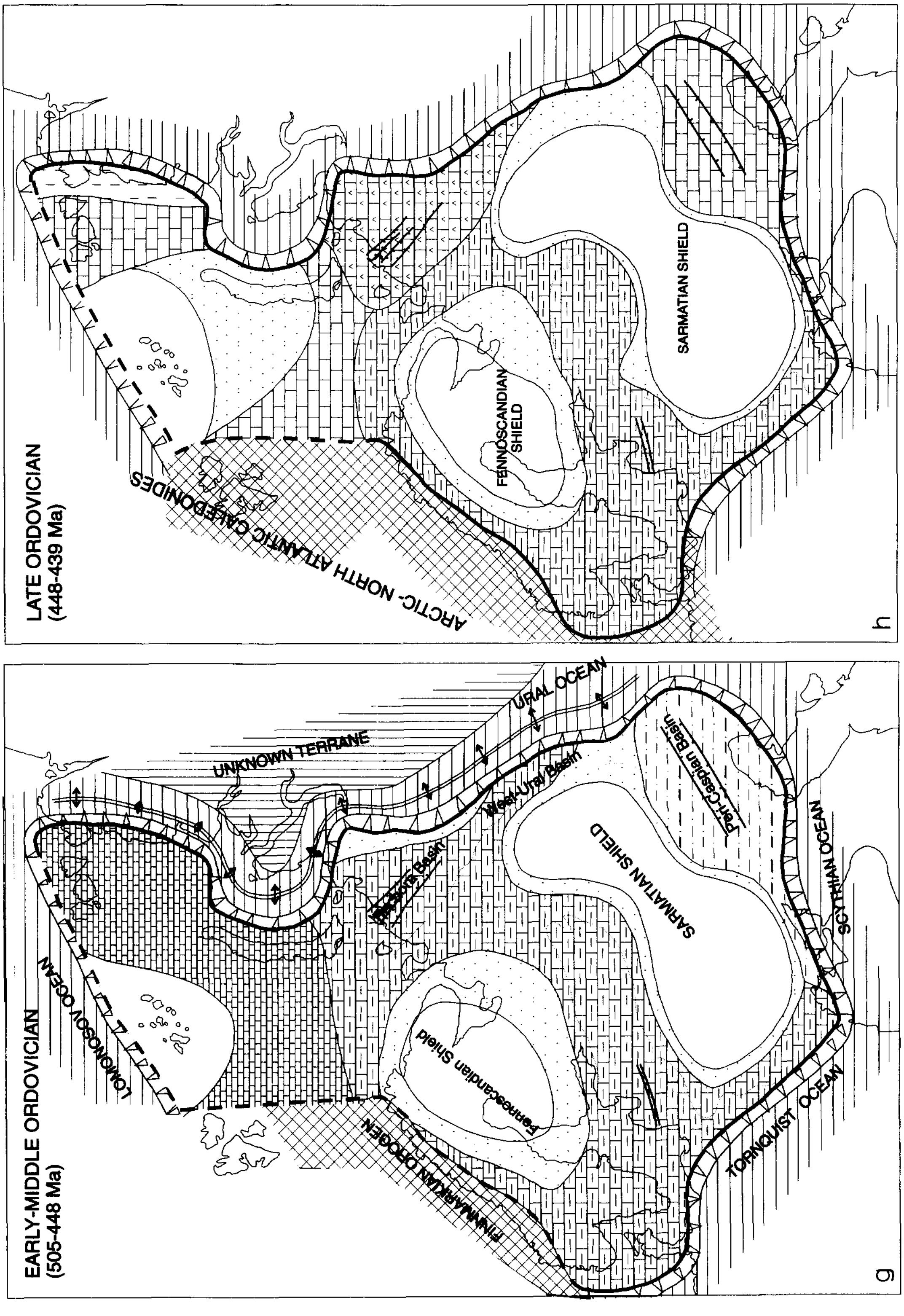

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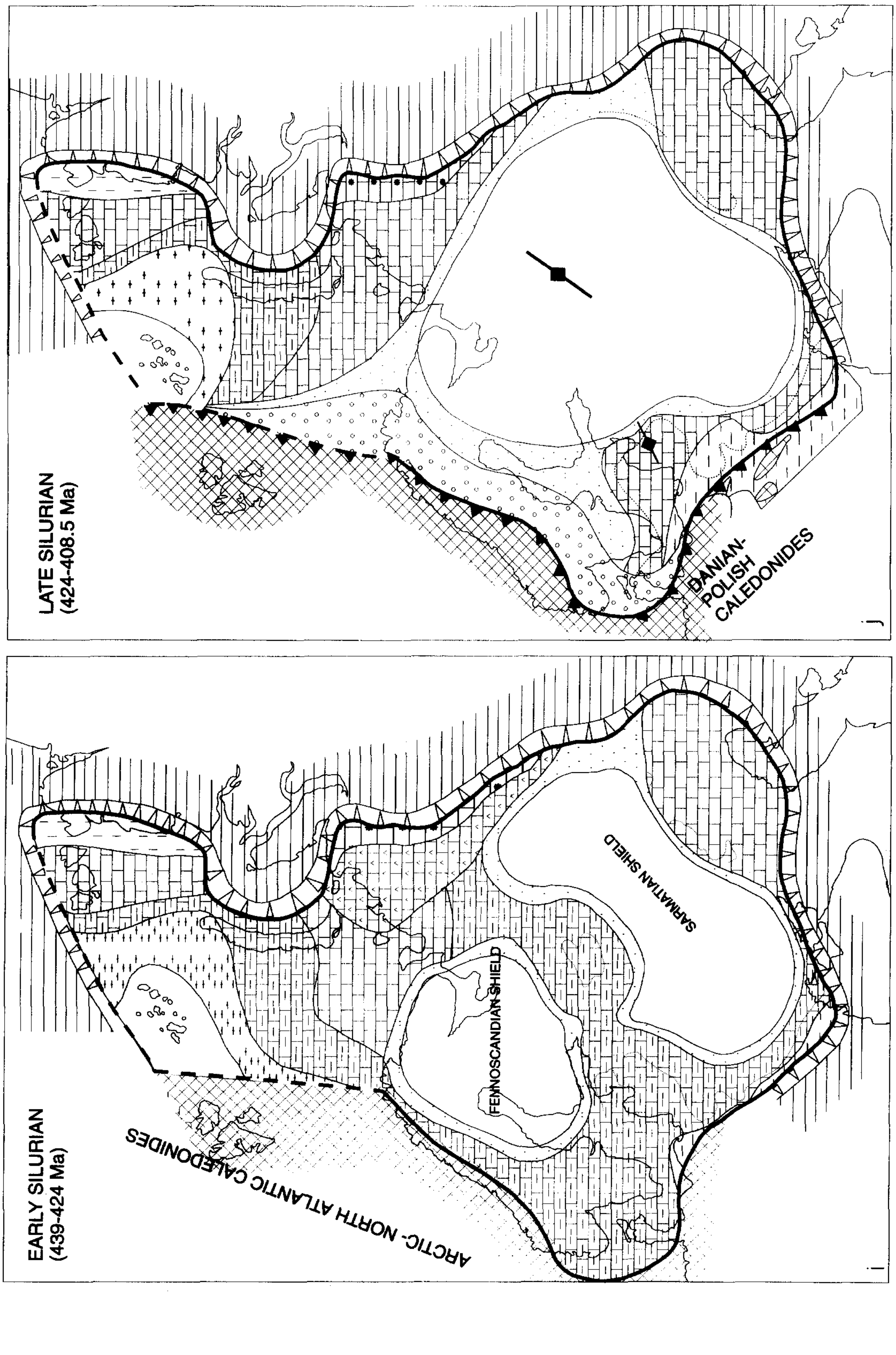

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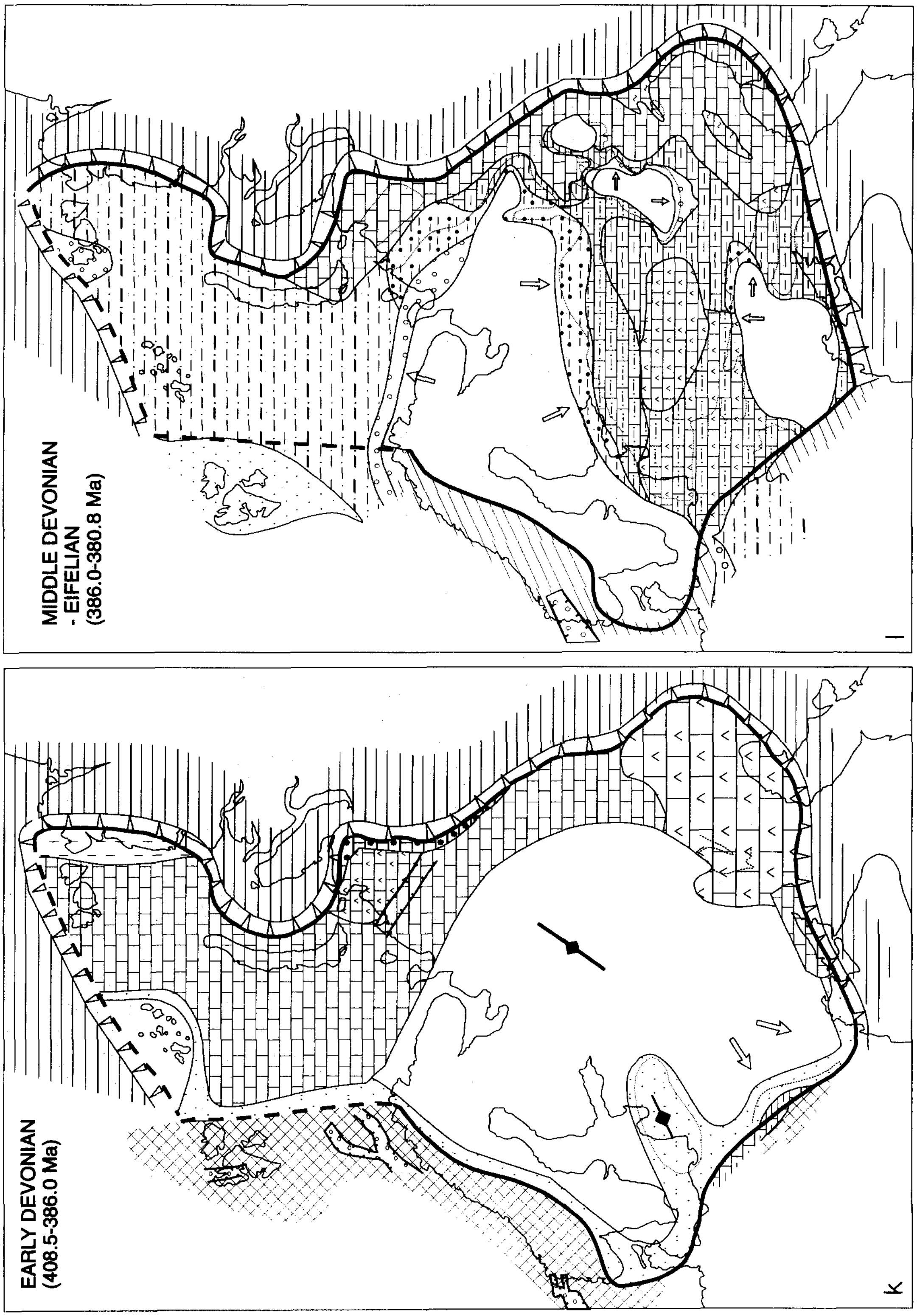

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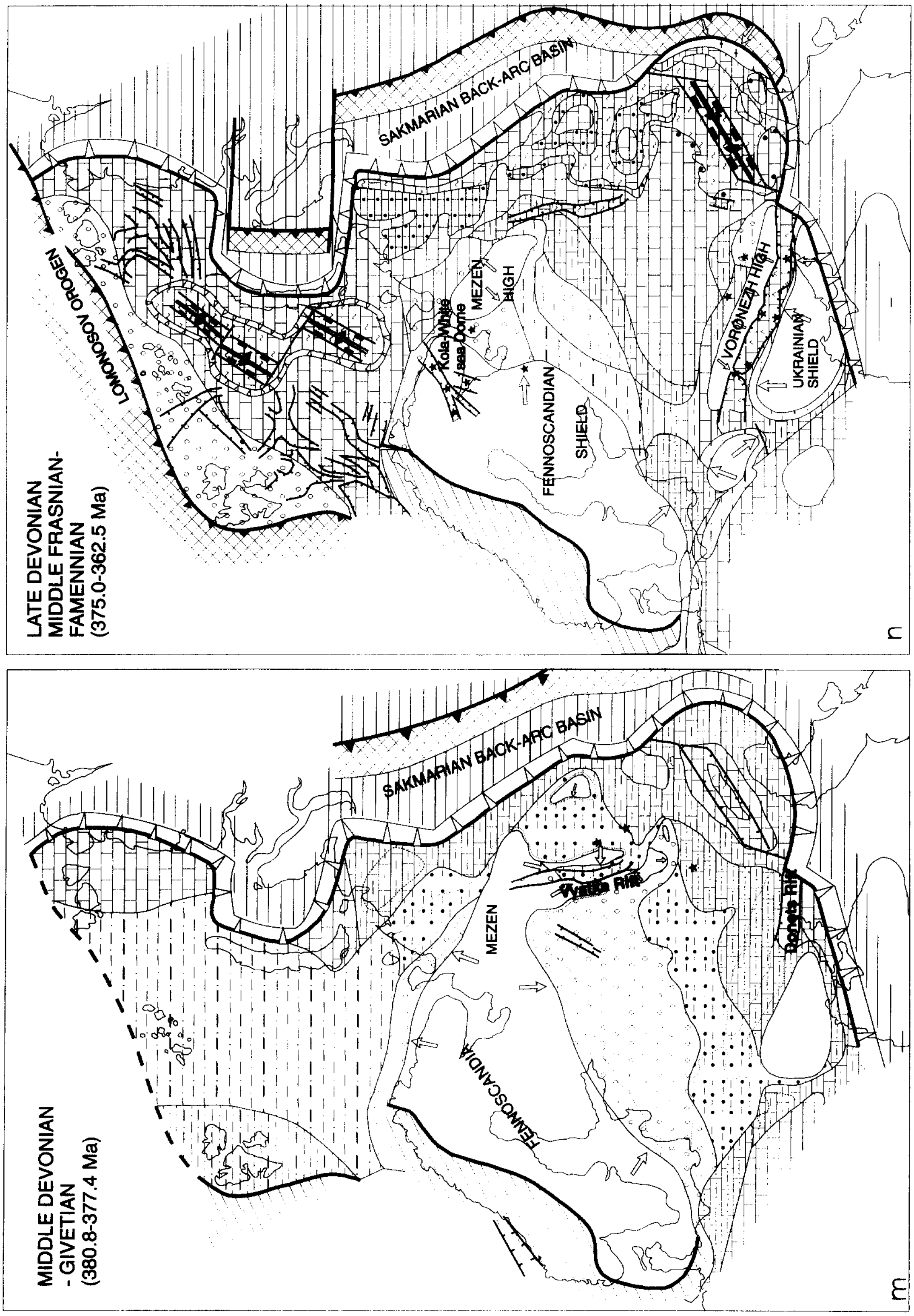

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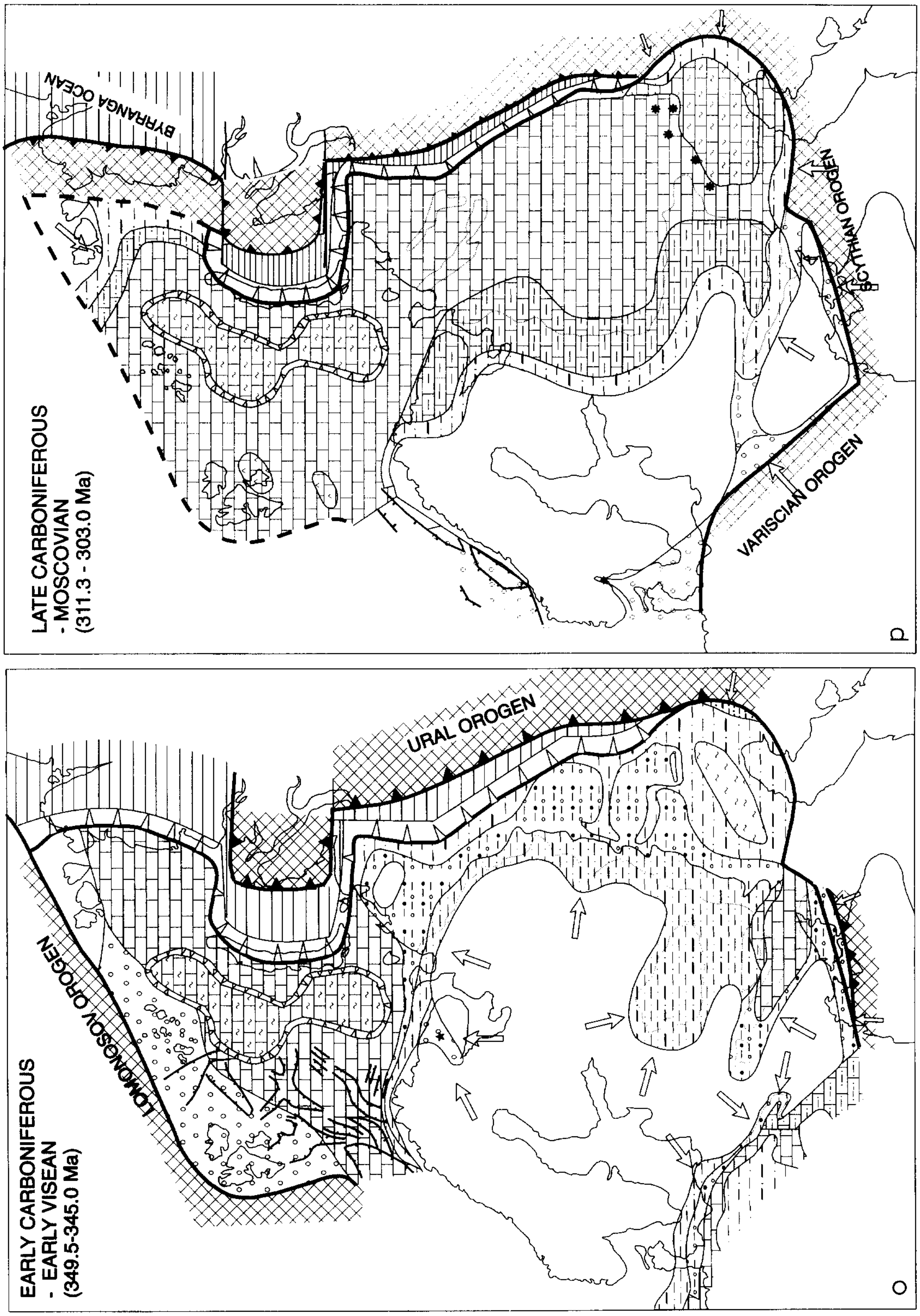

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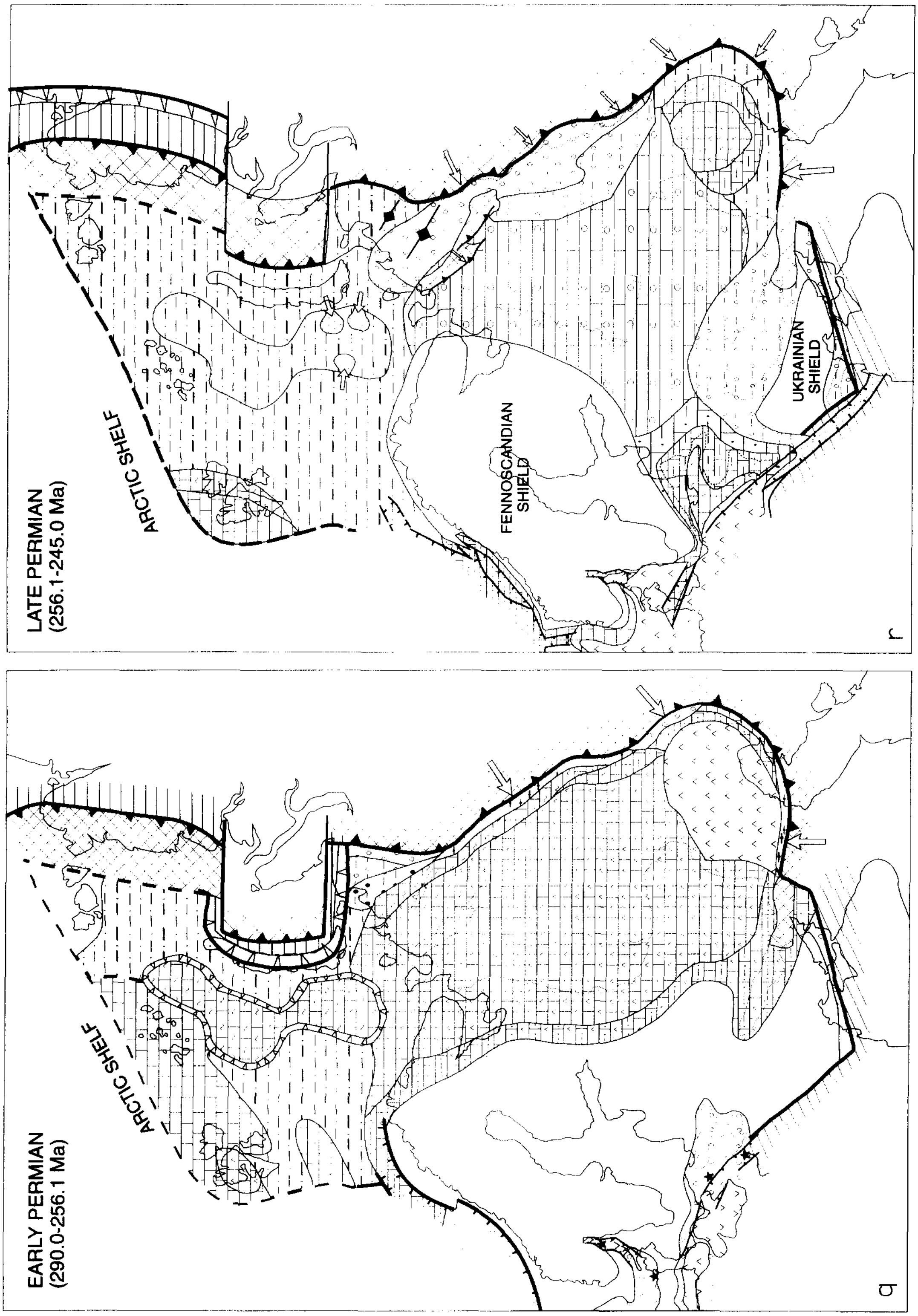

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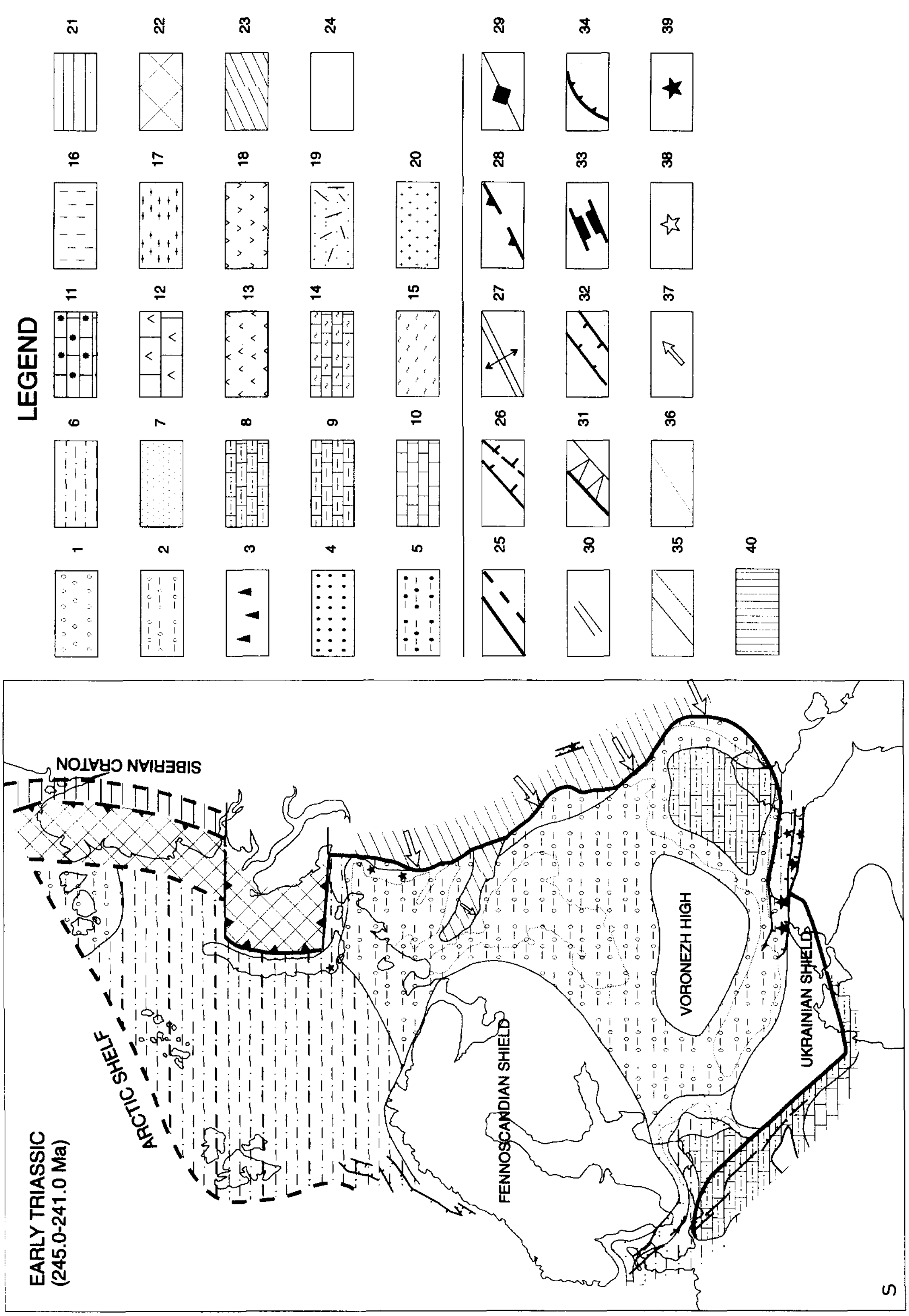

Fig. 7 (continued).

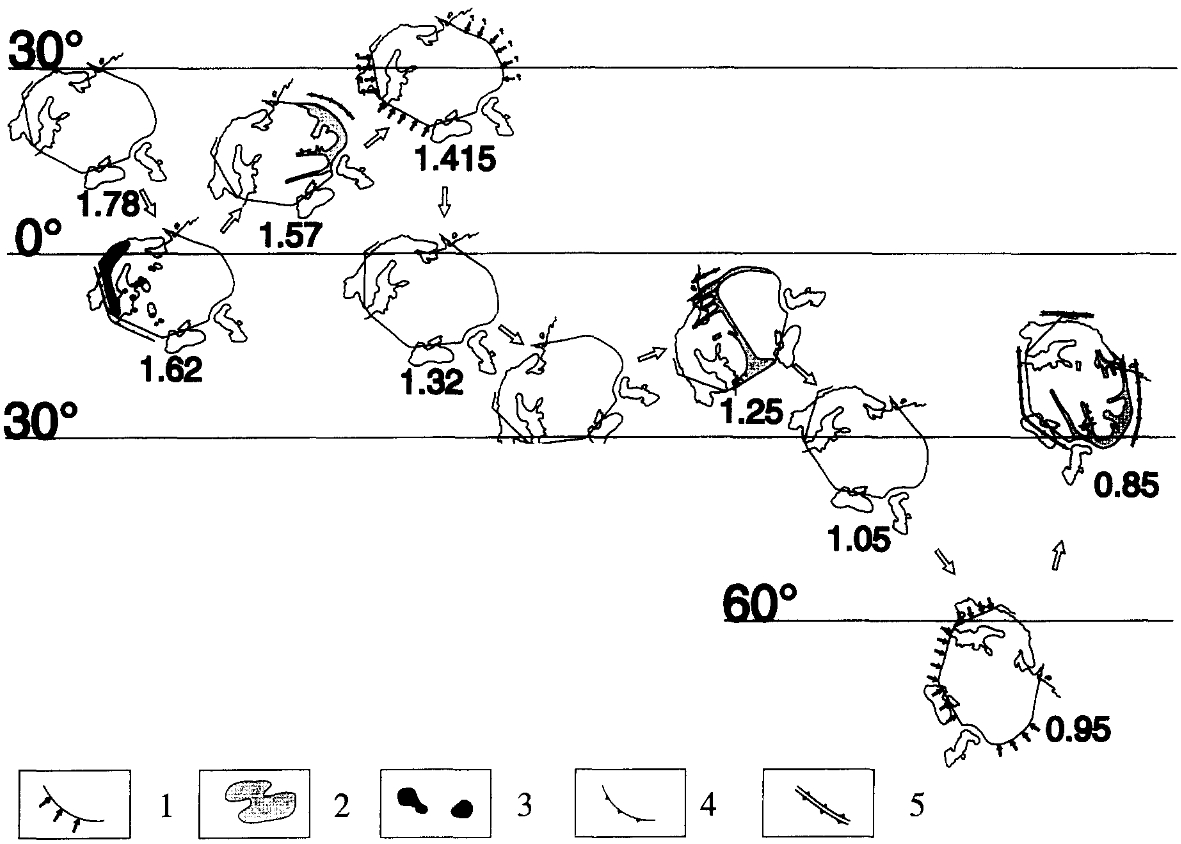

Fig. 8. Riphean drift pattern of Baltica (after Posenen et al., 1989) and reconstructions of the palaeotectonic environments. Legend: *1* = orogenic belt; *2* = intracratonic basin; *3* = rapakivi granites and Trans-Scandinavian magmatic belt; *4* = subduction zone; *5* = spreading zone.

of Baltica. In this context it should be noted that, on the Norwegian margin of Baltica, Laplandian tillites are preceded by a thick Vendian succession of continental, shallow-marine and in part even turbiditic sediments that were apparently deposited in transtensional basins (Nystuen and Siedlika, 1988; Winchester, 1988).

Laplandian sediments cover Riphean rifted basins mainly in the southwestern parts of the EEC and in the Peri-Ural Basin. The outlines of these lower Vendian basins differ significantly from those of the Riphean aulacogens. Evidence of early Vendian rifting in the interior of Baltica comes from southern Sweden (Winchester, 1988) and may have controlled on a much wider scale the distribution of interior basins; however, this is very difficult to ascertain on the basis of the available data.

During the Volyn stage, the large Volyn flood basalt province developed along the southwestern margin of the EEC (Volovnin, 1975; Znamenskaya et al., 1990; Garetsky and Zinovenko, 1994). These subaerial flood basalts attain maximum thicknesses of 500 m. In areas where basalts are associated with fault zones they are devoid of tuffs, whilst towards the basin margins pyroclastics dominate (Bronguleev, 1985; Garetsky and Zinovenko, 1994). The bulk of the flows consists of basalts with subordinate andesites and dacites.

The tectonic setting of the Volyn basalt province can be compared with that of the Mesozoic Parana–Etendeka and Karoo and the Palaeogene Deccan flood basalts. This suggests that the Volyn flood basalt province developed during the rifting stage which preceded the opening of the Tornquist Ocean. This concept is compatible with contemporaneous rifting and magmatic activity in the Arctic–North

Atlantic domain (Winchester, 1988; Andréasson, 1994). Rift-like early Vendian basins, associated with alkaline olivine- and trachy-basalts, occur also along the eastern margin of Baltica in the middle Urals (Maslov and Ivanov, 1995).

4.5. Late Vendian–Early Cambrian (620–530 Ma)

The boundary between the lower and upper Vendian series is of great significance as it marks the onset of a regional transgression, resulting in the establishment of very large shelf basins covering much of Baltica (Fig. 7e). This reflects a major reorganisation of regional subsidence patterns to which a number of factors may have contributed. During late Vendian and Cambrian times sea-levels rose eustatically (Vail et al., 1977), partly in response to deglaciation and partly in response to the opening of new oceanic basins during the break-up of the Late Riphean–early Vendian Pangea.

During late Vendian–Early Cambrian times, the poorly known orogenic Pechora–Barentsia–Urals terrane began to collide with the eastern margin of Baltica (Fig. 7e; Baikalian orogeny). In contrast, rifting in the North Atlantic–Central European domain accelerated during the late Vendian and culminated at the transition to the Cambrian in the separation of Baltica from the Riphean Pangea and the opening of the Iapetus and Tornquist oceans (Winchester, 1988; Kamo et al., 1989; Hoffman, 1991).

The four main late Vendian–Early Cambrian sedimentary basins of Baltica are the West-Norway, Dniestr (Lvov–Kishinev), Moscow–Mezen and the Peri-Timan–Peri-Ural basins (Fig. 7e). As the northwestern margin of the Moscow–Mezen Basin and the eastern margin of the Norway Basin are erosional, it cannot be excluded that much of the Fennoscandian Shield was transgressed in late Vendian–Early Cambrian times; this concept is supported by the occurrence of late Vendian erosional remnants in the Gulf of Botnia.

The rifted Dniestr and West-Norway basins, which are characterised by regional transgressions, developed during the Early Cambrian into passive continental margins. In Norway, shallow-marine clastics grade laterally into carbonates. In the Dniestr Basin, shallow-marine clastics overstepped the Volyn flood basalts.

The Moscow–Mezen Basin rests on the Mid-Russian–Kandalaksha–Dvina Riphean rift system and has the configuration of a broad post-rift sag basin. However, there is a great time gap between the end of Middle Riphean rifting and the beginning of post-rift subsidence. Unless the available dating of the Riphean sequences is erroneous, indicating dates that are far too old, other factors must be invoked to explain the apparent time gap between syn- and post-rift sedimentation. The fill of the Moscow–Mezen Basin consists of shallow-marine and continental clastics, deposited under a cold climate (Bronguleev, 1985; Sokolov and Fedonkin, 1985).

The Peri-Timan–Peri-Ural Basin is a typical foreland basin that developed in front of the collisional Timan–Pechora–Urals orogen. It overlies a Late Riphean–early Vendian passive margin sedimentary prism. Thin late Vendian tuff horizons, occurring in the Moscow–Mezen and Peri-Timan basins (Sokolov and Fedonkin, 1985; Yakobson and Nikulin, 1985), may be related to volcanic activity in the evolving Pechora–Ural fold belt.

The late Vendian–Early Cambrian depositional cycle terminated with the mid-Early Cambrian uplift of Baltica that was accompanied by weak inversion of the Soligalich aulacogen of the Mid-Russian aulacogen belt (Milanovsky, 1987; Garetsky, 1990; Kuzmenko et al., 1991). Uplift affected the eastern part of Baltica first (Milanovsky, 1987), whereas the outlines and structures of the more western basins changed (Garetsky, 1990). This so-called Soligalich event, which was accompanied by a regional regression (Figs. 4 and 6), coincides with the climax of orogenic activity in the Timan–Pechora–Urals fold belt (along which the continental Pechora–Barentsia blocks were sutured to Baltica; Puchkov, 1995), and with possible transtensional deformations along the Tornquist zone during the oblique opening of the Tornquist Ocean.

4.6. Late Early Cambrian–Early Devonian (530–386 Ma)

During Cambrian, and certainly during Ordovician to end-Silurian times, Baltica was a separate plate. However, following the Early Cambrian opening of the Tornquist and Iapetus oceans, the drift pattern of Baltica changed and it began to converge during Middle and Late Cambrian times with Laurentia–Greenland to which it was sutured at the end of the Silurian along the Arctic–North Atlantic Caledonides. With this, Baltica was incorporated into the Laurussian mega-continent (Ziegler, 1989; Torsvik et al., 1992).

In comparison with the distribution of late Vendian–earliest Cambrian basins, the pattern of late Early Cambrian to Early Devonian basins reflects again a radical change that must be related to a fundamental reorganisation of plate interactions (Fig. 7e–k). Along the eastern margin of Baltica, Baikalian orogenic activity ceased and gave way to crustal extension. In contrast, along its western margin, development of the Caledonian orogen commenced during the Late Cambrian and persisted into earliest Devonian times (Gee and Sturt, 1985; Harris and Fettes, 1988; Andréasson, 1994).

The main late Early Cambrian to Early Devonian sedimentary basins of Baltica are, along its western margin, the Baltic–Peri-Tornquist, West-Norway and Barents Sea basins, and, along its eastern margin, the Taymyr–Severnaya Zemlya, West-Urals and the Peri-Caspian basins. Rising sea-level accounted for progressive overstepping of basin margins so that by Late Ordovician–Early Silurian times only the central parts of the Sarmatian and possibly also of the Fennoscandian shields formed persisting land masses (Fig. 7h, i).

4.6.1. Western margin of Baltica

The Early to Middle Cambrian evolution of the western margin of Baltica was governed by progressive opening of the Iapetus and Tornquist oceans. However, during the Middle and Late Cambrian, these oceans began to close again. The Caledonian orogenic cycle, spanning Late Cambrian to earliest Devonian times, culminated in the suturing of Laurentia–Greenland and Baltica along the Arctic–North Atlantic Caledonides and the accretion of Gondwana-derived terranes to the southwestern margin of Baltica. These terranes are enveloped in the North German–Polish and the Mid-European Caledonides (Ziegler, 1988, 1989, 1990). In the course of the Caledonian orogeny, the western and southwestern passive margins of Baltica were destroyed and foreland basins superimposed on their proximal parts during the Silurian (Fig. 7j). Correspondingly, available stratigraphic information is restricted to tectonic and erosional remnants of these formerly very large shelf and foreland basins.

In the Peri-Tornquist–Baltic Basin, sedimentation resumed during the late Early to Late Cambrian with transgressive shallow-marine sands overlain by shales, reflecting increasing water-depths (Garetsky, 1981, 1990). A large embayment occupied the Baltic area from where a sea arm extended northeastwards through the White Sea into the Barents Sea. During the Ordovician and Early Silurian (Fig. 7g–i), carbonate-dominated shelves developed, giving way laterally to deeper-water graptolitic shales. There is only minor evidence for syn-depositional tectonic movements; for instance, Early Ordovician mild tensional tectonics may have controlled the development of the linear depocentre of the Baltic Basin (Myannil, 1965; Kaplan and Suveizis, 1970; Rotenfeld et al., 1974; Geodekyan et al., 1978; Floden, 1980; Suveizis, 1982). Late Silurian flysch series, derived from the rising Caledonides, were deposited in a foreland basin. In Poland, the foreland basin entered a shallow-water and ultimately a continental stage only during the mid-Early Devonian, whereas in southern Norway, Late Silurian series (Fig. 7j) are already developed in a continental red-bed facies (Garetsky, 1990; Ziegler, 1990).

In Norway and western Sweden, the main phases of the Caledonian orogenic cycle include the Mid-Cambrian to Early Ordovician Finnmarkian and the Silurian Main-Scandinavian orogenic pulses during which major nappe systems were emplaced on the margin of Baltica (Gee and Sturt, 1985; Andréasson, 1994). The foreland basin of the Scandinavian Caledonides was destroyed by erosion in Late Permian–early Mesozoic times (see below). The northward continuation of the Caledonides into the area of the western Barents Sea is suggested by results of deep reflection-seismic surveys (Gundlaugsson et al., 1987). Results of recent surface geological studies indicate that the Caledonian deformation front is located to the east of the Spitsbergen Archipelago (Gee and Page, 1994; Gee et al., 1995). For the western Barents Sea, the maps presented in Fig. 7f–k must be regarded as conceptual as there is virtually no stratigraphic information available on Cambrian to earliest Devonian series.

4.6.2. Eastern margin of Baltica

In contrast to the collisional setting of the western margin of Baltica, the evolution of its eastern margin was governed by tensional tectonics. These can be unravelled only in limited areas.

Cambrian flysch, unconformably covered by an Ordovician carbonate platform has been reported from the Taymyr–Severnaya Zemlya–Novaya Zemlya region (Uflyand et al., 1991; Kogaro et al., 1992). The tectonic position and origin of this flysch basin is unknown. It is assumed that collision between the Pechora–Barentsia and the North-Taymyr–Severnaya Zemlya terrane had terminated at the Cambrian–Ordovician boundary. The position of this collisional belt is unknown and its relation to the contemporaneous Finnmarkian orogen of Scandinavia is still an open question. Thus, it is inferred that the North-Taymyr–Severnaya Zemlya region formed part of the Pechora–Barents Sea area since Ordovician times.

On Novaya Zemlya, Cambrian to Early Devonian strata consist of shallow- to deep-water sediments, deposited in rifted basins (Kogaro et al., 1992). These basins were incorporated into the passive margin of the Ural palaeo-ocean (Zonenshain et al., 1993).

Surface geological data show that a Late Cambrian rifting event occurred along the North-Polar Urals and Pay-Khoy belts (Belyaev, 1994). At about the same time, thin sediments were deposited in the Pechora Basin (Dmitrovskaya, 1990). These are overlain by Ordovician and Silurian, mainly shallow-marine carbonates (Borisov, 1986; Beliakov, 1994; Kostyuchenko, 1994). Early Ordovician rifting is evident in the Pechora–Kolva Basin. In the Polar Ural–Pay-Khoy belt, rifting persisted during the Tremadoc–early Arenig and culminated in late Arenig separation of a continental terrane from Baltica, the onset of sea-floor spreading, and the opening of the Uralian Ocean (Belyaev, 1994; Savelieva and Nesbitt, 1996). Thereafter, the Pechora Basin formed part of a carbonate-dominated passive margin. During the Early Devonian (Lochkovian), a new rifting phase occurred during which the Ordovician Pechora–Kolva palaeorift was reactivated as a transtensional basin (Fig. 6; Dedeev et al., 1995; N.A. Malyshev, pers. commun., 1995). Prior to the Middle Devonian the Pechora Basin was gently uplifted and subjected to erosion; this was accompanied by minor compressional deformations (Beliakov, 1994; Dedeev et al., 1995).

It is unknown whether the East-Barents Sea area was also affected by early Palaeozoic rifting events; however, its early Palaeozoic subsidence is inferred on the basis results of deep wells drilled in the Pechora Basin and reflection-seismic data (Senin, 1993; Shipilov, 1993; Johansen et al., 1993).

In the West-Ural Basin, south of the Pechora Basin and north of the Peri-Caspian Basin, there is clear evidence for latest Cambrian–early Arenig rifting activity, accompanied by rift flank uplift that provided a clastic source (Puchkov, 1995; Maslov and Ivanov, 1995). Late Arenig onset of sea-floor spreading (Maslov and Ivanov, 1995) probably entailed the separation of continental crustal fragments from Baltica. The resulting Ordovician–Silurian passive margin shelf was apparently narrow (Puchkov, 1995; cf. Fig. 7h–j). In pre-Middle Devonian time, this shelf underwent minor uplift and erosion (Akhmetiev et al., 1993).

In the Peri-Caspian basin, only along its northern and western flanks, a few deep wells penetrate Ordovician, Silurian and Early Devonian sediments. This series consists of basal siltstones and sandstones that grade upwards into carbonates (Milanovsky, 1987; Nevolin, 1988; Dmitrovskaya, 1990; Akhmetiev et al., 1993; Chibrikova and Olli, 1993). There is a lack of information on the tectonic setting of this area; it may have formed part of the Late Cambrian–Early Ordovician rift system of the eastern Baltica margin.

4.6.3. Pre-Middle Devonian hiatus (402–388 Ma)

On Baltica, the Early Cambrian to Early Devonian sedimentary cycle terminated with a regional erosional unconformity that coincides with a global low-stand in sea-level (House, 1983; Johnson et al., 1985). Moreover, this break in sedimentation appears to be associated with broad lithospheric deformations and inversion of some of the rifted structures of the Mid-Russian aulacogen belt and the Baltic Basin. These deformations commenced during the latest Silurian and lasted until Middle Devonian times (Chaikin, 1986; Milanovsky, 1987; Kuzmenko et al., 1991). Data from the Moscow Basin demonstrate that the main uplift event took place between mid-

Lochkovian and mid-Emsian times (Alekseev et al., 1996) and thus is coeval with the uplift and transtensional deformation of the Pechora Basin (Dedeev et al., 1995; V.N. Malyshev, pers. commun., 1995).

Based on limited data, it is inferred that during the Early Devonian the area of the present Barents Sea was occupied by a broad carbonate shelf, the western margin of which was dominated by clastics derived from the Caledonides (Fig. 7k). Similar carbonate shelves occupied the passive Timan–Pechora and Uralian shelves, whereas in the Peri-Caspian area carbonates and evaporites prevailed. Broad uplift of the central and western parts of the EEC may reflect a lithospheric response to the build-up of compressional stresses during the terminal phase of the Caledonian orogeny in response to collisional coupling between the foreland and the Arctic–North Atlantic and North German–Polish Caledonides. The foreland basin of the Norwegian–Swedish Caledonides, largely destroyed during Late Permian–early Mesozoic times, is only shown conceptually on Fig. 7k.

Caledonian suturing of Baltica with Laurentia–Greenland entailed a fundamental plate reorganisation in which abandonment of the Arctic–North Atlantic subduction system and the development of a new one in the Ural Ocean played a major role. Moreover, earliest Devonian consolidation of the Arctic–North Atlantic and North German–Polish Caledonides was followed by major sinistral translations along the Arctic–North Atlantic mega-shear zone and rifting in the domain of the Mid-European Caledonides, causing subsidence of the Variscan 'geosynclinal system' of basins in a back-arc position with respect to the Palaeo-Tethys subduction zone (Ziegler, 1989, 1990).

4.7. Middle Devonian–earliest Carboniferous (386–350 Ma)

During Middle Devonian to earliest Carboniferous times, sinistral translation of Baltica and Laurentia–Greenland along the Arctic–North Atlantic mega-shear continued. At the same time, the continental Arctica terrane, which includes present-day Chukotka and the New Siberian Islands, was sutured to the northern margin of Laurentia–Greenland along the Inuitian orogen. During the Late Devonian the northern, passive margin of the Barents shelf collided with Arctica, giving rise to the development of the mainly hypothetical Lomonosov fold belt (Ziegler, 1989, 1990). The latter is reflected by the progradation of deltaic complexes, derived from northern sources, onto the northern parts of the Barents shelf carbonate platform (Fig. 7l–n). Along the southern margin of Baltica, intermittent back-arc rifting and compression governed the evolution of the Variscan 'geosynclinal system'. The intra-Ural Ocean Magnitogorsk arc–trench system paralleled the eastern margin of Baltica (Ziegler, 1989; Zonenshain et al., 1993).

Cyclically rising sea-levels accounted for Middle and Late Devonian transgressions on Baltica and the establishment of broad, carbonate-dominated shelves on the Moscow Platform and in the eastern Barents Sea (Fig. 7l–n). The distribution of Middle and Late Devonian sedimentary basins differed considerably from that of the Cambrian–earliest Devonian. The Fennoscandian Shield may have remained emergent throughout the Devonian.

During the Givetian, new rift systems developed on the eastern and southeastern parts of Baltica and during the Late Devonian rifting activity affected the entire eastern parts of the EEC from the Pripyat–Dniepr–Donets rift in the south to the eastern Barents Sea (Fig. 7m, n; Milanovsky, 1987; Gavrish, 1989; Garetsky, 1990; Nikishin et al., 1993). Figs. 9 and 10 provide a summary of tectonic activity in the late Palaeozoic rifts of the EEC; however, given the accuracy of available stratigraphic and palaeontological data, timing of the different events is only tentative.

The Devonian rifts of the EEC often cross-cut Riphean rifts but in some cases reactivated them. Detailed subsidence analyses suggest the occurrence of more or less discrete rifting pulses during the late Givetian, early, middle and late Frasnian, and the early Famennian (Fig. 6). The main rift phase is early Frasnian in age; it was coupled with the uplift of large domes and a widespread, partly kimberlitic, magmatic activity. There are considerable variations in the tectonic setting of the different rifts, the level of volcanic activity, the degree of lithospheric extension and uplift of intervening arches. Magmatic activity was not exclusively confined to rift zones (Figs. 9 and 10).

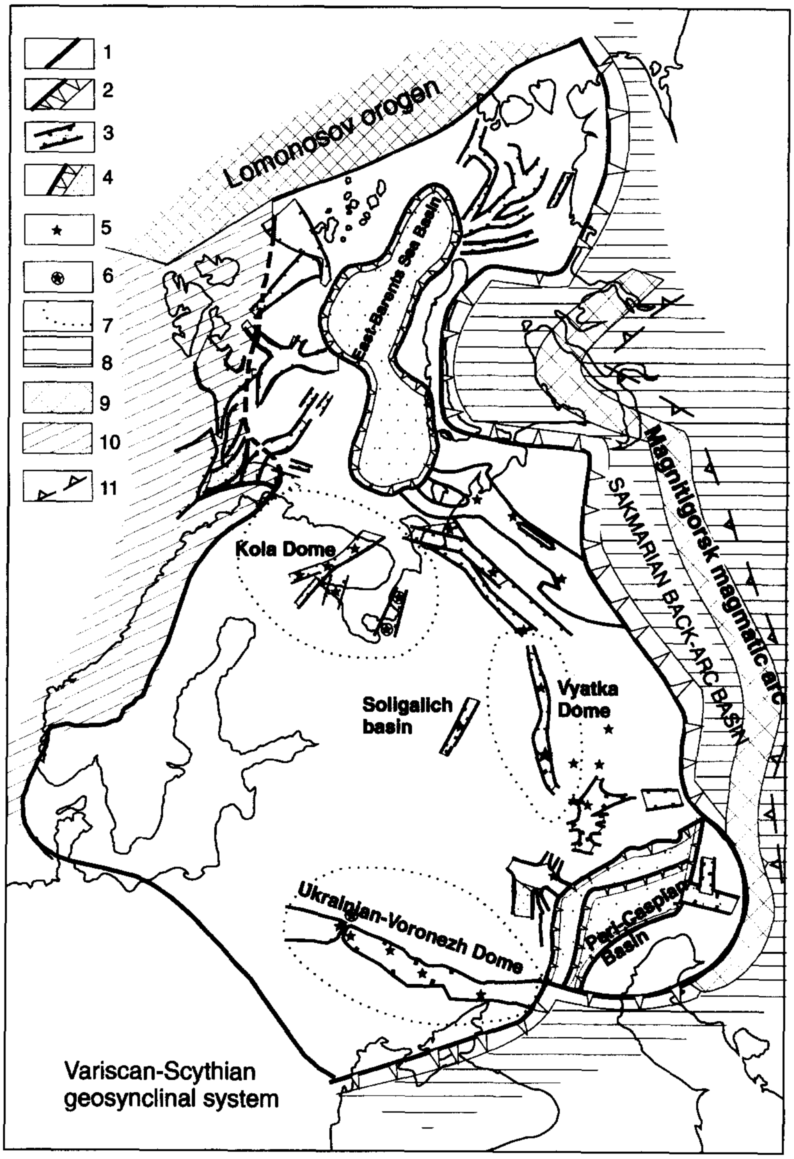

Fig. 9. Middle–Late Devonian–Early Carboniferous rift basins of the East European Craton. Legend: *1* = outlines of EEC; *2* = passive margins; *3* = rifts; *4* = area with very thin continental or oceanic crust; *5* = intraplate volcanism; *6* = kimberlites; *7* = outlines of synrift domes; *8* = oceanic basins; *9* = active fold belts; *10* = inactive fold belts; *11* = subduction zones.

4.7.1. Rifts associated with the southern margin of the EEC

Devonian–Early Carboniferous rift systems along the southern margin of Baltica consist of the back-arc basins forming the Variscan 'geosynclinal system', the some 1500 km long Pripyat–Dniepr–Donets–Donbas–Karpinsky rift (PDD), and the Peri-Caspian rift system (Fig. 9). It has been suggested that the development of the PDD and the Peri-Caspian rift system, similar to the Variscan 'geosynclinal system', is related to back-arc extension governed by the geometry of the north-dipping Palaeo-Tethys subduction zone (Ziegler, 1989, 1990).

Along the northern and western periphery of the Peri-Caspian Basin, a well documented complex system of Middle–Late Devonian rift structures occurs

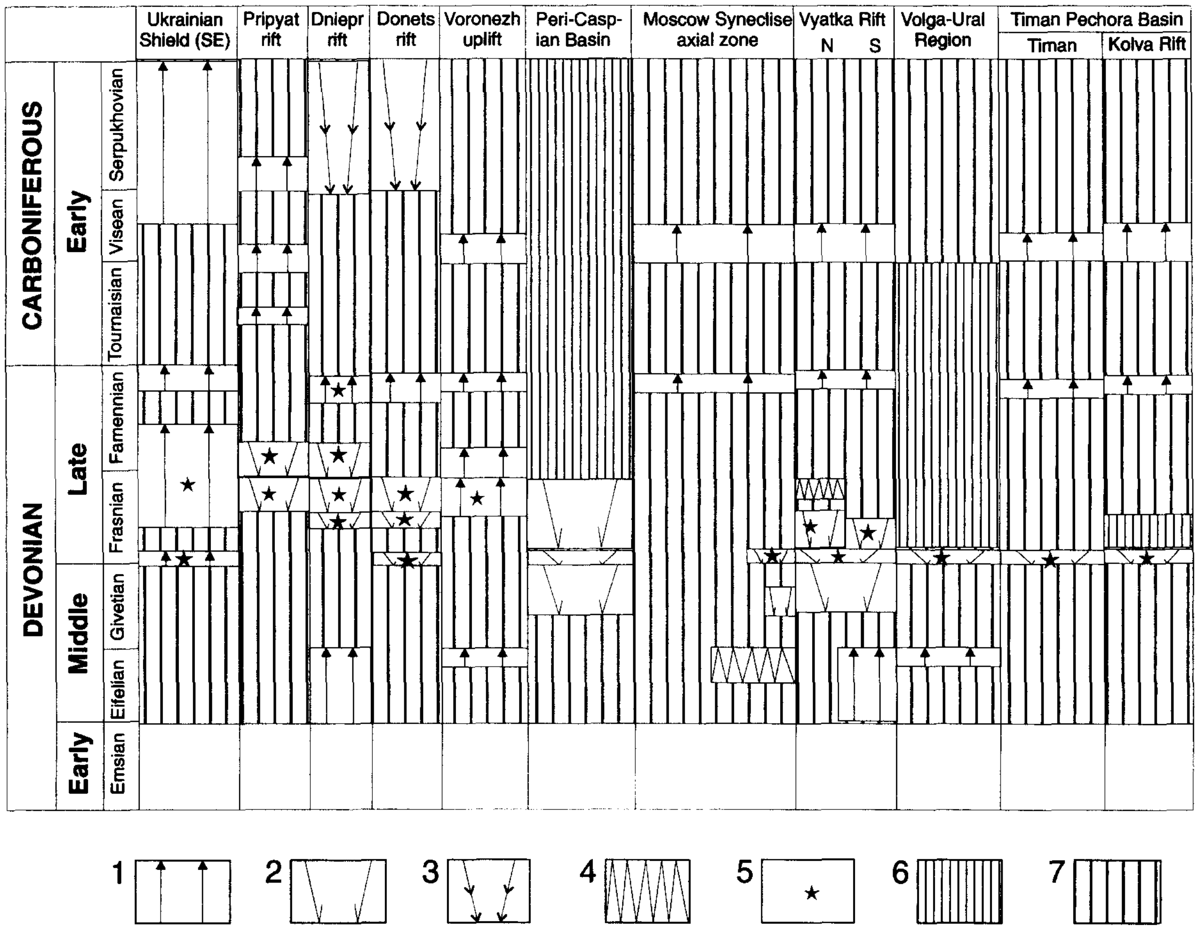

Fig. 10. Correlation of Middle Devonian–Early Carboniferous tectonic and magmatic events on the EEC. Legend: *1* = uplift episode; *2* = rifting phase; *3* = syn-compressional rapid subsidence phase; *4* = inversion event; *5* = volcanism; *6* = non-compensated subsidence; *7* = compensated subsidence.

(Fig. 9). Reflection-seismic profiles, crossing the margins of the Peri-Caspian Basin, show numerous Middle–Late Devonian normal faults (Kiryukhin et al., 1993), indicating that this basin subsided in response to crustal extension. Its present sedimentary fill of 23 km and its thin, high-velocity crust suggest that Devonian rifting may have progressed to crustal separation and the opening of a limited ocean basin (Nevolin, 1988; Zonenshain et al., 1993). The Peri-Caspian Basin shows no evidence for major syn-rift doming of its flanking areas (Milanovsky, 1987).

The PDD evolved during late Middle and Late Devonian times by northwestward rift propagation, possibly nucleating in the Peri-Caspian area (e.g., Gavrish, 1989). Prior to the onset of crustal extension, the Ukrainian Shield and Voronezh High were partially covered by marine Middle Devonian sediments (Alekseev et al., 1996). Volumetrically significant magmatism (Wilson and Lyashkevich, 1996), accompanied by a broad lithospheric doming of the rift flanks, commenced shortly after or contemporaneous with the onset of crustal extension. The main phase of rifting occurred in Frasnian–Famennian times with a moderate reactivation occurring in the Visean after a period of quiescence in the earliest Carboniferous (Stovba et al., 1996; cf. Fig. 6). The Karpinsky and Donbas segments of the PDD were subsequently heavily inverted. The sedimentary fill of the PDD decreases in thickness from 19 km in the Donets sector to about 2.5 km in the Pripyat

Trough, in tandem with a decrease in rift width and magnitude of crustal stretching (Gavrish, 1989; Garetsky, 1990; Chekunov et al., 1992; Stephenson et al., 1993; Kivshik et al., 1993; Kuzsnir et al., 1996; Stovba et al., 1996; van Wees et al., 1996).

As the level of magmatic activity in the PDD cannot be explained by crustal extension alone, mantle hot-spot activity has been invoked (Wilson, 1993; Scheglov et al., 1993; Lukjanova et al., 1994; Lyashkevich, 1994; Kuzsnir et al., 1996; Wilson and Lyashkevich, 1996). However, as the main phase of lithospheric doming of the PDD commenced after the onset of crustal extension, as shown in Fig. 10, a combination of 'passive' and 'active' rifting presumably governed its evolution.

4.7.2. *Rifts associated with the eastern margin of Baltica*

Devonian rifts occupy an almost 1000 km wide belt paralleling the 4500 km long eastern passive margin of Baltica which had developed at the end of the Early Ordovician rifting cycle. Rifting activity resumed during the Middle Devonian and persisted into Early Carboniferous times. From north to south, major elements of this rift system are the Barents Sea, Kola, Kolva, Timan, Vyatka and Soligalich rifts (Fig. 9).

The Barents Sea contains a number of long-lived rifted structures. The rifted basins of the western Barents Sea are associated with the Arctic–North Atlantic mega-shear (Ziegler, 1988, 1989). These basins are characterised by sedimentary thicknesses in excess of 10 km. Seismic lines, calibrated by wells, suggest that main rifting activity occurred during the Early to Middle Carboniferous; however, Late Devonian rifting cannot be excluded (Johansen et al., 1993; Shipilov, 1993; Nottvedt et al., 1993).

In the eastern Barents Sea, reflection-seismic data indicate the presence of a complex system of rifted basins which contains up to 18–20 km of sediments (Johansen et al., 1993; Senin, 1993; Shipilov, 1993; Verba, 1993). The lack of sufficiently deep well control impedes determination of the main rifting stages. However, in analogy with the on- and offshore Timan–Pechora area and the Kola Peninsula (Milanovsky, 1987; Kogaro et al., 1992; Scheglov et al., 1993), it is assumed that rifting activity resumed in the eastern Barents Sea during the Middle Devonian and persisted into the Early Carboniferous (Alsgaard, 1993; Junov, 1993a,b; Nikishin et al., 1993). This notion is supported by the occurrence of Middle–Late Devonian extensional structures on Novaya Zemlya, which are associated with Frasnian tholeiitic and trachytic basalts, cross-cutting Ordovician ones (Kogaro et al., 1992). Reflection-seismic data (Shipilov, 1993; Johansen et al., 1993; Ignatenko and Cheredeev, 1993; Alekhin, 1993; Popova and Krylov, 1993; Junov, 1993a,b) show that the eastern Barents Sea was occupied during the Middle Devonian to Late Carboniferous by a deep-water basin, infilling of which commenced only during the Late Permian (Fig. 11; cf. Nikishin et al., 1995). This basin, which is characterised by a thin, high-velocity crust of possibly oceanic nature (Fig. 2; Senin, 1993; Shipilov, 1993), probably subsided in response to Middle Devonian–Early Carboniferous crustal extension.

The Timan and Kolva rifts (Fig. 6), characterised by Late Devonian volcanism, do not show any evidence for syn-rift uplift of flanking arches (Milanovsky, 1987; Ehlakov et al., 1991; Belyaeva et al., 1992; Dedeev et al., 1995). The Late Devonian basalts of the Pechora–Kolva rift display a geochemical signature typical for a back-arc environment.

In contrast, regional arching of the Kola–White Sea area, accompanied by alkaline and kimberlite magmatism and weak extensional tectonics (Fig. 9; Kramm et al., 1993), probably occurred during the Late Devonian but is poorly controlled by stratigraphic data; the petrogenesis of the Kola–White Sea magmatic province is indicative of hot-spot activity (Scheglov et al., 1993; Kramm et al., 1993; Lukjanova et al., 1994).

Middle Devonian subsidence of the Vyatka rift was associated with the uplift of a broad lithospheric arch (Kuzmenko et al., 1991), whereas the Soligalich rift shows no evidence of such doming (Fig. 9).

Devonian rifts associated with the eastern margin of Baltica can be interpreted as forming part of a major back-arc extensional system which is related to a west-dipping subduction zone associated with the postulated intra-oceanic Magnitogorsk arc–trench system (Ziegler, 1989; Khain and Seslavinsky, 1991; Zonenshain et al., 1993; Nikishin et al., 1993; Zonenshain and Matveenkov, 1994; Sengör et al., 1993). Although Devonian palaeotectonic reconstructions of the Ural system are still very hy-

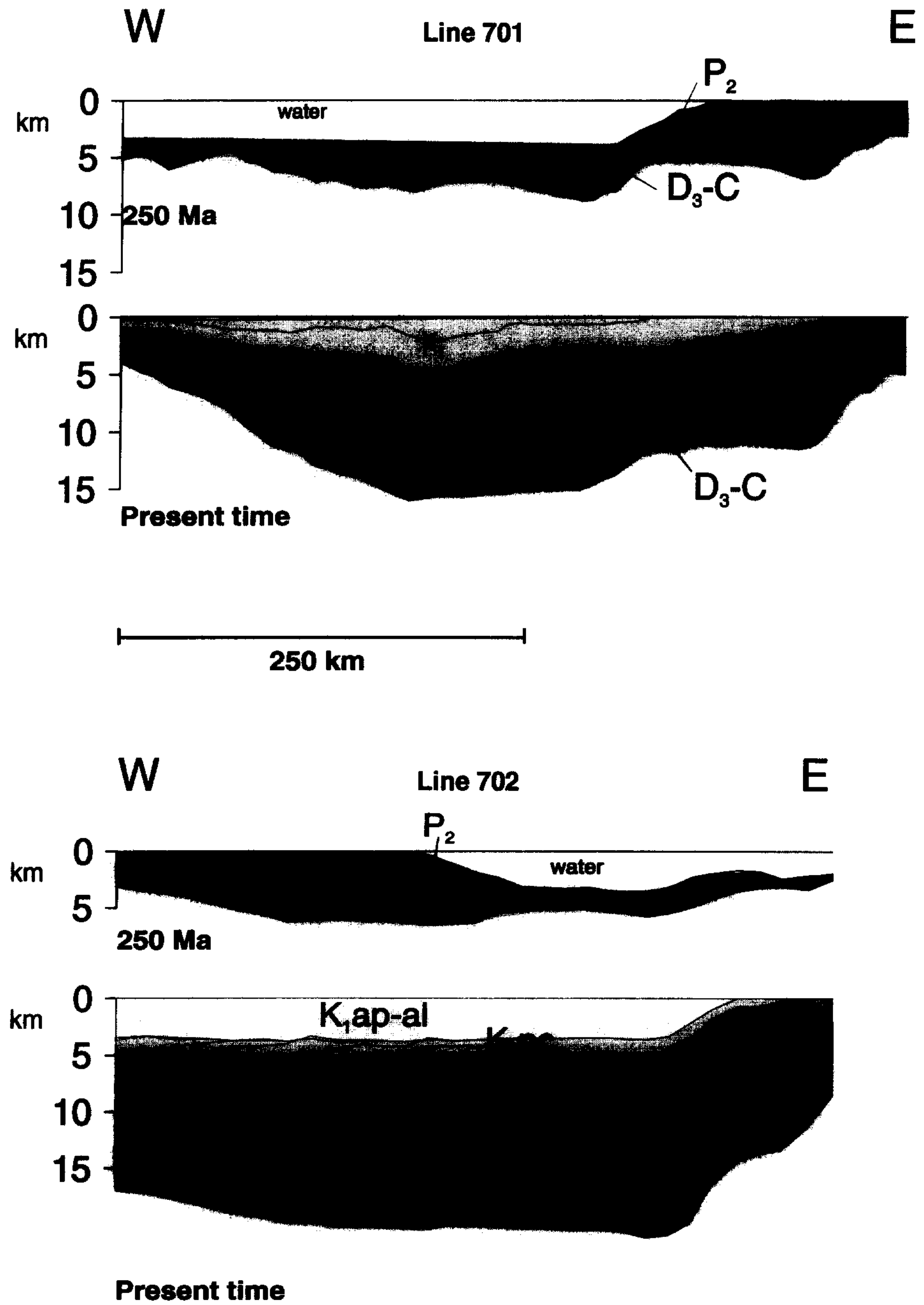

Fig. 11. Evolution of the East-Barents Sea basin, based on seismic lines 701 and 702 (without decompaction of sediments). For location of profiles see Fig. 2.

pothetical, all data available support the model of a Middle–Late Devonian subduction system within the Ural Ocean, the polarity of which, however, is a matter of dispute. Ziegler (1989), Zonenshain et al. (1993), and Sengör et al. (1993) postulate a west-dipping subduction zone. In contrast, Iudin (1990) and Puchkov (1993) favour an east-dipping subduction system, in which case a back-arc rift model would not apply to the rift system of the eastern margin of Baltica.

A model of westward subduction beneath the Peri-Ural rift belt is compatible with the following observations. Firstly, whereas an early Palaeozoic ophiolite complex can be followed throughout the Uralian orogen, the southern Urals also contain Middle Devonian ophiolites west of the Mugodjary zone; the latter are interpreted as representing remnants of an oceanic back-arc basin (Korinevsky, 1989; Puchkov, 1993; Zonenshain and Matveenkov, 1994). Both ophiolite zones are located to the west of a Devonian arc-related magmatic belt. Secondly, at the latitude of the Pechora Basin, magmatic rocks within the Urals indicate, on the basis of geochemical criteria, a change to subduction-related magmatism at the Early to Middle Devonian boundary (Bochkarev, 1990) and, as such, suggest a change in subduction polarity at that time. Thirdly, conodont stratigraphy (Puchkov, 1993; Savelieva and Nesbitt, 1996; A.V. Maslov, pers. commun., 1995) indicates that in the central Urals, Middle Devonian arc volcanics rest on Silurian–Early Devonian sediments; this suggests that arc volcanism, related to the development of a west-dipping subduction zone, began no earlier than during the Middle Devonian.

Observations from the southern Urals indicate that back-arc spreading in the West-Mugojari region occurred during the late Eifelian and that middle Eifelian–early Frasnian, subduction-related volcanism is related to a west-dipping subduction system. Moreover, at the transition from the Frasnian to the Famennian, accelerated orogenic activity is indicated in the Zilair–Magnitogorsk Basin by flysch deposits, derived from an eastern source, overlying a deep-water pelagic series (Veimarn et al., in press); this was accompanied by a phase of compressional deformation of the Sakmarian back-arc basin (Korinevsky, 1989; Milanovsky, 1989; Zonenshain and Matveenkov, 1994). Thereafter, back-arc extension may have resumed. However, during the late Tournaisian–early Visean, west-dipping subduction terminated at the onset of the Uralian orogeny during which the Sakmarian back-arc basin was closed and subducted to the east beneath the Kazakhstan terrane (Fig. 7o).

Therefore, Devonian–Early Carboniferous rifting on the eastern margin of the EEC may be related to intermittent back-arc extension. Obviously, a more precise correlation of tectonic and magmatic events within the craton and inside the Uralian belt is required to test such a hypothesis.

4.7.3. Dynamics of Middle Devonian–earliest Carboniferous rifting

The above suggests that processes related both to back-arc extension and hot-spot activity may have played a role in Middle Devonian to earliest Carboniferous rifting along the southern and eastern margin of Baltica. A possible genetic relationship among these is suggested by their contemporaneity. Although it appears that hot-spot activity is not the basic driving mechanism of lithospheric extension (e.g., Hill et al., 1992), it may contribute towards the localisation of rifts by weakening the lithosphere (Wilson, 1993). It is unknown to what extent such thermal mantle processes contributed towards the development of the pre-Middle Devonian regional unconformity on the EEP (Fig. 10).

Following the end-Silurian Caledonian suturing of Laurentia–Greenland and Baltica, global-scale changes in plate interaction, governing the Middle Devonian accretion of Avalonia to Laurussia and increasing collisional coupling between the latter and Arctica, were apparently accompanied by changes in the Palaeo-Tethys and Uralian subduction systems. These gave rise to poly-phase back-arc extension and mantle upwelling beneath Baltica. In contrast, the contemporary evolution of Laurentia–Greenland was dominated by compressional stresses related to the development of the Antler and Inuitian orogens along its western and northern margins, respectively (Ziegler, 1989). 'Passive' and 'active' rifting processes appear to have variably contributed to the development of the late Palaeozoic rifts of Baltica. Strike-slip movements were important in the Arctic–North Atlantic domain, the Pechora–Barents Sea and along the Timan–Varanger belt.

4.8. Carboniferous–Early Permian (365–256 Ma)

The first collisional contacts between Gondwana and Laurussia were established during the Famennian in the West-Mediterranean domain. As a consequence of their collisional coupling, the clockwise rotational motion of Gondwana was imparted on

Laurussia which also rotated clockwise during its progressive suturing with Gondwana during the Carboniferous Variscan and the Permo-Carboniferous Alleghenian orogenic cycles. During latest Carboniferous and Early Permian times, dextral motions between Gondwana and Baltica were compensated by a conjugate system of wrench faults which transected the Variscan fold belt and its northern foreland, resulting in the collapse of the Variscan orogen (Ziegler, 1989, 1990).

Suturing of Arctica with the northern margin of Laurussia along the Inuitian–Lomonosov orogen was completed during the earliest Carboniferous (Fig. 7o). During the Early Carboniferous, sinistral motions along the Arctic–North Atlantic mega-shear gradually ceased and gave way at the transition to the Late Carboniferous crustal extension (Fig. 7p). The Carboniferous rotational movement of Laurussia was coupled with collision of the Kazakhstan and Siberian cratons with each other and the Magnitogorsk arc–trench system, the development of an east-dipping subduction zone and progressive closure of the oceanic Sakmarian back-arc basin. Collision of the Uralian orogen with the eastern, passive margin of Baltica commenced in the south during the Bashkirian and progressed in time northwards (Ziegler, 1989, 1990; Zonenshain et al., 1993; Puchkov, 1995).

These fundamental changes in plate interaction and motions had major repercussions for the evolution of the Baltica sub-plate. Moreover, strong glacio-eustatic sea-level fluctuations, in combination with tectonic processes, provided for major changes in basin geometries (see Fig. 7o–q; Vail et al., 1977; Ross and Ross, 1987). Although tectonic environments on Baltica were, at the beginning of the Carboniferous, still similar to those of the Late Devonian, the evolution of the Variscan and Uralian orogens accounted for major palaeogeographic changes during Late Carboniferous and Early Permian times. This was paralleled by changes in stress systems controlling the subsidence of intracratonic basins on the EEC. Analyses of these basins show that their evolution was characterised by several phases of accelerated and decelerated subsidence and uplifting phases (Fig. 6). During the Carboniferous–Early Permian, the eastern parts of Baltica were dominated by carbonate and evaporitic shelves in which the sediment-starved, deep-water Peri-Caspian and East-Barents Sea basins stand out. During Late Carboniferous and Permian times, the Barents Sea area formed part of the extensive Arctic shelf that included the Canadian Arctic Archipelago and the New Siberian Islands. During these times, the Fennoscandian Shield formed a relatively low-relief highland (Ziegler, 1989).

In the domain of the Variscan 'geosynclinal system', intermittent back-arc extension ceased by mid-Visean time and gave way to regional back-arc compression, governing the development of the collisional Variscan orogen that is characterised by major nappes, extensive high pressure–low temperature metamorphism, and plutonic activity. The Variscan foreland basin developed during the Namurian and Westphalian and was partly destroyed by thin-skinned thrusting during the Westphalian terminal phases of the Variscan orogeny (Ziegler, 1990; cf. Fig. 7p).

The Variscan orogen links to the east with the Scythian orogen which fringes the southern margin of the EEC; in the Scythian orogen crustal shortening persisted into the Permian. In the Pripyat–Dniepr–Donets–Donbas–Karpinsky rift system, the main phase of crustal extension ceased at the end of Devonian times. Its late Visean–Serpukovian early post-rift stage was characterised by very rapid subsidence (Fig. 6g; Stovba et al., 1996; van Wees et al., 1996). Collision-related compressional stresses, related to the development of the Scythian orogen, may have contributed to this phenomenon. Similar syn-compressional subsidence accelerations are observed during the Palaeocene and Neogene evolution of the North Sea Basin (Cloetingh and Kooi, 1992; Ziegler et al., 1995) and the Late Cretaceous pre-inversion stage of the Polish Trough (Dadlez et al., 1995).

In the southern parts of the Urals, an east-dipping subduction zone developed in conjunction with the probably intra-Visean collision of the Kazakhstan block with the Uralian arc–trench system. Activity along this new subduction system controlled the gradual closure of the Sakmarian back-arc basin (Bocharova and Scotese, 1993; Zonenshain and Matveenkov, 1994; Koroteev et al., 1995). Collision of the evolving Uralian orogen with the southeastern parts of the EEC passive margin commenced at the transition from the Serpukhovian to the Bashkirian, as indicated by the development of a classical foreland basin (Puchkov, 1995). During the Late Carboniferous and Early Permian, this collision front propagated northward and reached the area of the Northern Urals (Fig. 7q). Middle and Late Carboniferous rapid subsidence events on the EEC (Fig. 6) mould be related to stresses exerted on the EEC from the Uralian collision front. During the Asselian–Artinskian, the Ural foredeep subsided very rapidly under the load of the advancing nappe systems. The Uralian orogeny climaxed during the Kungurian. By this time the eastern margin of Baltica was fringed by the Himalaya-type Ural mountain belt (Zonenshain et al., 1993).

During the Early Permian phases of the Uralian orogeny, the Devonian–Early Carboniferous Timan and Pechora–Kolva grabens were partly inverted (Fig. 12). Inversion of the Donbas–Karpinsky segment of the Pripyat–Dniepr–Donets rift system is thought to be related to the evolution of the Scythian orogen (Milanovsky, 1987; 1989). Two-sided thrust-loading of the Peri-Caspian Basin by the southern Urals and the Karpinsky swell caused its rapid subsidence during the Asselian–Artinskian and the progradation of major deltaic complexes into deeper waters (Zamarenov, 1970; Kiryukhin et al., 1993). During the Kungurian, the remnant basin was filled in with thick salt deposits (Milanovsky, 1987).

Whereas the Urals were characterised during the latest Carboniferous and Early Permian by intense crustal shortening, the by now inactive Variscan fold belt and its northern foreland record transtensional and transpressional tectonics, accompanied by an extensive basaltic and rhyolitic volcanism. This deformation phase can be related to a dextral translation of Gondwana relative to Baltica during the Alleghenian orogeny of the Appalachians. At the same time, rifting activity persisted in the Arctic–North Atlantic domain (Ziegler, 1989, 1990).

4.9. *Late Permian–Triassic (256–208 Ma)*

During the Late Permian, the Variscan–Appalachian suture between Gondwana and Laurussia, which together formed the core of the Permo-Triassic Pangea, had become inactive. Crustal shortening continued, however, in the Pay-Khoy, in Novaya Zemlya and Taymyr into Triassic–Early Jurassic times, whereas the central and southern parts of the Urals had become tectonically inactive (Milanovsky, 1989). On the other hand, crustal extension persisted in the Arctic–North Atlantic rift system and commenced in the Tethys domain (Ziegler, 1989, 1990).

During the Late Permian, the Barents Sea was occupied by a large shale-dominated basin in which the eastern Barents Sea deep-water basin, located in the foreland of Novaya Zemlya, stands out (Johansen et al., 1993). From the Arctic shelves a sea arm extended through the Arctic–North Atlantic rift and terminated in the evaporitic Zechstein basins of Western and Central Europe (Milanovsky, 1987; Ziegler, 1989). The southeastern parts of Baltica were covered by an extensive carbonate platform that became covered by clastics during the Late Permian (Fig. 7r). Post-rift subsidence continued in the PDD basin. The Late Permian was characterised by a cyclical lowering in global sea levels, reaching a minimum at the Permian–Triassic boundary (Ross and Ross, 1987).

Due to diachronous termination of crustal shortening in the Ural–Pay-Khoy–Novaya Zemlya orogenic belt, the Late Permian history of the corresponding sectors of the associated foreland basin differs considerably. Whereas thrusting terminated in the Urals during the Permian, it persisted in the Pay-Khoy–Novaya Zemlya–Taymyr belt during the Late Permian and Triassic and terminated only at the transition to the Jurassic (Milanovsky, 1989; Kogaro et al., 1992).

During the Late Permian, the Ural foredeep, the Peri-Caspian and the Moscow–Mezen–Volga–Ural basins were filled with clastics, mainly derived from the Ural mountains. The Peri-Caspian Basin continues to subside rapidly with sedimentation rates of clastics and carbonates matching subsidence rates (Milanovsky, 1987).

At the beginning of the Late Permian, the eastern Barents Sea–Pay-Khoy–Novaya Zemlya Basin corresponded to a deep-water trough. However, in the course of the Late Permian, major deltaic complexes prograded into this basin from the east, southeast and southwest (Junov, 1993a,b; Popova and Krylov, 1993; Ignatenko and Cheredeev, 1993; cf. Figs. 7r and 11). In the Pay-Khoy–Novaya Zemlya region, these clastics consist of material derived from an

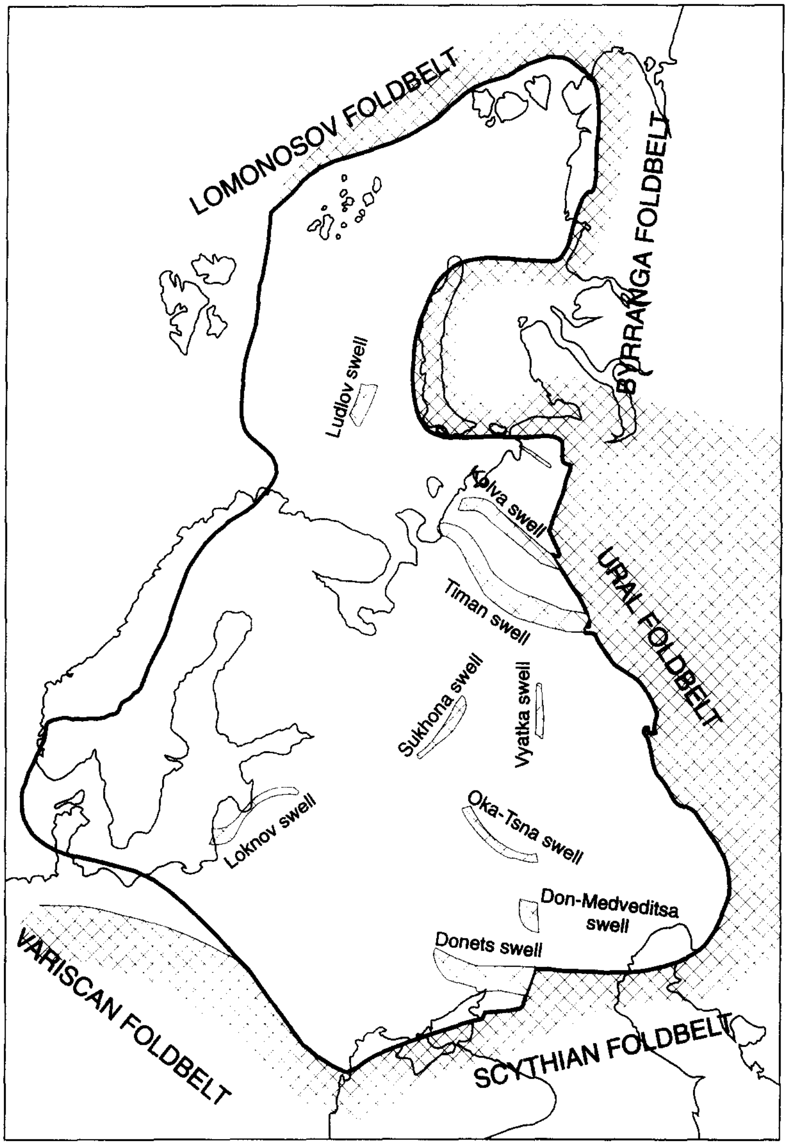

Fig. 12. Late Palaeozoic–early Mesozoic inversion structures of the East European Craton.

orogenic belt (Kogaro et al., 1992). In the Pechora Basin, continental to shallow-marine molasse-type sediments accumulated and apparently prograded along the axis of this foreland basin into the Barents Sea (Popova and Krylov, 1993). Deltas prograding from the Kola Peninsula into the Barents Sea testify to contemporaneous uplift of the Fennoscandian Shield; this is compatible with fission-track data from southern and central Sweden (Zeck et al., 1988; D. Gee, pers. commun., 1996) that indicate that uplift of the western parts of the Fennoscandian Shield and the Scandinavian Caledonides commenced during the Late Permian and persisted into Early Mesozoic times, possibly in response to a large radius doming of the Arctic–North Atlantic rift zone (Ziegler, 1988, 1990).

Although the timing of collision of the Pay-Khoy–Novaya Zemlya–Taymyr orogenic system with the Barents Sea passive margin is unknown, the oceanic basin separating them was closed no later than at the beginning of the Late Permian. Thereafter, the eastern Barents Sea was incorporated into the Ural–Novaya Zemlya–Taymyr foreland basin. Superposition of major Late Permian and Late Triassic deltaic clinoform complexes in the eastern Barents Sea (Kerusov, 1993), separated by an Early Triassic flooding event, testifies to a two-phase Novaya Zemlya orogeny in which Early Triassic rapid deepening of the foreland basin presumably corresponds to a period of rapid thrust-loaded subsidence (Fig. 11).

Increasing instability of the core of Pangea during the latest Permian and Triassic, was manifested by an acceleration of rifting activity. During the Early Triassic, the Arctic–North Atlantic rift system propagated southwards into the area of the North Sea and to the west of the British Isles. Rifting intensified also in the Tethys domain and began to propagate westwards during the Triassic. A branch of the evolving Tethys rift system can be seen in the Polish Trough (Dadlez et al., 1995), which is superimposed on the Tornquist line, marking the southwestern margin of the Precambrian EEC. By Late Triassic times, the Tethys and Arctic–North Atlantic rift systems interfered with each other in the North Atlantic domain (Ziegler, 1988, 1990).

On the EEC, the distribution of Triassic basins is similar to that of the Late Permian; however, a gradual reduction of the area covered by sedimentary basins is evident (Fig. 7r, s). Despite rising sea levels, sedimentation ceased towards the end of the Triassic on much of the East European Platform, resulting in the development of a regional hiatus (Milanovsky, 1987). The dynamics controlling this broad uplift are unknown. On the other hand, sedimentation continued on the Barents shelf and in the rift-related basins of Western and Central Europe, the Arctic–North Atlantic and Tethys domains (Ziegler, 1988, 1990).

Subsidence curves for the Peri-Caspian Basin, the Russian Platform, the West-Siberian Basin and the Scythian Platform reflect a similar rapid subsidence event and accelerated sedimentation rates straddling the Permian–Triassic boundary, as seen also in the Eastern Barents Sea and the Mid-Polish Trough (Dadlez et al., 1995). Contemporaneous rifting events are evident within and along the Ural belt, controlling subsidence of grabens and the extrusion of basalts (e.g., Chelyabinsk graben; Milanovsky, 1989), as well as along the Scythian belt (Manych rift system; Nikishin et al., 1994). Basalts occurring in the Polar Ural–Pay Koy foreland basin (Andreichev, 1992) may form part of the rift-related volcanic province of the West-Siberian Basin (Surkov and Zhero, 1981; Peterson and Clarke, 1991).

At the transition to the Jurassic, the Pay-Khoy–Novaya Zemlya–South Taymyr orogenic system was affected by a last, albeit major phase of crustal shortening and granitic plutonism (Milanovsky, 1987; Uflyand et al., 1991; Kogaro et al., 1992; Zonenshain et al., 1993). Foreland thrusting in the eastern Barents Sea (Komaritsky, 1993) was paralleled by further inversion of Devonian rifts in the Timan–Pechora area (Figs. 6 and 12). There is also evidence for end-Triassic compressional deformation within the Urals (Milanovsky, 1989; Sokolov, 1992).

Along the southern margin of the EEC, a Late Triassic collisional event affected the Caucasus–Scythian area; it was accompanied by further inversion of the Karpinsky Kryazh belt (Fig. 12; Nazarevich et al., 1986; Kazmin and Sborschikov, 1989; Nikishin et al., 1994).

5. Craton boundary evolution and intraplate stresses

Repeated changes in the configuration of sedimentary basins on the Precambrian EEC and in the mechanisms governing their subsidence (rifting, post-rift sag basins, foreland basins, etc.) reflect changes in the stress systems affecting the craton and its thermo-mechanical properties. Changes in the intraplate stress regime are mainly controlled by processes affecting the plate boundaries (Zoback, 1992; Zoback et al., 1993), which through time, repeatedly coincided with the margins of the EEC. However, different plate boundary processes affected opposite margins of the craton synchronously, as seen, for instance, during the late Vendian–Early Cambrian and the Late Carboniferous–Early Permian. Therefore, intraplate palaeostress regimes were often very complex and are difficult to reconstruct.

In this respect, a comparison of the drift pattern of the EEC, its interaction with neighbouring plates, and its intraplate stress history is of special interest. Figs. 8 and 13 summarise the complex Riphean and Palaeozoic drift pattern of Baltica, which was char-

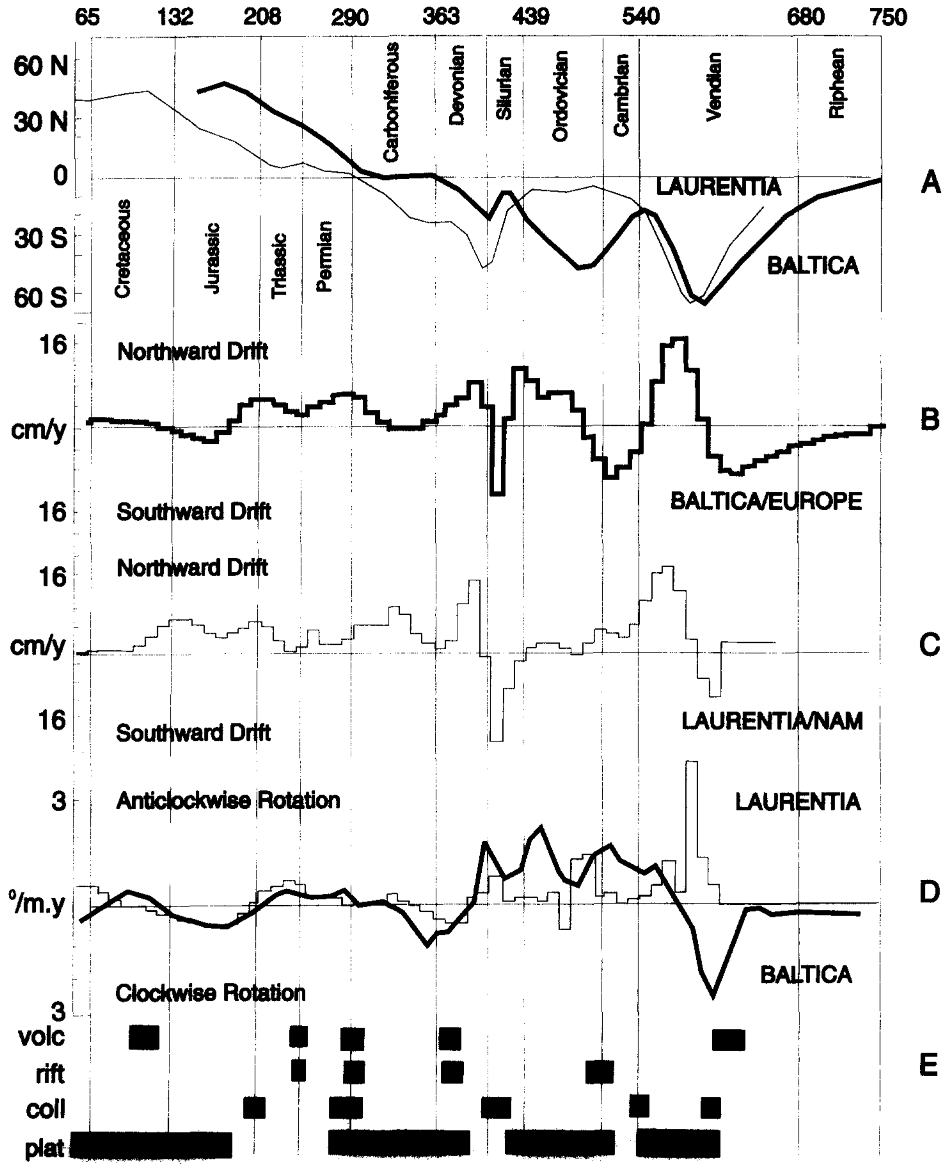

Fig. 13. (A) Latitudinal drift velocity of Baltica (reference location: 60°N, 10°E) and Laurentia (reference location 40°N, 270°E) since Riphean time. Time scale after Harland et al. (1990), except for Vendian–Cambrian boundary. (B) and (C) Latitudinal drift rates for Baltica and Laurentia, separated into southward and northward movements. (D) Angular clockwise and anti-clockwise rotation rates for Baltica and Laurentia (after Gurnis and Torsvik, 1994). (E) Main tectonic and magmatic events in Baltica; volc = flood basalt volcanism, rift = main rifting phases, coll = main collisional events, plat = periods of regional intracratonic platform subsidence.

acterised by repeated changes in direction, velocity and rotational motions. The timing of plate motion changes and the evolution of sedimentary basins on the craton indicates that the stress history of the EEC was controlled by global plate kinematics, plate interactions and their changes through time. Based on palaeomagnetic data (Posenen et al., 1989; Gurnis and Torsvik, 1994), periods of major plate reorganisations can be compared with the following significant events which are recorded in the evolution of sedimentary basins on the EEC (cf. Figs. 8 and 13):

(1) The end Early (±1350 Ma) and end Middle

(±1050 Ma) Riphean phases of basin inversion, although chronostratigraphically poorly constrained, appear to coincide with significant changes in plate motions (Milanovsky et al., 1994).

(2) Late Riphean rifting along the eastern margin of the EEC (±1050–650 Ma) is contemporaneous with an acceleration of the southward movement of Baltica.

(3) Dalslandian incorporation of Baltica into the Riphean Pangea, ending at 0.9 Ga, was followed by an abrupt change in drift pattern.

(4) Late Vendian–earliest Cambrian (±620–530 Ma) rifting along the western margin of the EEC, culminating in opening of the Iapetus and Tornquist oceans and thus its separation from the Riphean Pangea, is marked by a sharp change in the drift pattern of Baltica and a cycle of regional subsidence.

(5) Middle Early Cambrian (±540 Ma) intracratonic inversion tectonics (Soligalich deformation phase) coincide with the suturing of the Barentsia terrane to the northeastern margin of the EEC, uplift of the latter and an abrupt change in its drift pattern.

(6) Collision of the northern Taymyr–Severnaya Zemlya terrane with the EEC at the transition from the Cambrian to the Ordovician (±510 Ma) coincides with the Finnmarkian orogeny of the North Scandinavian Caledonides and a change in the drift velocity of Baltica.

(7) During the Early Ordovician (±510–475 Ma) minimum in latitudinal plate velocity, rifting affected the eastern margin of the EEC and culminated in the opening of the Palaeo-Ural ocean.

(8) Late Silurian (±425–410 Ma) collision of Baltica and Laurentia–Greenland and their suturing along the Arctic–North Atlantic Caledonides entailed an abrupt change in plate motion.

(9) The Middle Devonian to earliest Carboniferous (±385–362 Ma) cycle of rifting and possible plume activity was accompanied by a deceleration in the motion of Baltica.

(10) The intra-Visean (±345 Ma) onset of the main-phase of Variscan–Sakmarian orogeny coincides with the development of an east-dipping subduction zone in the Ural Ocean.

(11) The climax of Uralian orogeny at the Carboniferous–Permian boundary (±290 Ma) coincides with the onset of anti-clockwise rotation of Baltica which, by this time, formed part of Pangea.

(12) Slowing of plate motions at the Permo-Triassic boundary (±245 Ma) coincides with extensional tectonics and intraplate magmatism along the western, southern, and eastern margins of the EEC and orogenic activity in the Pay-Khoy–Novaya Zemlya system.

The global stress map shows that present-day intraplate stresses are related to the direction of plate motion (Zoback, 1992; Zoback et al., 1993; Cloetingh et al., 1994). If this applies also to the past, then changes in plate motion ought to correlate with changes in intraplate stress regime. Data available for the EEC provide an opportunity to further evaluate this concept.

6. Conclusions

The Precambrian EEC consists of a collage of continental and arc-related terranes which were assembled and welded together during a sequence of orogenic cycles spanning Archaean, Early Proterozoic, and Riphean times. During the Late Riphean and Vendian, Baltica formed part of a Pangea-type supercontinent from which is was separated again at the end of the Vendian. During Cambrian to Late Silurian times, Baltica played the role of an independent plate. Upon its Caledonian suturing to Laurentia–Greenland, it formed part of the Laurussian plate which was integrated into Permo-Triassic Pangea during the Variscan–Appalachian and Uralian orogenic cycles.

The sedimentary cover of the EEC consists of a mosaic of superimposed basins of different age and origin. Basins developed in conjunction with repeated rifting cycles, separated by periods of thermal subsidence and collisional phases during which flexural foreland basins developed and intraplate compressional stresses controlled the inversion of pre-existing tensional basins.

The contribution of eustatic and tectonically induced sea-level fluctuations, controlling sedimentation and erosion on the EEC, warrants further analysis and comparison with other cratons.

The EEC records several phases of intraplate magmatism of different origin, including, plume, intracratonic and back-arc 'passive' rifting and post-orogenic magmatism related to slab detachment and the collapse of orogenic belts. Due to insufficient geochemical, isotope and trace element data, it is not possible at present to separate the different types of magma generation and to propose dynamic processes responsible for the development of the observed cycles of intracratonic magmatism.

The Middle and Late Devonian rifting cycle of the EEC may be related to changes in the subduction geometry of the Uralian and Palaeo-Tethys arc–trench systems, possibly controlling back-arc extension. There are indications for contemporaneous mantle plume activity.

Preliminary correlations between the stress history of the craton, the evolution of its sedimentary basins, and lithospheric plate motions through time, indicate that the kinematics of plate interaction controlled the origin and development of intracratonic sedimentary basins.

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