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Sedimentary Geology 156 (2003) 35–58

Sedimentary Geology

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Pre-Mesozoic geodynamics of the Precaspian Basin (Kazakhstan)

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Received 1 March 2000; received in revised form 1 December 2001; accepted 19 July 2002

Abstract

This paper concentrates on the Pre-Mesozoic history of the Precaspian Basin representing more than half of the 20 km of its sedimentary thickness. We present the evolution of sediments deposited in the basin through time by several regional sections of the basin. These image, from the Devonian, the creation and evolution of a deep intercontinental basin which existed from Carboniferous to Permian. Palinspastic reconstructions show also the evolution of the basin's margins, their relations with the southern pre-Uralian foredeep in the east and carbonate build-ups in the southeast.

We present a reconstruction for the early history of the Precaspian Basin, which became an individual basin only at the end of the Permian as a result of the Uralian orogeny.

The basement of the Central Precaspian depression is characterised by a thin crust where a low-velocity layer is absent. A layer of abnormally high velocities exists below the crust, which on the basis of structural and gravitational field data is interpreted as eclogite.

We put forward a scenario for the age and emplacement of eclogite lenses at the base of the crust. They are possibly implemented in the hangingwall, from a westward subduction zone existing during the Vendian, before the collision giving birth to the Proto-Urals.

Most of the basin subsidence is due to the thinning (mainly during Devonian) and increase of the average density of the crust. Present crustal density has characteristics of intermediate to lower crust interpreted by some authors as oceanic. We propose that the loading effect of the eclogites emplacement may explain also a part of the Precaspian Basin subsidence.

Keywords: Precaspian Basin; Subsidence; Crustal thinning; Eclogites; Isostasy

1. Introduction

The Precaspian Basin is located at the southeastern corner of the East European platform (Fig. 1). The thick Lower Permian salt is one of the main features of the basin, and traditionally, the location of the basin is defined within the area of the salt plugs. A Pre-Kungurian tectonic–sedimentary structure, which is up to 1500 m in height, borders this area to the northwest. The structure stretches continuously in a sub-meridian direction from Kotel'nikovo in the south via Volgograd through Saratov in the north where it sharply turns to the east and stretches at the latitude of Uralsk through Orenburg (Fig. 1). Folded structures of the Urals border the depression to the east. The

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PII: S0037-0738(02)00281-6

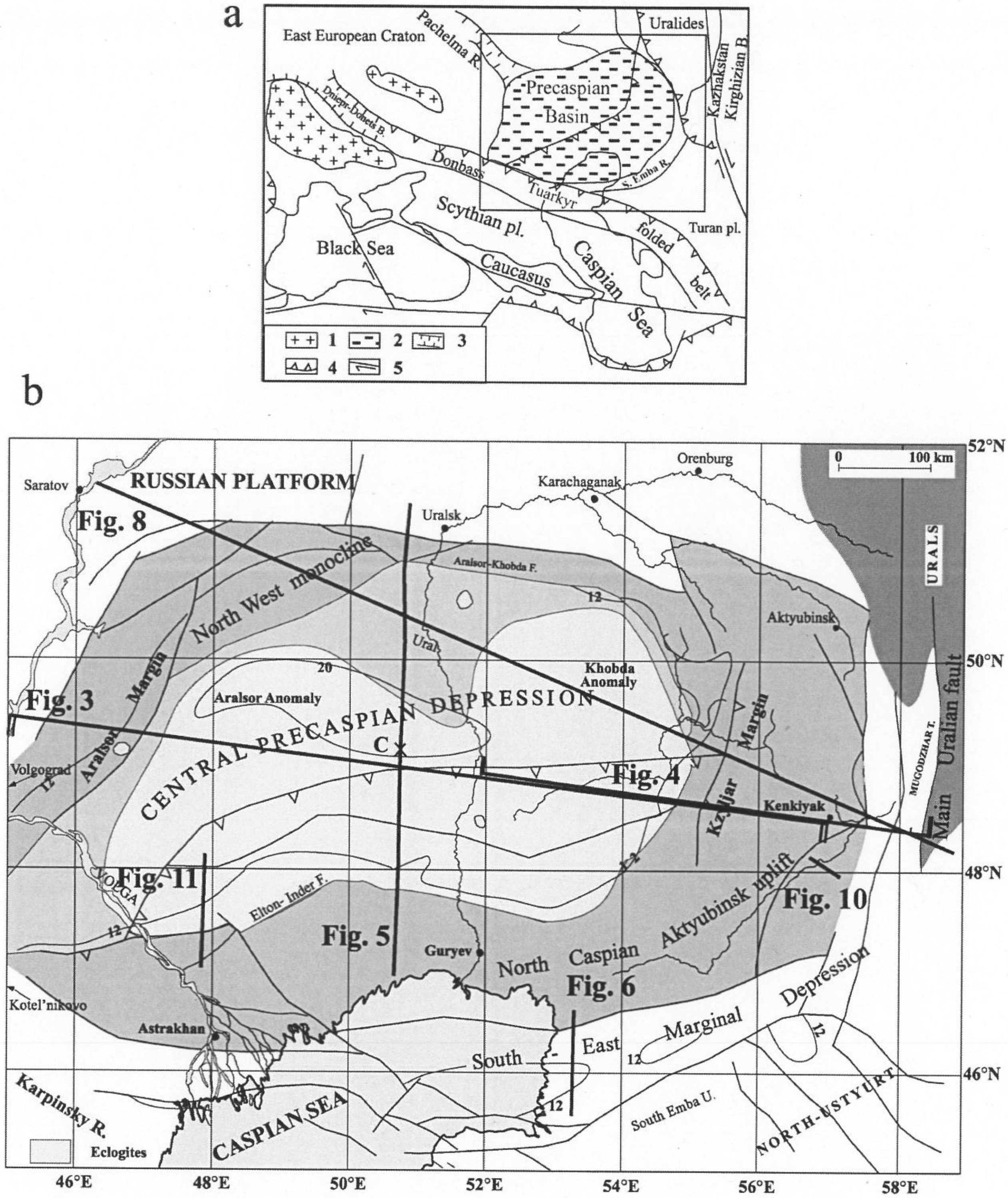


Fig. 1. (a) Tectonic location of Precaspian Basin scheme. 1—Basement uplifts, 2—Precaspian Basin, 3—border faults of rift systems, 4—thrust, 5—transcontinental shear zone. (b) Main units of the Precaspian Basin, position of sections shown in Figs. 3–6, 8, 11.

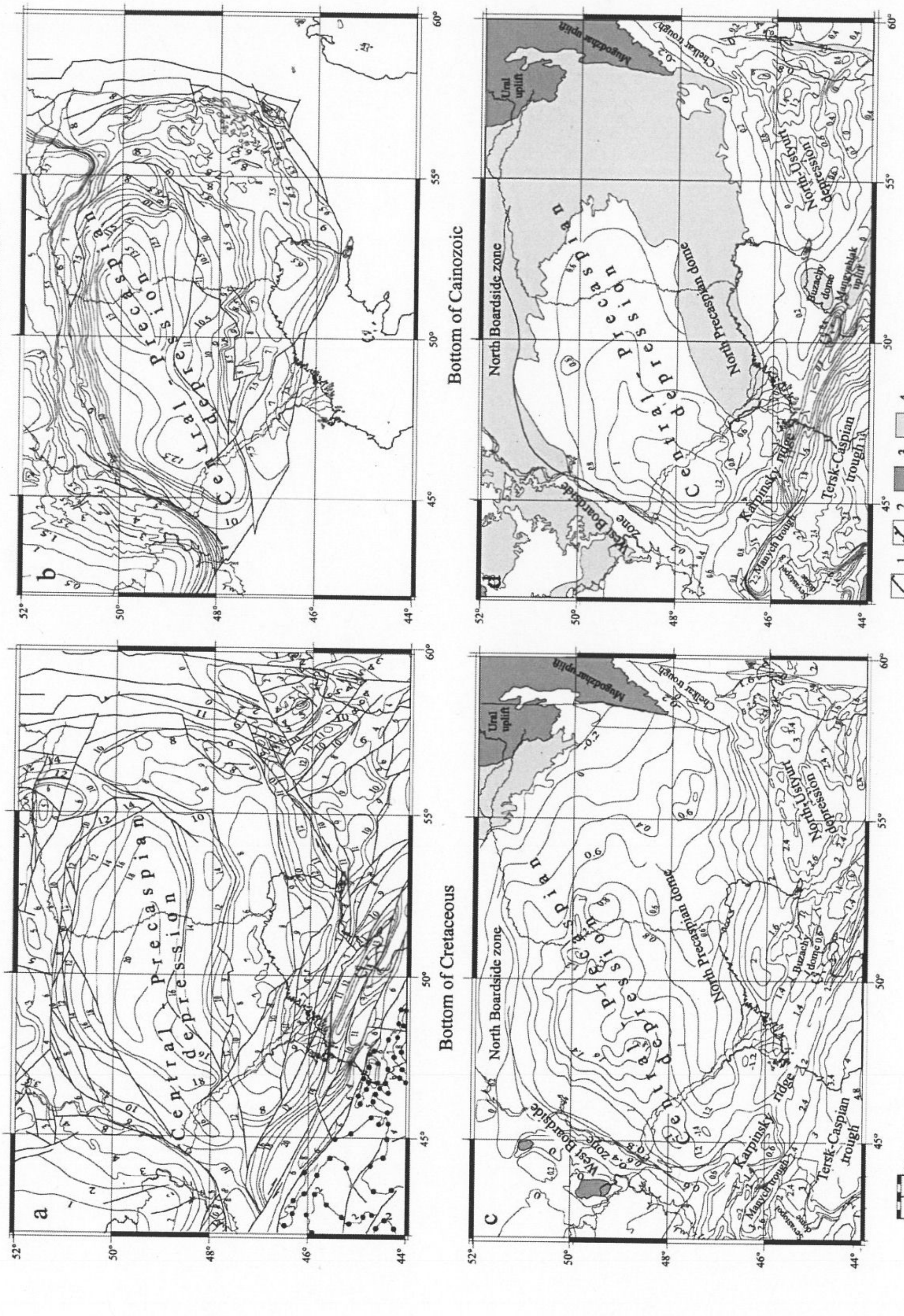


Fig. 2. Isobath maps of Precaspian Basin at the level of main reflectors, isolines in km: (a) basement (grey lines with points: isolines at depth base of Palaeozoic); (b) base of Devonian; (c) base of Cretaceous; (d) base of Cainozoic. 1: Faults; 2: contour of the area where Cretaceous sediments lay on pre-Jurassic rock complex; 3–4: outcrops of pre-Cretaceous (c) or pre-Palaeogene (d) rock complexes, 3: basement, 4: sedimentary cover.

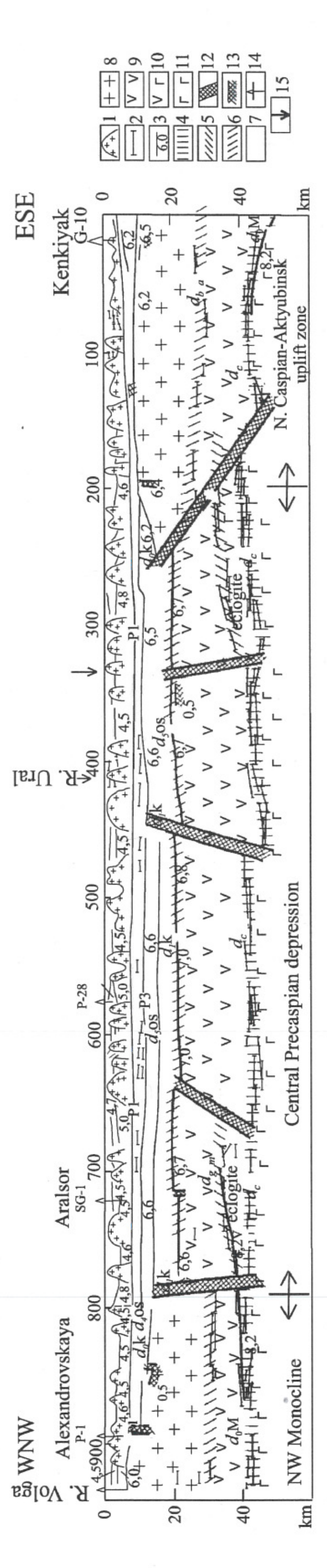


Fig. 3. Geological–geophysical crustal section Volgograd–Chelkar (from Volozh et al., 1975, modified), location in Fig. 1. (1) Salt domes; (2) reflectors: P_1—top of subsalt sediments, P_3—base of Devonian; d_{ba}—boundary upper/lower crust, dgm—surface of abnormal gravity maximum, d_c—bottom of crust; (3) refraction events with velocity km/s: d^4_{OS}—first subsalt, d^5_{OS}—second subsalt, d_0k—surface of upper crust, d_1k—top of lower crust, d_0M—Moho; (4) base of crust; (5) top of abnormal layer; (6) top of lower crust; crustal complexes: (7) sediments, (8) upper crust, (9) lower crust, (10) abnormal gravity maximum (eclogites), (11) upper mantle; (12) tectonic faults; (13) upper part of magnetic bodies; (14) deep boreholes; (15) point of sections crossing.

palaeozoic South Emba fault-related uplift borders the depression in the southeast. The Donbass–Tuarkyr system (Fig. 1a) of inversion uplifts forms the southwest boundary of the depression.

The Precaspian Basin has been a long-lived structure, contiguous to orogenic areas and oceanic sutures. This specific situation could explain the structure of the thin crust and the presence of a high-velocity layer in the mantle, at the base of the crust.

We use available data to propose a Riphean age for the oldest sediments and the presence of eclogite lenses injected at the base of the crust along a Vendian front of deformation. The loading effect of the eclogites is proposed to explain a part of the basin subsidence. The major remaining part of subsidence is explained by the isostatic response to the presence of a thin and high density crustal layer below the basin. This paper builds on the paper by Brunet et al. (1999) for the presentation of the data: i.e., isobath maps and palinspastic reconstructions through the basin and its margins, and for the discussion about the main driving mechanisms of the basin's subsidence. Obviously, the geodynamic understanding of the Precaspian Basin's evolution depends on a detailed study of the sedimentary infill and of the deep structure of the basin.

2. Main characters of the sedimentary infill

2.1. General setting

Four large structural–sedimentary elements can be recognised in the basement surface within the Precaspian salt-plug area. These are: the North–West monocline, the Central Precaspian depression, the North Caspian–Aktyubinsk uplift, and the South–East Emba system of deep basins (Fig. 1).

Comparison of isobath maps (Fig. 2) indicates that the Central Precaspian depression and its margin structures (North–West monocline and North Caspian–Aktyubinsk uplift zone) are the most permanent structures of the Precaspian Basin. Available data from thousands of kilometres of seismic lines allowed the construction of these maps (Volozh, 1991): depths of the basement surface and of the main seismic horizons corresponding to surfaces of structural unconformity. These seismic horizons are recognised by all authors and the various published depth maps are rather similar, but the attribution of an age to the deepest seismic horizons is a subject of controversy. It leads to very different propositions for the age and geodynamics at the origin of the basin. Fig. 2 presents a basement depth map (Fig. 2a), depths of horizon P_3 that we assume as the base of Devonian (see the discussion below) (Fig. 2b), horizon III—Pre-Cretaceous (Fig. 2c) and horizon I—Pre-Cainozoic (Fig. 2d). The reader is referred to Brunet et al. (1999) for complementary maps of horizons P_2 (Middle Carboniferous unconformity) and P_1 (base of the Kungurian salt).

In our interpretation, the sedimentary cover of the Central Precaspian depression exceeding 20 km (Figs. 2 and 3) contains: Riphean terrigenous–carbonate (4 km), Lower Palaeozoic (Vendian to Middle Ordovician) terrigenous (2 km), Upper Ordovician to Silurian carbonate (2 km), Devonian to Lower Permian terrigenous (4 km), Kungurian to Kazanian salt (4 km), Upper Permian to Triassic red bed (2 km), and Jurassic, Cretaceous, and Cainozoic carbonate–terrigenous deposits (2.5 km) (Volozh, 1991).

The Central Precaspian depression experienced the highest subsidence rate in the basin, as indicated by the sedimentary infill of distal and deep-marine deposits that can be observed on the following sections through the basin.

2.2. Palinspastic reconstructions of sections

A deep-water intercontinental basin existed within the Central Precaspian depression from Carboniferous to Permian. Its creation and evolution is illustrated along a selection of palinspastic reconstructions transecting the basin and showing the evolution of its margins (Figs. 4, 5 and 6).

Fig. 4 demonstrates the relations between the northeastern part of the Precaspian Basin and the southern pre-Uralian foredeep. The latter was mainly affected by an Early Permian phase of thrusting representing a stage of E–W Uralian compression (Giese et al., 1999).

Fig. 5 is a general cross-section of the basin perpendicularly to its main W–E direction, and Fig. 6 presents some carbonate build-ups on the southeastern margin.

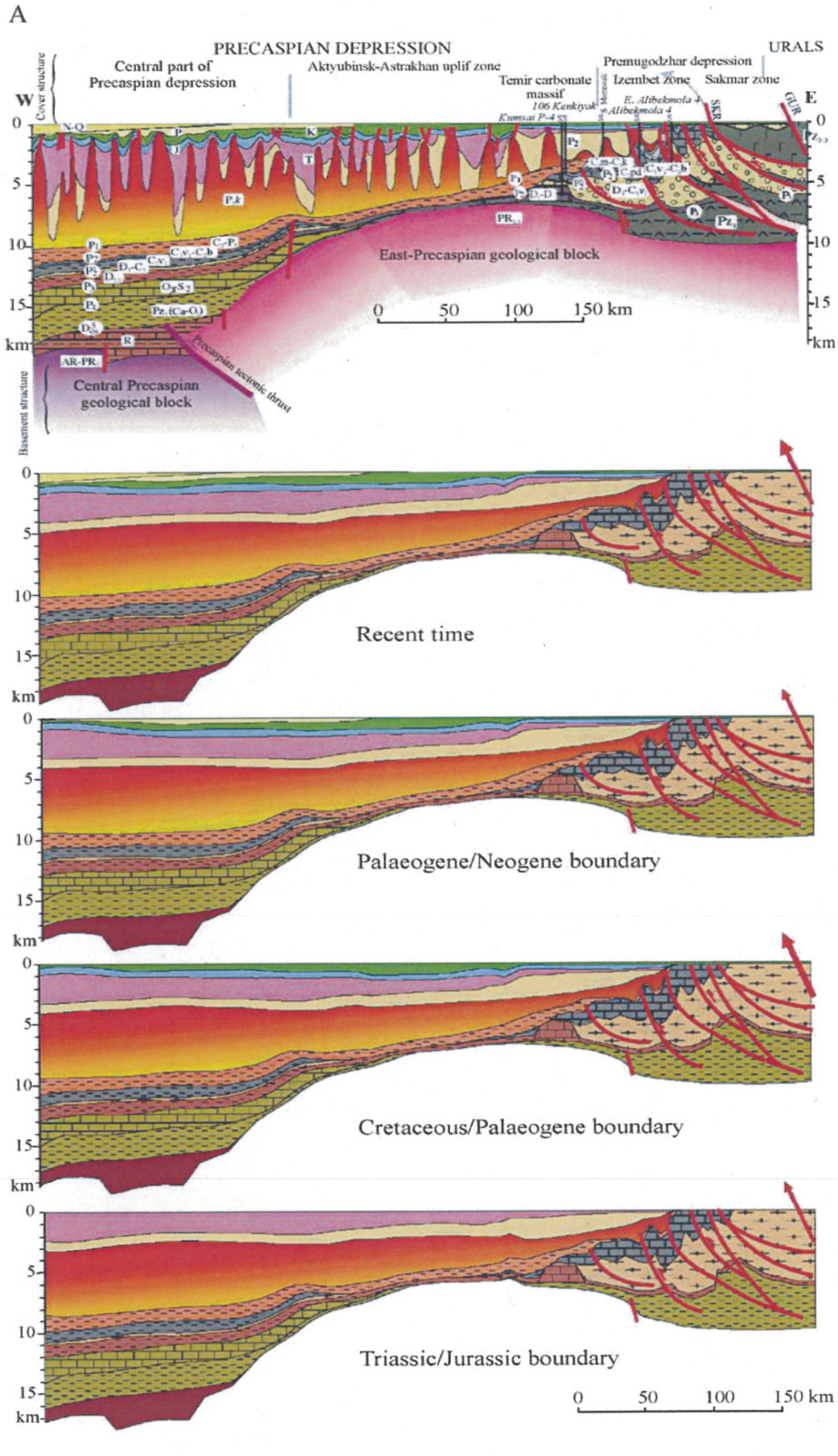
Before and during Devonian, the area of the Precaspian Basin was in a shallow water environment. Then the palaeo waterdepth considerably increased (more than 1.5 km) during the Carboniferous. Palaeodepths were determined (Volozh, 1991), mainly during Carboniferous, by the heights of the sediment scarps of the carbonate platform, located to the north of the basin (Fig. 5), and also by the heights of intercontinental carbonate build-ups (Karaton, Tengiz, Yuzhnaya) on the southern slope of the North Caspian–Aktyubinsk uplift zone (Fig. 6). The basin was filled during the Permian. From Triassic to Present, several additional kilometres of sediments were deposited in shallow marine or continental conditions. The post-salt sedimentary cover is mainly affected by salt-flow creating numerous salt diapirs. Rim-syncline infill is predominantly composed of Mesozoic sediments. The palinspastic profiles (Figs. 4, 5 and 6), have been tentatively restored to show initial thicknesses of uniform layers of salt and post-salt sediments deposited on the whole basin surface. These thicknesses were derived by integrating the salt volumes in all the salt domes and sediments volumes in the rim-synclines through all the seismic profiles (Volozh, 1991). A numerical approach has also been attempted on this type of reconstruction (Ismail-Zadeh et al., 2001).

2.3. Proposed early history of the basin

The Precaspian salt basin became an individual basin only at the end of Permian, when orogenic Urals belt arose along its eastern margins and inversion uplift occurred in the Donbass–Tuarkyr rift system (Volozh et al., 1997, 1999). Before these events, parts of the depression were incorporated in various sedimentary basins and the Precaspian Basin did not yet exist in its present configuration.

Our view of the regional geodynamical evolution before the Precaspian Basin Permian origin is exposed in the following paragraph. It should be noted that very different other reconstructions exist in the literature. They vary according to timing of terranes collision in the south and east and also on the history of oceans opening and closing between these terranes and the East European platform.

In the Middle Proterozoic, the Central Precaspian depression was included into the structure of the


A
PRECASPIAN DEPRESSION
URALS
Central part of
Precaspian depression
Aktyubinsk-Astrakhan uplif zone
Temir carbonate
massif
Premugodzhar depression
Izembet zone
Sakmar zone
W
E
Cover structure
East-Precaspian geological block
Central Precaspian
geological block
Basement structure
Precaspian tectonic thrust
0 50 100 150 km
km
Recent time
Palaeogene/Neogene boundary
Cretaceous/Palaeogene boundary
Triassic/Jurassic boundary
0 50 100 150 km

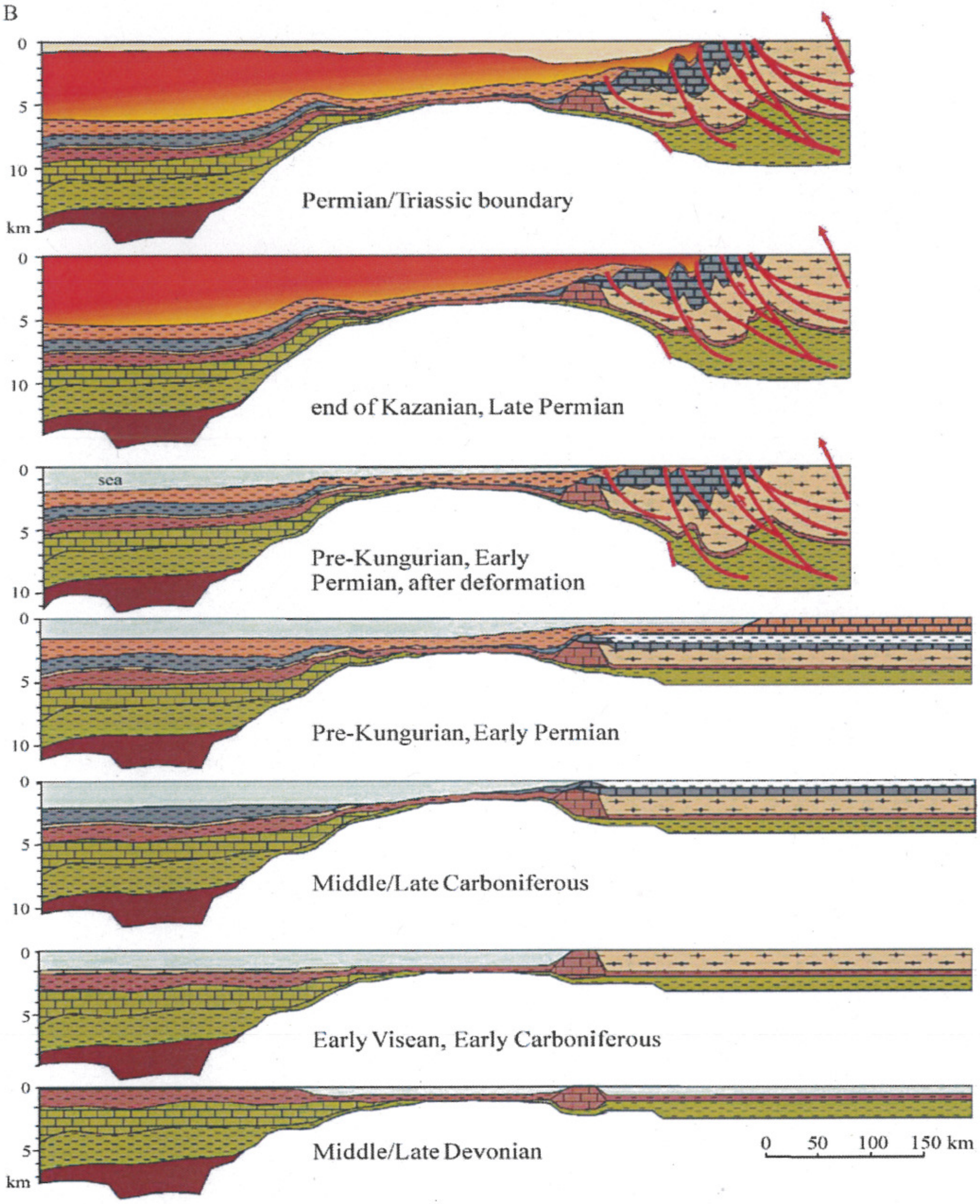


Fig. 4. Palinspastic reconstruction along a profile on the east Precaspian margin, line Ural river–Izembet (Urals) (location in Fig. 1). Some levels are grouped during the evolution but detailed with different colors on the real section. First section is the Present real section, the following are reconstituted without movements of salt. Index circled, reflectors: P_1—top of subsalt sediments, P_2, $P_{2'}$, P_3—base of Devonian, P_4, d^5_{OS}. Stratigraphic levels: N–Q, Neogene–Quaternary; P, Palaeogene; K, Cretaceous; J, Jurassic; T, Triassic; P, Permian; P_2k, Kungurian; C, Carboniferous, C_3k, Kasimovian; C_2m, Moscovian; C_2b, Bashkirian; C_1v, Visean; D, Devonian; S, Silurian; O, Ordovician; Ca, Cambrian; Pz, Palaeozoic; R, Riphean; AR–PR, Archaean–Proterozoic; SKR; Sakmaro–Kokpektyn Fault; GUR, Main Urals Fault.

Baltica continental passive margin (Fig. 7) (Mossakovsky et al., 1998). At this period, a shelf basin opened towards a deep-water oceanic basin in the east. At Late Riphean–Early Vendian, a foredeep basin was created in front of the Proto-Urals, within the domain of the Central Precaspian depression

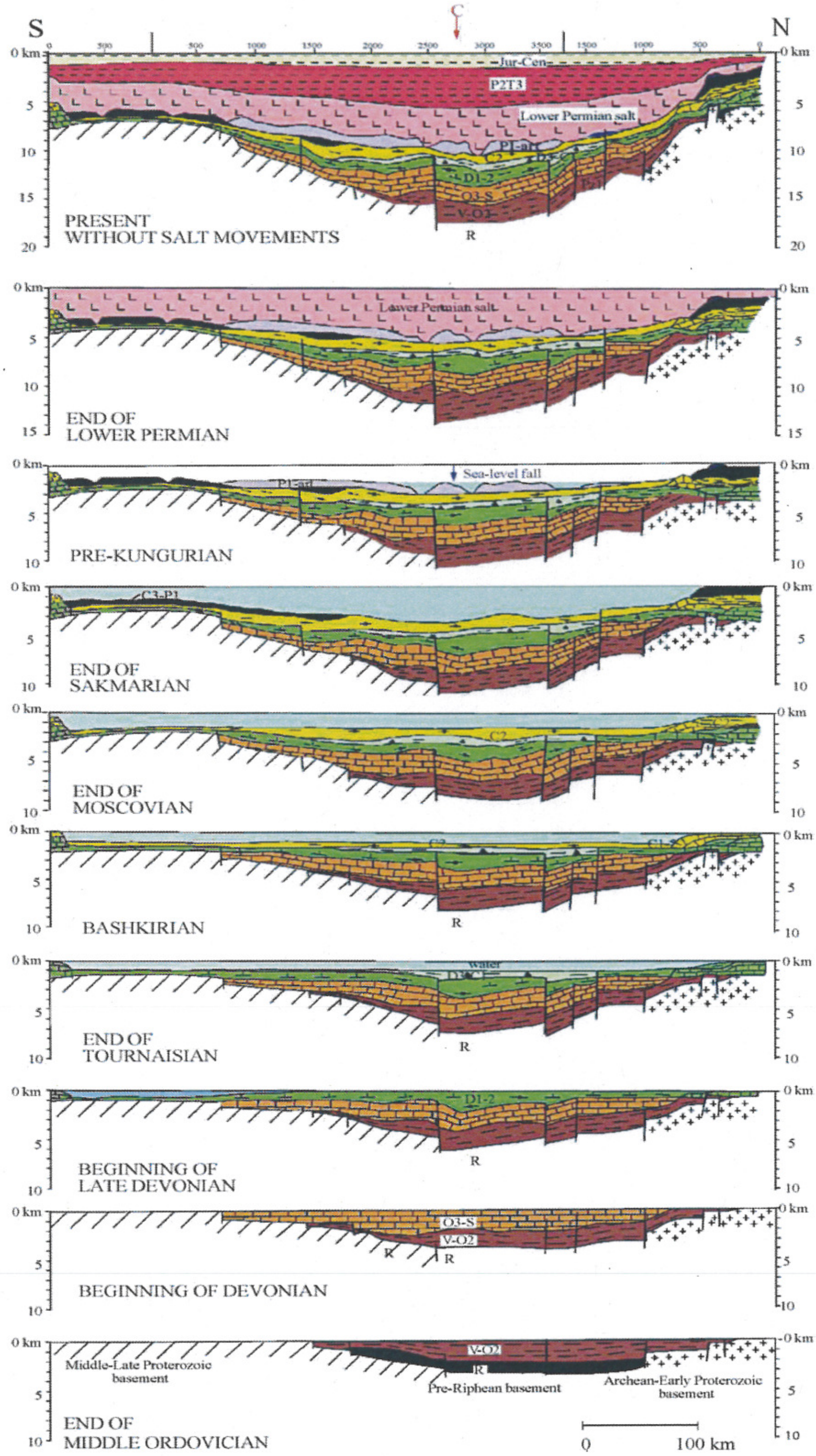


Fig. 5. Reconstruction of evolution of the Precaspian Basin along a N–S line (location in Fig. 1) (after Volozh, 1991; Brunet et al., 1999). Layers are reconstituted without the movements of salt.

limited to the south and southeast by an orogenic structure (currently corresponding to the North Caspian–Aktyubinsk block as remnant of a larger structure). At the end of Proterozoic, collision of Baltica and Gondwana continents led to the formation of the supercontinent Rodinia. The Proto-Urals Ocean closed after subduction dipping towards the west (Kheraskova et al., 2001). It formed the accretionary Proto-Urals folded belt in the Late Vendian (Puchkov, 1997; Glasmacher et al., 1999; Fig. 8). At the end of Vendian–Early Cambrian, the Proto-Urals orogen was dismantled in connection with break-up of Rodinia (Powell et al., 1993; Dalziel et al., 1994; Glasmacher et al., 1999; Table 1). Rift basins were created here by the collapse of the orogen. Later, they evolved into the Urals Ocean in the east and into the South Emba intercontinental rift in the south (Fig. 1). At the end of Ordovician, the North Caspian–Aktyubinsk uplift zone was involved in the subsiding area, as a result of the Urals Ocean opening (Burtman et al., 2000). It lasted till Middle Devonian and rifting occurred within the South Emba trough. This resulted in an expansion of the epicontinental basin and its connection with the oceans surrounding the East European platform (Urals and Palaeo-Tethys). The whole Precaspian territory was included in sea shelf basin at Late Ordovician–Silurian. In the south, it was part of the back-arc basin system of the Palaeo-Tethys Ocean. At the end of Silurian, collision processes started on the northern periphery of the Palaeo-Tethys. These processes have resulted in the formation of a collided belt in the southern framework of the Precaspian Basin. In the Early Devonian, the border between the Precaspian Basin and the Caledonian orogenic belt foredeep basin was created within the South Emba trough. This basin accumulated terrigenous material, transported from the orogenic belt. It predetermined the formation of a basin, not compensated by sedimentation, within the Central Precaspian shelf depression. In the Middle Devonian, an important stage in the evolution of the Precaspian region, major structural changes occurred as well within the basin as on its periphery. The onset of the Uralian deformation is dated from the Mid-Late Devonian. It is based on the radiometric ages of HP metamorphism and beginning of the Zilair flysch sedimentation in southwest Urals (Brown et al., 1998; Hetzel et al., 1998; Giese et al., 1999). During this period, distension was active in the Precaspian Basin. Subduction occurred on the Kazakh active margin, then collision between the East European and Kazakhstan plates, as the arrival of the Ustyurt microcontinent in the South Emba area (Zonenshain et al., 1990). This resulted in the final structural configuration of the Central Precaspian depression as an isolated deep-water intercontinental basin. Folded belts or uplifts created southern and eastern barriers around the Precaspian Basin. They periodically cut the communications with the open ocean leading to the formation of thick (several kilometres) Permian salt layers (Volozh, 1991; Snyder et al., 1994; Kuznetsov, 1998).

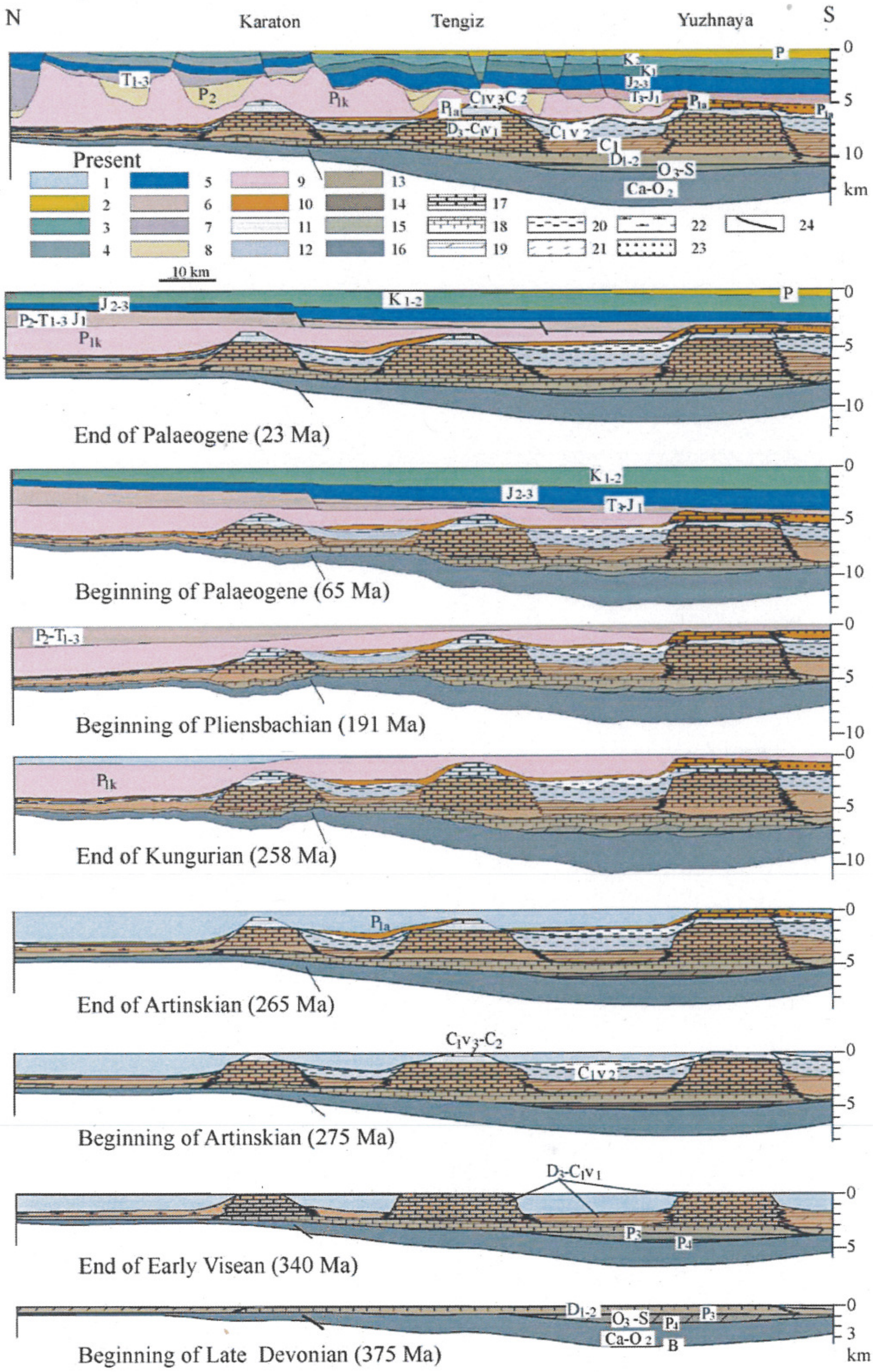
Thus, the western half of the depression was incorporated in a sedimentary basin which continuously developed since Late Riphean, while the southeastern half belonged to a large orogenic area up to the beginning of Devonian. During Devonian through Early Carboniferous, sedimentation took place all over the entire region of the depression which represented the shelf of the deep-sea marginal basin (Volozh, 1991; Zholtayev, 1989). The latter was located in front of the subduction zone, which separated the East European continent from the Urals Ocean. The Precaspian Basin, in its present shape and size, appeared during Permian when it was isolated from the southern and eastern areas.

3. Deep structure of the basin

Geophysical data (gravitational, magnetic, and seismic) can illustrate a distinct expression of the Central Precaspian depression in the structure of the mantle, basement, and sedimentary cover (Volozh et al., 1975; Brunet et al., 1999; Murzagaliev, 2000).

A distinct regional negative magnetic anomaly corresponds to the Central Precaspian depression on the magnetic field map and the most intense Khobda and Aralsor gravitational maximums (about 40 mgal) coincide with the depression on the gravity map (Brunet et al., 1999; Murzagaliev, 2000). These maximums show a regional level of the gravitational field higher than that of the uplifted structures of the Russian plate.

The basement of the Central Precaspian depression is characterised by abnormally high velocities.



N
Karaton
Tengiz
Yuzhnaya
S
Present
10 km
End of Palaeogene (23 Ma)
Beginning of Palaeogene (65 Ma)
Beginning of Pliensbachian (191 Ma)
End of Kungurian (258 Ma)
End of Artinskian (265 Ma)
Beginning of Artinskian (275 Ma)
End of Early Visean (340 Ma)
Beginning of Late Devonian (375 Ma)
km

Boundary velocities vary from 6.7 to 7 km/s (Fig. 3); layer velocities (in the upper part of the crust) show a stable velocity of more than 6.5 km/s (Volozh et al., 1975), to a maximum of 6.9 km/s in the Khobda area (Kostyuchenko et al., 1999). Thus, the characteristic lower crust velocity of 7–7.2 km/s, observed on the margins, is not present below most of the Central Precaspian depression. Besides velocities, other characteristics of crust and mantle are anomalous. In contrast to other areas, the crust of the Central Precaspian depression is 12–15 km thick. We can recognise two prominent seismic reflectors at the lower levels of the crust, at the depths of 32 and 42 km. A refracting horizon with $V_b = 8.0–8.1$ km/s identified as the Moho-surface coincides with the upper reflector. A reflecting horizon, about 42 km in depth, corresponds to the Moho on the margins (where reflecting and refracting horizons are merged), then extends towards the centre of the basin almost flatly. In the centre of the basin, it is separated from the Moho, which is uplifted, by an 8- to 10 km layer. Analysis of the structure, gravitational field, analysis of density contrast between the different layers and composition of the crust indicates (Volozh et al., 1975; Volozh, 1990, 1991) that the layer between the two described reflectors (32 and 42 km) may be interpreted as eclogites. We propose that the presence of these eclogites was a key factor in the development of the very deep and particular Precaspian Basin. The lower reflector is situated near the upper boundary of an anomaly layer (8.6–8.7 km/s) which was identified by other studies between 45- and 60-km depth (Kostyuchenko et al., 1999 with upper mantle data from Egorkin and Razinkova, 1980).

4. Discussion

The characteristics of the Central Precaspian depression predetermine the special features of the entire Precaspian Basin. These are: (1) the sedimentary cover (important thickness, significant stratigraphic volume, and completeness of the section), (2) the crust (small thickness and absence of low-velocity layer), (3) the mantle (low depth of the Moho surface, presence of another seismic boundary beneath the Moho surface), (4) geophysical fields (high gravitational and very low magnetic fields). That is why all geodynamic hypotheses and formation mechanisms of the Precaspian Basin primarily have to explain the structure and evolution of the Central Precaspian depression.

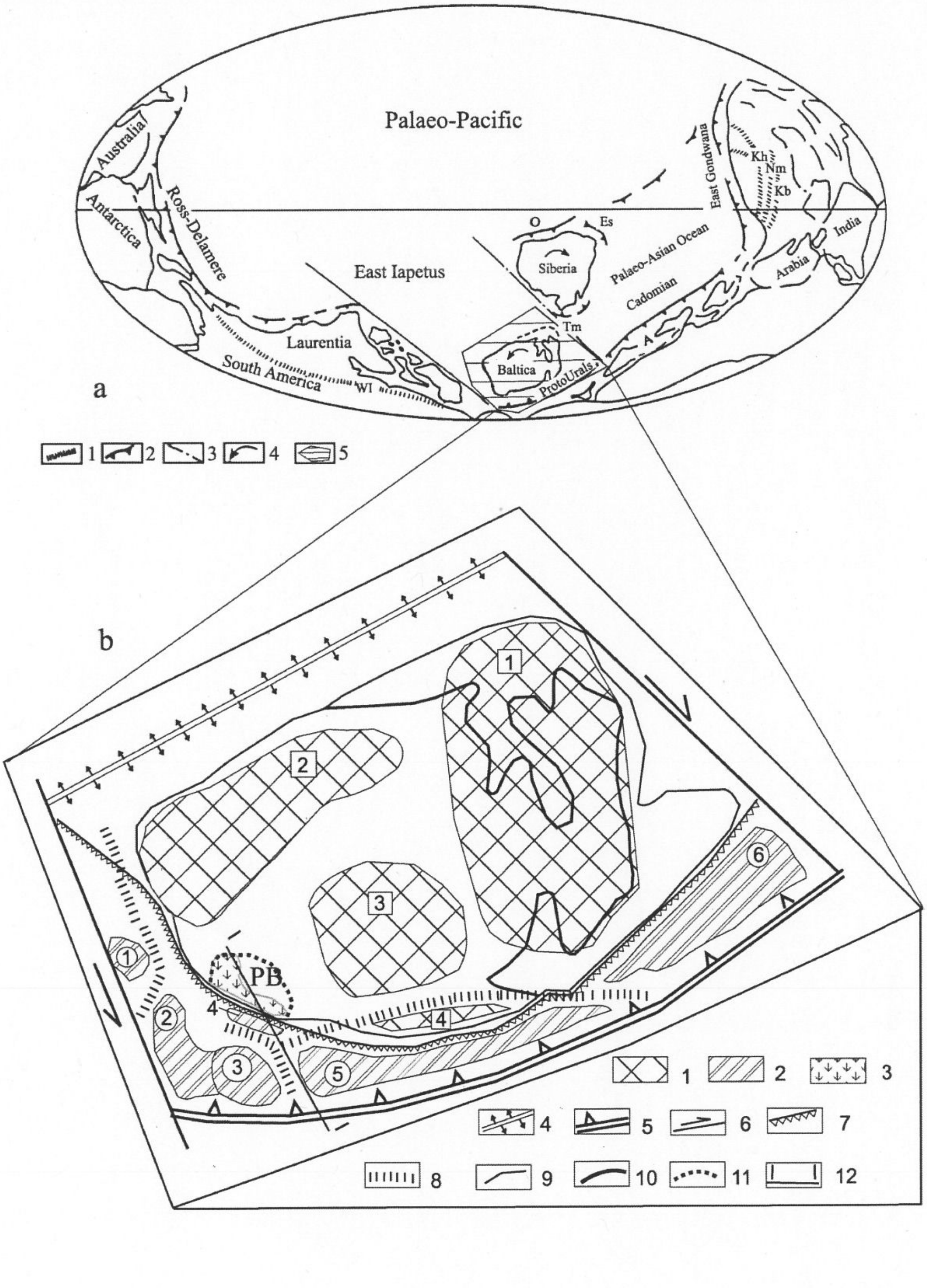
Numerous models have been proposed to explain the nature of this depression. The first group includes models that focus on both the small thickness and high-velocity composition of the crust of the Central Precaspian depression, considering it as an oceanic crust. According to these models, the thin crust of the depression represents the remnant of a marginal oceanic crust which was formed in either pre-Urals or Urals Ocean. In the first case, the crust is Riphean in age (Nevolin, 1978), and in the second, the crust is Riphean to Early Palaeozoic in age (Brazhnikov, 1993; Rikhter, 1997), or even Middle Devonian (i.e., Zonenshain et al., 1990; Volchegursky et al., 1995).

The models of the second group suppose that rifting is the leading mechanism with a thinning of the continental crust during several rifting phases from Riphean to Devonian (Kirukhin et al., 1980, 1981; Kunin 1989).

To explain the subsidence of the Precaspian depression, models of the third group examine the processes and phenomena related to a metamorphic alteration of the crust (Artyushkov, 1993).

It is notable that none of the depicted models has been supported by numerical calculations based on real geological–geophysical data. We have undertaken such an analysis supported by both Peri-Tethys and INTAS Programs. The subsidence of the central part of the basin is illustrated by a set of subsidence curves (Fig. 9). We analysed existing geodynamic hypotheses with consideration of real compositional

Fig. 6. Palaeogeological reconstruction across profile Karaton–Tengiz–Yuzhnaya, southeast Precaspian margin (location in Fig. 1). First section is the Present real section, the following are reconstituted without movements of salt. (1) Sea water; (2–16) stratigraphical units: 2—Palaeogene, 3—Upper Cretaceous, 4—Lower Cretaceous, 5—Middle Upper Jurassic, 6—Upper Permian to Triassic–Lower Jurassic undivided, 7—Triassic–Lower Jurassic, 8—Upper Permian, 9—Kungurian salt, 10—Gzhelian–Artinskian, 11—Middle Visean–Moscovian, 12–Lower Carboniferous–Lower Visean, 13—Middle Devonian, 14—Lower Devonian, 15—Upper Ordovician–Silurian, 16—Cambrian–Lower Mid Ordovician; (17–23) lithological complexes: 17—reefs, 18—carbonate, 19—clayey carbonate, dolomite, 20, 21—terrigenous (sand and clay), 22—clay, 23—sandstones, 24—faults; in the reconstitution, some levels are grouped in average color, $K_1–K_2$, $P_2–T_{1-3}$.


Palaeo-Pacific
Australia
Antarctica
Ross-Delamere
East Iapetus
Laurentia
South America
WI
O
Es
Siberia
Palaeo-Asian Ocean
Cadomian
Tm
Baltica
ProtoUrals
A
East Gondwana
Kh
Nm
Kb
India
Arabia
a
1 2 3 4 5
b
1
2
3
4
5
6
PB
1 2 3
4 5 6 7
8 9 10 11 12

characteristics and physical properties of the crust. This led to more realistic models of the Central Precaspian depression (Brunet, 1995; Brunet et al., 1996). The existing geological data on structure and composition of the sedimentary cover and crust give constraints that must be taken into account. For example, an oceanic crust of Riphean age is too old to explain the Palaeozoic Devonian to Permian subsidence of the basin.

4.1. Age of the oldest sediments

The large age discrepancy for the basin origin results from the uncertainty on the age of the first sediments deposited in the central part of the Precaspian Basin. It should be noted that the base of the sedimentary cover within the areas surrounding the Central Precaspian depression consists of Devonian terrigenous–carbonate. It is the same lithology as that found in the base of the sedimentary pile of the central part of the basin. This similarity of lithology has led to different hypotheses on the age of origin of the basin: either an older Riphean age is proposed or the same Devonian age is attributed to the sediments in the centre of the basin and on its periphery.

But some stratigraphic data exist on the deep horizon (P_3) in the subsalt section. These data seem to exclude all groups of models that explain the formation of the Central Precaspian Basin as a result of Devonian rifting and consequent spreading of oceanic crust. The authors of these models proceeded from the assumption that the deepest main horizon P_3 is restricted to the top of Devonian deposits. They consider also that the thick sedimentary complex (more than 8 km), comprised between this horizon and the basement, is not older than Devonian in age.

The drilling data obtained from some areas in the east and southeast of the Precaspian depression (east Akjar, Kumsai, Baktygaryn, etc.) (Akhmetshina et al., 1993), strongly support the stratigraphic position of the main reflector P_3 as close to the base of Lower Devonian (Fig. 10). This reflector was penetrated by hole G-5 (east Akjar), at a significant distance from the slope of the depression, in the zone of the Aktyubinsk uplift. Within this area, two holes G-1 (Baktygaryn) and G-4 (Kumsai) penetrated thick Devonian carbonate deposits. Logging (GIS—Geophysical interpretation of wells) data determine the stratigraphic position of boundaries. The drilling was stopped at 300 (G1)–500 (G4) m above horizon P_3. Devonian deposits, which are evenly cored in G4, are known in the depth interval of 4830–6007 m. The drilled section contains numerous organic remnants that permit to definitely determine the age of hosting carbonate rocks. The presence of Conodonts species *Ozarkodina remscheidenis remscheidenis* Ziegler in carbonates from hole G-5 (intervals 5738–5745 and 5745–5751 m) is indicative of the lowermost portion of the Lower Devonian Lochkovian stage. The presence of Foraminifers *Tubeporina tenue* Sabirov in Hole G-1 (interval 6204–6212 m) determines the time interval as the Lower Devonian Pragian stage. The fragment of the time-scale seismic section in the vicinity of hole G-5 (Fig. 10) shows that the main reflector P_3 is restricted to the base of the Lower Devonian (Lower Lochkovian) carbonate deposits. As shown in the available substantial volume of seismic data, main horizon P_3 is well expressed in the whole region. It is traced from margin towards centre of basin along almost all seismic profiles. For example, Fig. 11 shows the correlation of P_3 horizon between the southwest Astrakhan margin and the basin.

Seismo–stratigraphic and structural constraints along P_3 horizon within the Precaspian Basin, permit to depict gentle and rounded structures of secondary order (Aktyubinsk uplift, Central Precaspian depression, etc., Fig. 1). In the most subsided part of the Central Precaspian depression, the base of the Lower

Fig. 7. Geodynamic reconstructions of Riphean–Vendian history. (a) Reconstruction of continents in Riphean–Vendian (after Mossakovsky et al., 1998, modified for the vergence of the subduction in area of (b)). 1—Vendian rift, 2—suture, 3—transform fault, 4—direction of rotation, 5—block shown on (b). A: Avalonian, ES: East Sajans, Kb: Karatau–Bajkonur, Kh: Khantajshir, Nm: Nej-Mongolian, O: Omolon, Tm: Tajmyr, WI: West Iapetus. (b) Detail for Proto-Urals ocean 1–2—different blocks: 1—Baltica (*number within square*: (1) Baltic dome, (2) Sarmatian dome, (3) Volga–Urals dome, (4) Uraltau block); 2—Proto-Urals Ocean: (*number within circle*: (1) Moesian, (2) Scythian [pre-Caucasus and Karabogaz domes], (3) Turan [North Ustyurt, Karakum, Central Turkmenia, Zeravshan], (4) East Precaspian block, (5) Turgaj–Syrdarya, (6) Pechora–Barentz Sea); 3—Eclogites; 4–6—East European lithosphere plate boundaries: 4—divergent, 5—convergent, 6—transform, 7—main overthrust Urals folded belt; 8—rifts; 9–11—boundaries: 9—continents, 10—recent continent (shoreline), 11—Precaspian Basin PB; 12—line of profile 11.

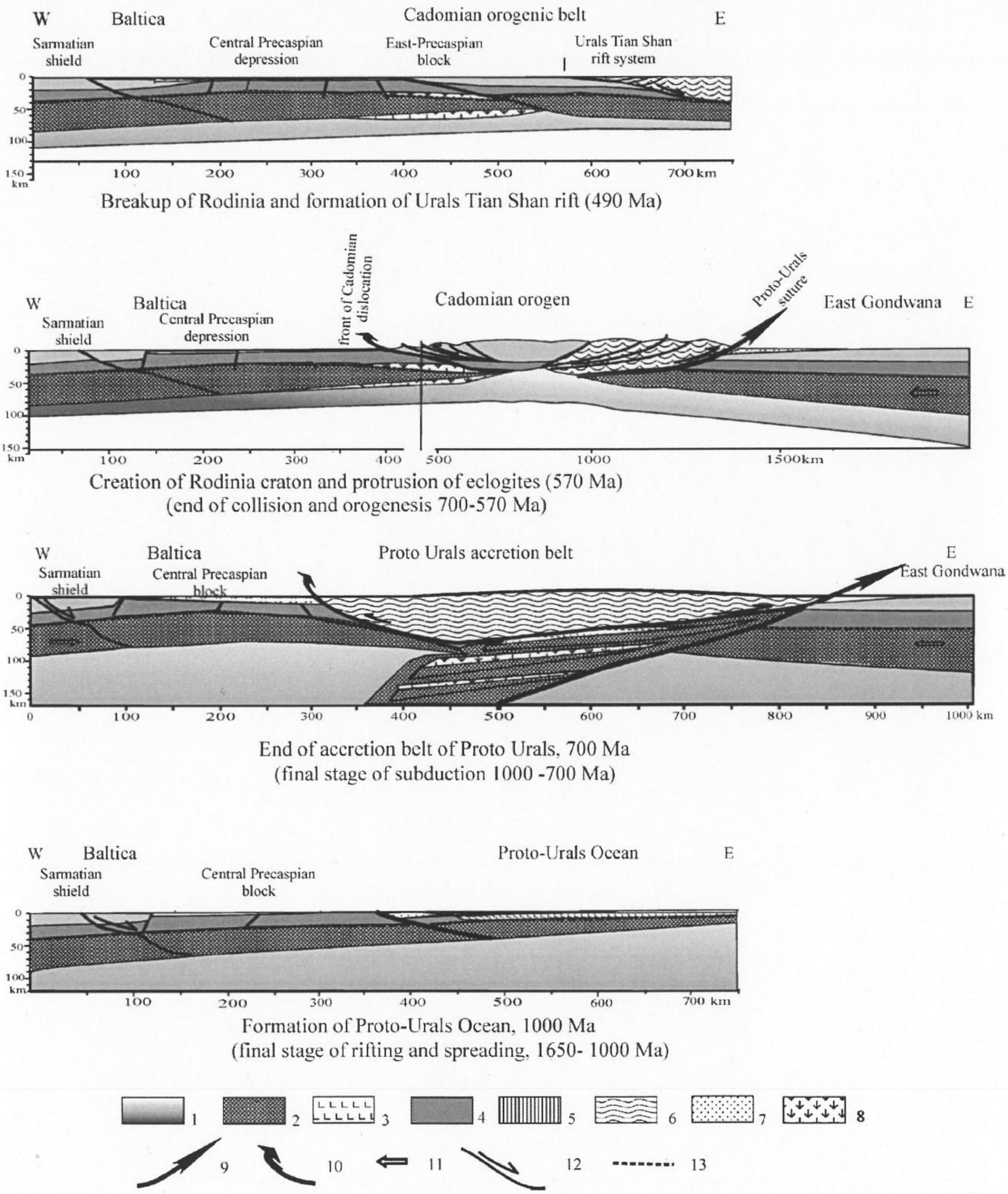


Fig. 8. Riphean–Early Palaeozoic geodynamic reconstruction along an East–West section, (location in Fig. 1) demonstrating eclogites emplacement. 1—Asthenosphere, 2—upper mantle, 3—oceanic crust, 4—lower continental crust, 5—upper continental crust, 6—accretion-collided and subduction–accretion fold belts, 7—sedimentary cover, 8—lenses of eclogites and mantle, 9–10: main faults: 9—suture, 10—front of dislocation, 11—tectonic stress direction, 12—faults with moving direction, 13—Moho boundary.

Table 1

time Ma	Sedimentary environment within Precaspian basin: Central Precaspian block	East-Precaspian block		Duration, Ma	Main geological times[main seismic reflectors]		Geodynamical events within East European lithosphere plate: continent: Old East-European platform	Continental margins: PalaeoAsia sector	Continental margins: Palaeo-Tethys sector
680–540						d_{oc}^{5}	destruction, break up of supercontinent Rodinia	subduction	opening of palaeoTethys
540–450	foredeep basin	orogen		90	Camb–O_2	$P_{\bar{4}}=d_{oc}^{4}$	extension and subsidence	active marginis destruction and formation of triple rift system (Urals, Tian-Shan and Tugarakchan) towards east Central Precaspian basin	passive margin formation
450–408	epicontinental basin	passive margin basin		42	O_3–S	P_3	stabilisation	opening of Urals-Turkestan ocean and formation of the passive margin	accretion of microcontinents and islands arcs
408–390	shelf basin	shelf basin		18	D_1–D_1em		uplift, erosion		accretive fold belt formation
390–375	outer-shelf basin	inner shelf basin		15	D_2–D_3f		rifting	formation of accretive fold belt along Urals margin within East European continent	Sarmatian–Tuarkyr transcontinental shear zone formation
375–312	deep sea basin formation	relatively deep sea environ.: Donbass–Tuarkyr rift (DTR) basin in SW margin	PreMugodzhar and South Emba flysch basin	30	D_3–C_1v_1				accretive fold belt formation
				33	C_1v_2–C_2b	P_2			
312–270	filling the deep basin	folding within DTR	PreUrals molasse basin	13	C_2m–C_2k		uplift: uplift	collision of East-European with Kazakh continents	formation of active Eurasian continent margin along palaeoTethys boundary; and formation of volcano-pluton belt
				34	C_3–P_1a	P_1	stabilisation	formation of Urals–Tian Shan orogen	
270–255	epicontinental basin	epicontinental basin		10	P_1k–P_2t	D	uplift	formation of Urals–Gererud transcontinental shear zone	
255–203	Precaspian-Scythian-Turan intra-continental basin			45	T	V	extension	mobile platform	orogen
203–175	cutting (erosion)			30	J		uplift		formation of Donbass–Tuarkyr intraplate fold system

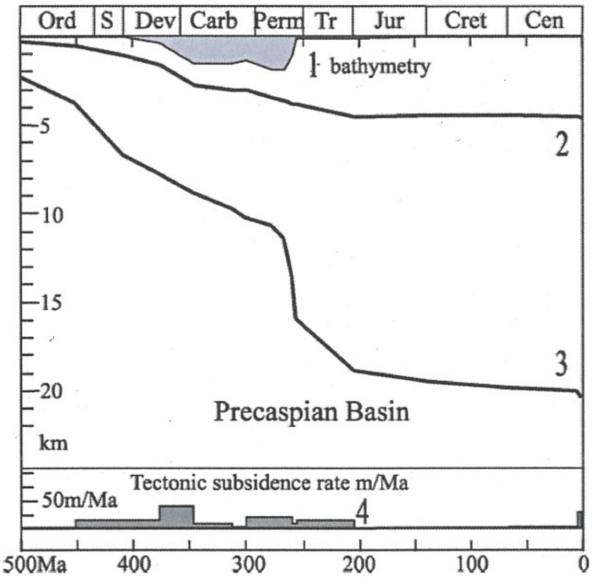


Fig. 9. Subsidence curves in the centre of the Precaspian Basin (after Brunet et al., 1999) at the location of the point C, location in Figs. 1–5, time scale is from Odin (1994). 1: Tectonic air-loaded subsidence curve, 2: total subsidence of the basement with sediments, 3: palaeowater depth, 4: tectonic subsidence rate.

Devonian deposits is situated at a depth of almost 14 km (Fig. 2b) shown by the deepening of the seismic horizon P_3. Hence, a thick (more than 6–8 km) unit of pre-Devonian carbonate and terrigenous rocks is present between the top of the basement and the base of Devonian (seismic horizon P_3) to Lower Permian series.

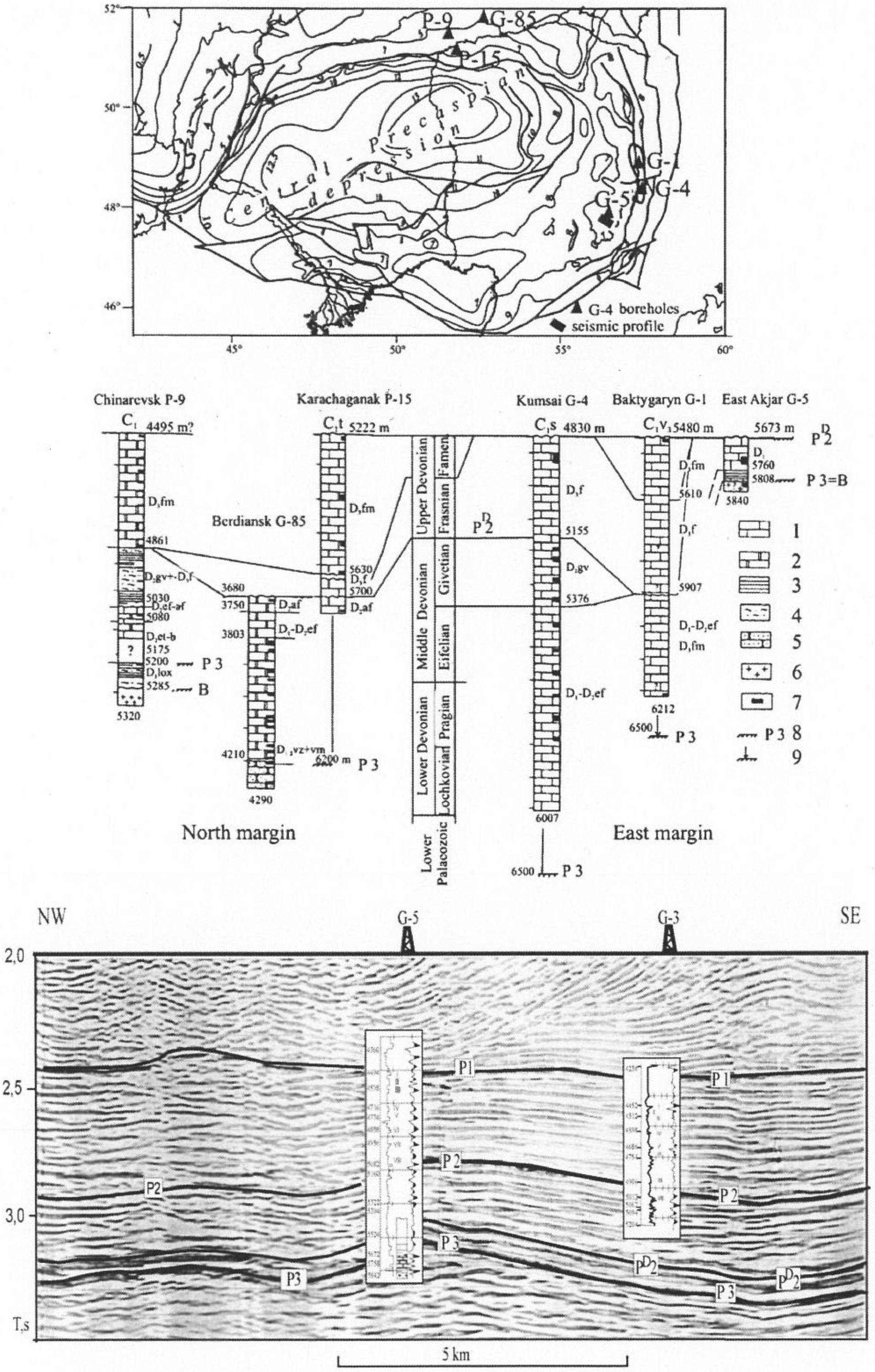
We do not have data constraining precisely the initial age of the sedimentary unit existing below the P_3 Lower Devonian horizon. Below P_3, there exist other seismic horizons (Figs. 3 and 4) P_4, P_5 or D^5_{OS} that were never drilled, and the basement. Sediment deposition began during Riphean, by comparison, for example, with Pachelma rift and Pre-Uralian foredeep where thick Riphean deposits are known. Velocity data and wave images of Riphean deposits were correlated with information from Bashkirian region where seismic records have been carried out and have been published (Scripij et al., 1990). We also have some data on Lower Palaeozoic sediments provided by boreholes on the margins of the basin (Zamarenov, 1989; Volozh 1991; Jatskevich, 1996). Some boreholes in the northwest Precaspian (3—Ershov; 11, 14—Krasnokut; 1, 2—Vostochno–Kudinov; 300—Zhirnov; 10—Petroval'sk) penetrated: Ordovician–Silurian terrigenous and carbonate sediments (about 100 m) containing Brachiopods, Trilobites, Corals and Crinoids. In the northeast Precaspian, boreholes (85—Berdiansk; 110—Predurals) drilled Cambrian–Ordovician terrigenous and carbonate sediments (about 200–300 m) with Graptolites, Algae and Ostracods. Jatskevich (1996) wrote that Upper Proterozoic sediments were covered by thick (1000–2000 m) post-Vendian deposits (Cambrian, Silurian and Ordovician) in Riasano–Saratov basin (in the former situation of the Pachelma Riphean Basin and in prolongation of the Precaspian Basin). These Lower Palaeozoic sediments were eroded during pre-Devonian time. Thus, the thickness of Lower Palaeozoic terrigenous sediments is 100–300 m and carbonate sediments 100–200 m as well in northwest as in northeast. Palaeoreconstructions show that Lower Palaeozoic sediment thickness increased towards the basin up to 3.5–4 km and that older layers appear below these deposits.

Even if the age of first sediments is still partly undefined, the existing data do not support a Middle Devonian spreading, since older sediments cannot cover a Devonian oceanic crust.

4.2. Implementation and role of eclogites

We propose (Volozh, 1990, 1991) that the high-velocity lenses recognised by geophysical data correspond to eclogites and are an important cause of the subsidence of the basin. Thus, we explored the possibilities of timing and process of their implementation. Our first idea was a transformation into eclogite at the base of the crust during either the Riphean or the Mid-Late Devonian. This second hypothesis was

Fig. 10. Stratigraphical site of seismic reflector P_3 (by Akhmetshina et al., 1993, with modification). Comparison of levels reached by wells on the north and east margins of the Precaspian Basin; location of wells on a map of seismic horizon P_3 depth; fragment of seismic line on the east margin, through the wells G5–G3. 1: limestone, 2: dolomite, 3: argillite, 4: aleurolite, 5: sandstone, 6: metamorphic (basement), 7: samples with fauna, 8: reflectors and their index, 9: seismic reflectors offside borehole. Horizons: af–Afonian, bs–Biiskian, vs–vn Viazov–Vaniashkian, cv–lv Evlano–Livennian.


P-9
G-85
P-15
Central - Precaspian depression
G-1
G-4
G-5
G-4 boreholes
seismic profile
52°
50°
48°
46°
45°
50°
55°
60°
Chinarevsk P-9
Karachaganak P-15
Berdiansk G-85
Kumsai G-4
Baktygaryn G-1
East Akjar G-5
Upper Devonian
Middle Devonian
Lower Devonian
Lower Palaeozoic
Famen
Frasnian
Givetian
Eifelian
Pragian
Lochkovian
North margin
East margin
P 3
P 3=B
B
NW
SE
G-5
G-3
P1
P2
P3
2,0
2,5
3,0
T,s
5 km

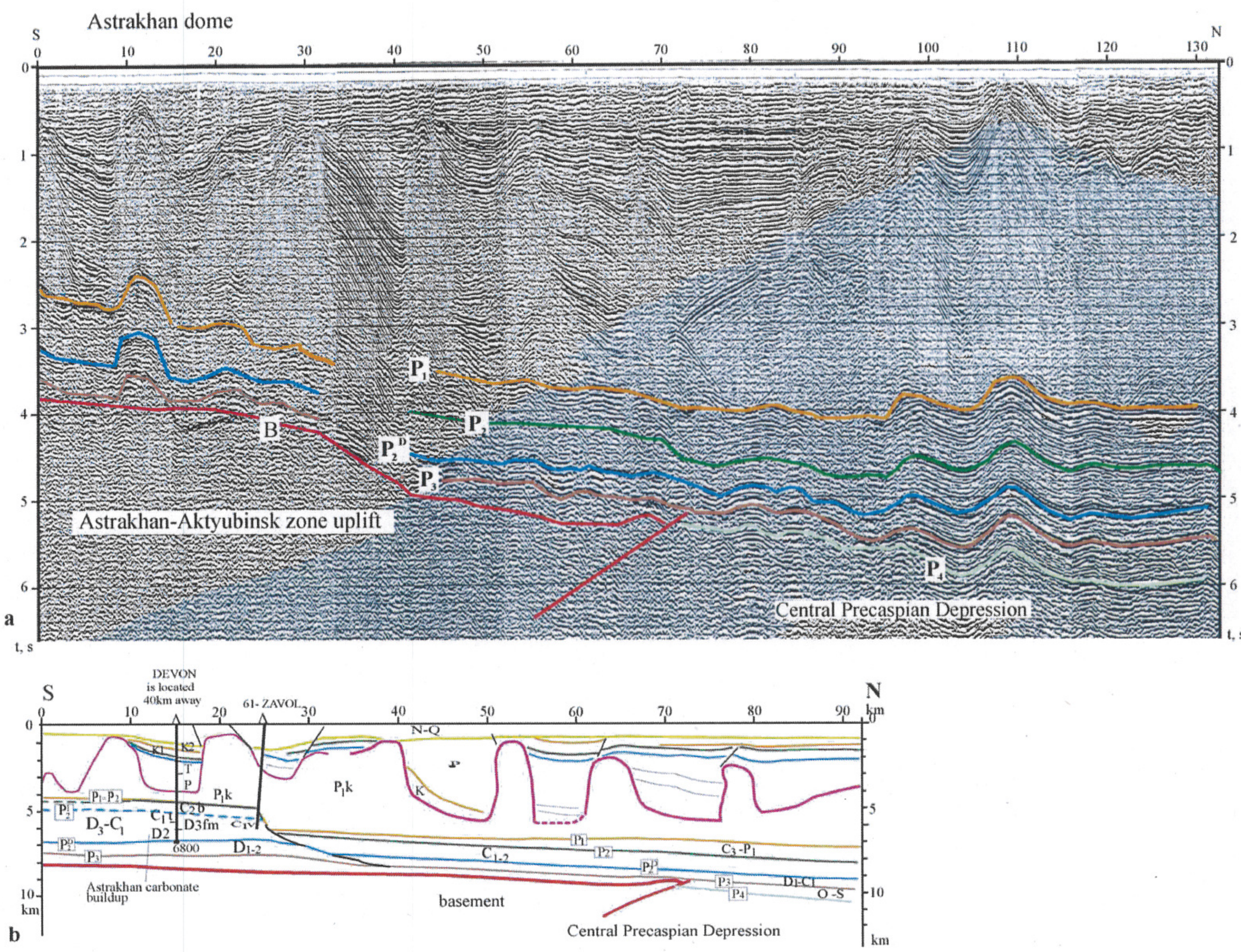


Fig. 11. North–south section on the southwest Astrakhan margin of the Precaspian Basin. (a) Seismic profile; (b) depth converted section of a part of the profile (a).

proposed to match the age of eclogites found in the Urals (Matte et al., 1993; Chemenda et al., 1997; Hetzel, 1999).

The assumption of transformation of gabbro (ρ_g= 2.9–3 g/cm^3) to eclogite (ρ_{ec}=3.5–3.6 g/cm^3) (Artyushkov, 1993) at the base of the crust, is not sufficient to entirely explain the formation of the Precaspian depression. The initial thickness of lithospheric layer h_{tr} subject to transformation is taken to be equal to 8 km and the respective density of sediments, asthenospheric mantle, eclogite, gabbro are: ρ_s=2.47 g/cm^3 (average density of the sedimentary column in the centre of the Precaspian Basin, Brunet et al., 1999), ρ_a=3.2 g/cm^3, ρ_{ec}=3.55 g/cm^3, ρ_g=2.9 g/cm^3. Then using the expression for total subsidence (with sediments) of the basin W in the form (Artyushkov, 1993):

$$W = h_{tr} x \rho_a(\rho_{ec} - \rho_g)/\rho_{ec}(\rho_a - \rho_s),$$

we find that W is less than 7 km and could only explain a part of the 20 km of sediments deposited in the basin.

Moreover, this model is not satisfactory, as it lacks arguments in favour of the reality of eclogitisation in the lower crust where pressure conditions are not sufficient. Indeed, we can estimate the thickness of the column, over the layer candidate to eclogitisation. It includes a few kilometres of consolidated deposits, and a crustal layer up to 20 km thick (including the portion of the crust to be possibly transformed into eclogites). Thus, the total thickness of the crust plus sediments would have been less than 30 km. Such parameters of the crust result in thermal and pressure conditions prohibiting eclogitisation, as the latter needs pressures not less than 15–20 kbar according to experimental data (Ringwood, 1982). The transformation of a mafic subcrustal body to eclogite was proposed by Fowler and Nisbet (1985) for the Williston intracratonic basin. However, the conditions were very different: around 55-km thick crust and few sediments (some kilometres), and hence minor tectonic subsidence was induced by eclogites load in comparison to the Precaspian Basin's subsidence.

Another model of eclogites formation was proposed by Lobkovsky et al. (1996a,b) and Ismail-Zadeh et al. (1997) with the accumulation of magmatic melt in the asthenospheric bulge created by an extension, a phase transition to eclogite. This model is more specifically proposed at the base of the lithosphere, and seems to be unlikely for large masses of eclogites at the base of the crust, where it also lacks solid physical argumentation.

After analysing existing problems, we revised our previous concepts on when and how eclogites appeared at the base of the crust of the Central Precaspian depression.

From the current concept, we have kept the main ideas that follow. As we have illustrated above (Fig. 3), lenses of eclogites indicate distinct lower surfaces. A clearly expressed reflector is related to these surfaces. This reflector has a sub-horizontal surface which is restricted to the same depth (40–42 km) as that of the base of the crust outside the supposed eclogite-bearing zone. Spatially, lenses of eclogites tend to occur along the zone of the deep-rooted fault that borders the Central Precaspian depression to the southeast. The plane of this fault dips to the southeast. The North Caspian–Aktyubinsk terrane with a Riphean basement is overthrusted along this plane on the Central Precaspian block that contains pre-Riphean thinned crust (Figs. 3 and 8).

According to seismic data, the fault is sealed with a Lower Palaeozoic association of rocks. Thus, we consider it to represent a front marking the Cadomian deformation during the creation of the Rodinia continent and the formation of the accretive Proto-Urals belt (Figs. 3, 4, 7 and 8).

These data support the following scenario. In the Central Precaspian depression, crustal rocks subducted during Riphean, have been transformed into eclogites at depths of 50–60 km. The eclogite lenses, returned from depth, may have been emplaced at the Crust/Mantle Boundary (CMB=Moho), during the end of Late Riphean through Early Vendian. Fig. 8 illustrates a possible context and timing of this process. It is supposed to occur along the boundary between the Central Precaspian depression and the Proto-Urals folded belt, as a result of microplates and island arcs accretion to the Baltica active margin.

On the Canadian margin (Clowes et al., 1995), seismics investigated in details deep structures of crust and lithosphere in other folded belts of such type. Reflected and refracted seismic deep profiling data show specific structural lithosphere features within these belts. The lower lithosphere part, beneath the Moho, consists of crustal nappes and packages of mantle plates as thick as 50 km. It is a collage of eclogitised rocks with increased viscosity, which is located hypsometrically below the lithospheric mantle of the subducting edge of the continental plate. During collision, thick eclogitised rocks of the lower part of hanging block of the orogenic belt find themselves between more rigid lithospheric blocks, split and rise upward, protruding into the CMB. By analogy of our data with these investigations, a similar mechanism of evolution of the Precaspian Basin is proposed during the Riphean–Vendian interval of its history (Fig. 8).

The eclogites of the Precaspian Basin are not directly injected along the thrusting and not exhumed to the surface as recognised now in many mountain belts. They are only bounded by an important thrusting and could be injected or underplated below the crust. Ellis and Beaumont, (1999) show a similar geometry with underplated mafic material above the subduction zone of Cascadia.

The mechanisms which have been proposed for eclogites exhumation in the belts are numerous in literature: erosion, extension, buoyancy, return flow, pinching and others (i.e., Platt, 1993; Matte and Chemenda, 1996; Chemenda et al., 1997; Garagash et al., 1997; Ring et al., 1999; Leech, 2001). However, in general, the question of the exact mechanism of the eclogites emplacement remains partially unsolved. Specific unanswered questions in the Precaspian Basin are the considerable size of the eclogites lenses (Fig. 1) and their implementation through a less dense mantle.

4.3. Need of an improved model of subsidence of the Central Precaspian depression caused by the weight of eclogite lenses and of a dense thin crust

As discussed, no published model on the formation and evolution of the Precaspian Basin can explain all the features of its structure. A new model is needed to satisfactorily describe the following main features of structure and development of the Precaspian Basin:

(a) important thinning of the continental crust and presence of uplifts in its lower surface filled with high velocity and high density material thought to be eclogites lenses according to our interpretation;
(b) preservation at the same location of the area of maximal subsidence (at least during 600 Ma) within the Central Precaspian depression. This testifies of the accommodation to a source of forces responsible for subsidence, inside the lithospheric plate;
(c) absence of evidence of appreciable change in crustal structure during the Middle Carboniferous–Triassic time, even if it is a stage of very active evolution of the Precaspian Basin sedimentation;
(d) observable amplitudes of general subsidence and bathymetry changes of the basin.

The model we propose, provides basin subsidence partly caused by weight due, on one hand to eclogite lenses injection at the base of the crust, and on the other hand by the transformation from a normal continental crust to a thin and heavy crust.

Fig. 12 shows a simple isostatic balance between an initial "normal" 40-km crust and the column we observed, composed of a thin heavy crust ($\rho_c = 2.93$ g/cm^3), as indicated by seismic data ($p = 6.9$ km/s). In this way we consider the results of the different driving mechanisms which have induced the amount of total subsidence but not the timing of their action nor if the events (rifting) where polyphase or monophase.

Two hypotheses have been explored. The first hypothesis (Fig. 12) considers that the mantle is underlying a thinned high velocity and density crust (= absence of eclogites), the subsidence produced is around 4 km. The isostatic balance shows that the tectonic subsidence obtained for this configuration is slightly less than the observed subsidence (Fig. 9). In the second case, the presence of an 8-km eclogite layer is adopted at the bottom of the observed thinned anomalous crust in the centre of the basin. We obtain 0.8 km of additional tectonic subsidence and thus a total value very close to the observed subsidence (derived from subsidence curve, Fig. 9). The supplementary load of the heavy lens would correspond to the deposit of approximately 3.8 km of sediments of density 2.47. A gravity model is in progress (Brunet et al., 2000, 2001, manuscript in preparation), testing the hypothesis of eclogite presence by comparing anomalies produced by eclogites or mantle material below the crust. It indicates that heavy load is necessary at depth,

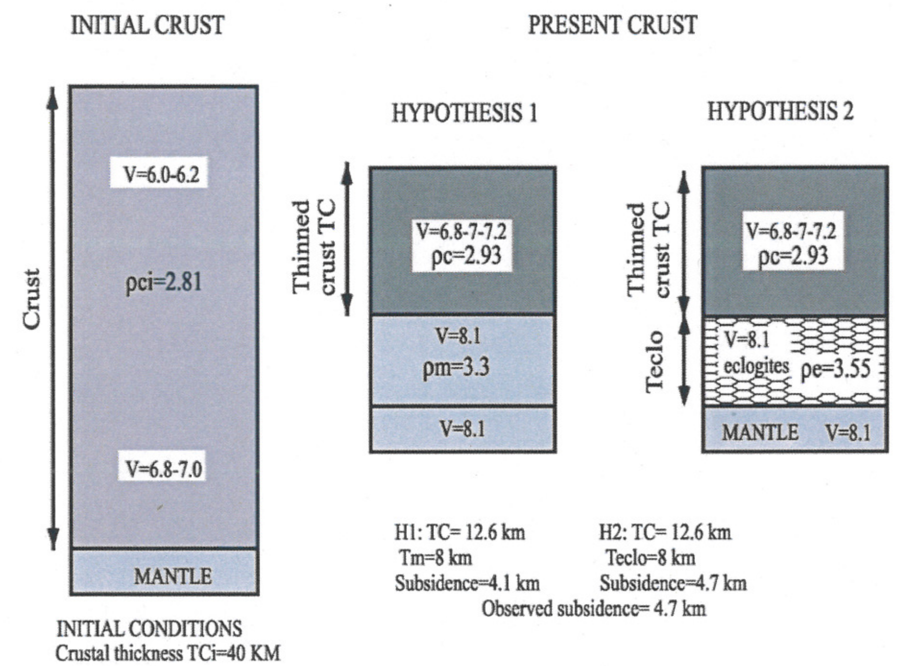


Fig. 12. Isostatic balance between an initial crustal column and the present day crustal column in the middle of the basin. Two hypotheses are tested: the 8-km layer below the crust is considered: (1) as upper mantle, (2) as eclogites. The results are compared with the tectonic subsidence from geological data (subsidence curve Fig. 9) in the centre of the Precaspian depression.

to compensate the 20 km of low average density (2.47) due to the presence of salt.

During Late Riphean–Early Vendian (our hypothesis for timing of eclogites implementation), the amount of sediments deposited does not seem to exceed 2–3 km. This is a little less than the 3.8 km expected from the load of the 8 km layer of eclogites. It means that either the adopted load is too large (too important thickness or density of eclogites) or the timing considered is wrong. A younger eclogite implementation would correspond to periods of more important sedimentation. Another possibility is that one part of the potential of subsidence due to the eclogites weight is not fully expressed at once at the implementation. Rocks may be submitted to an increase of density subsequent to implementation (cooling change of phase with fluids intervention). Subsidence may be also partly counterbalanced by in-plane tangential compressional stresses acting on the plate. The latter solution is retained by Garagash et al. (1997) as the Precaspian Basin was continuously (Table 1) in a zone of active influence of tangential compressional stress during almost 400 million years (600–203 Ma) with some release during the extensional phases. Appreciable influence of the tangential stress field has ceased on the Precaspian Basin subsidence around 200 Ma, after the accretion of the young Scythian and Turan platforms to the south and southeast of the Precaspian Basin. Since then, the Central Precaspian depression has become an isostatically balanced structure.

We have chosen a more realistic model combining several mechanisms through time, which better match existing geological and geophysical data. The model suggests the succession of the main geologic events in the past as follows: Riphean, Ordovician and mainly Devonian rifting phases producing the thinning of the crust, probably a compressive context during the Uralian orogeny and loading influence of the high-velocity layer at the base of the crust. The main mechanism driving the basin subsidence is an isostatic adjustment to the change of crustal state. It results from rifting, allowing an important reduction in thickness and the increase of the average density of crust through the disappearance of the low velocity upper crustal layer. The presence of eclogite lenses below the Precaspian Basin may explain only a small part of the 20-km sedimentary deposits in the Central Precaspian depression. Other tectonic events are less important in their consequence on the amount of tectonic subsidence produced. These are a collision at the end of the Riphean, allowing a small amount of flexural foreland deposits, some possible other rifting phases during Carboniferous and Permo-Triassic and finally the consequence of the propagation of Uralian compressive forces and mantle movements leading to transient accelerations of subsidence (Brunet et al., 1999).

5. Conclusions

(1) The main, most remarkable, structural features of the sedimentary cover and crust of the Central Precaspian Basin are: (a) a large sedimentary cover thickness, (b) a reduced thickness of the crust and increase of its average density, and (c) a possible presence of high-velocity lenses at the base of the crust interpreted as eclogites. These features have originated before the formation during Permian of the saliferous basin (as an isolated area of sedimentation) and were connected with Riphean–Early Palaeozoic then Devonian history and evolution of the Precaspian region.

(2) We present a new model of formation and evolution of the Precaspian Basin. This model fits the available geologic–geophysical data on its deep structure. The eclogite loading may explain a part of the subsidence but other mechanisms (mainly the crustal thinning due to rifting and increasing of crustal average density), are required to explain the remaining major part of the subsidence.

(3) As this model shows, basins like the Precaspian basin are rather unique. Another example of a superdeep basin with around 20 km of sediments is the South Caspian basin (Brunet et al., 2003). Their origin needs a favourable combination of several geodynamic factors. Such conditions appear to exist within external corners of large continents bordered by accretionary folded belts.

Acknowledgements

The study was undertaken in the context and with the support of the international Programs INTAS and Peri-Tethys. The study was also supported by the Russian Foundation for Fundamental Research (grant no. 00-15-98531). Our thanks go to Andrey Ershov who made the balanced reconstruction of the section near the Urals. We are grateful to Jan Hegner and two anonymous reviewers for their constructive comments on a first version of this paper and to Sierd Cloetingh for his help in english editing.

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