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Accommodating large-scale intracontinental extension and compression in a single stress-field: A key example from the Baikal Rift System

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ABSTRACT

The Baikal Rift System in southern Siberia is one of the main intracontinental extensional features on Earth. The rift system represents the northwestern boundary of the Amuria plate and in that respect can be considered as an evolving plate boundary. The Baikal Rift System has been widely studied both in terms of geology and geophysics and many models have been proposed for its formation and evolution. However, the age of the initiation of deformation and the mechanism driving this deformation are still largely debated. While major extension has occurred since the Late Miocene–Pliocene, the onset of extension seems older than the India–Asia collision, implying that several driving mechanisms may have acted together or in relay through time. In this work, we review the available data and models for deformation in an area encompassing the Baikal Rift System, the Sayan ranges to the west and the Transbaikal to the east. Using a synthesis of this data and our own field and mapping observations, we show that the Baikal Rift System, along with transpressional deformation in the Sayan ranges and transtension in the Transbaikal area, can be explained through major left-lateral strike–slip systems. The deformation is strongly controlled by inherited crustal and lithospheric structures, and is distributed over a wide area within the western Amuria plate that consequently cannot be considered as a rigid block. Such distributed deformation is likely to have a strong effect on the structure of the future continental margin if extension evolves towards the formation of oceanic crust.

1. Introduction

Together with oceanic or intracontinental subduction, extension and subsequent rifting or the propagation of major lithospheric strike–slip faults are the main mechanisms that can generate new tectonic plate boundaries. While various rifting processes have been widely investigated and documented (e.g., Polyansky, 2002; Zorin et al., 2003; Chorowicz, 2005; Petit and Déverchère, 2006; Gueydan et al., 2008; Schmeling, 2010), plate fragmentation by propagating strike–slip faults in a compressional tectonic setting is still poorly understood.

In Siberia, the Baikal Rift System (BRS) is considered as a narrow rift structure (Buck, 1991) extending from the southeastern edge of the Siberian craton up to the southern edge of the Aldan craton (Fig. 1). The BRS forms the northwestern limit of the Amuria plate and accomodates its progressive eastward motion away from the Siberian craton (e.g. Polyansky, 2002; Petit and Fournier, 2005; Petit and Déverchère, 2006). Rifting along the eastern edge of the Siberian craton occurs in parallel with transpressive deformation affecting the Mongolian crust in the Sayan ranges to the west (e.g. Delvaux et al., 1997; Arzhannikova et al., 2011; Jolivet et al., 2011) (Fig. 2). To the east, the Transbaikal area extends from Lake Baikal towards the borders of Mongolia and North China (Fig. 2). It is characterised by a strong tectonic fabric and especially by Mesozoic grabens that developed in a NE–SW direction mostly parallel to the general direction of the BRS (e.g. Delvaux et al., 1995; Tsekhovsky and Leonov, 2007) (Fig. 2). Evidence for active deformation (mainly normal faulting) has been reported for some of these basins close to the BRS (e.g. Bulnaev, 2006; Chipizubov et al., 2007; Lunina and Gladkov, 2007, 2009). This kinematic setting requires that the generally N–NE motion of the lithosphere in western Mongolia (with respect to the stable Siberian craton) is sharply transformed to a SE direction in the adjacent Transbaikal area.

Many models have been proposed to describe the rifting process in the BRS and to explain the transition between transpression and extension near the southern tip of the Siberian craton (e.g. Jolivet et al., 2009; Seminsky, 2009 and references therein; Mats and Perepelova, 2011 and references therein). The mechanism driving extension is still

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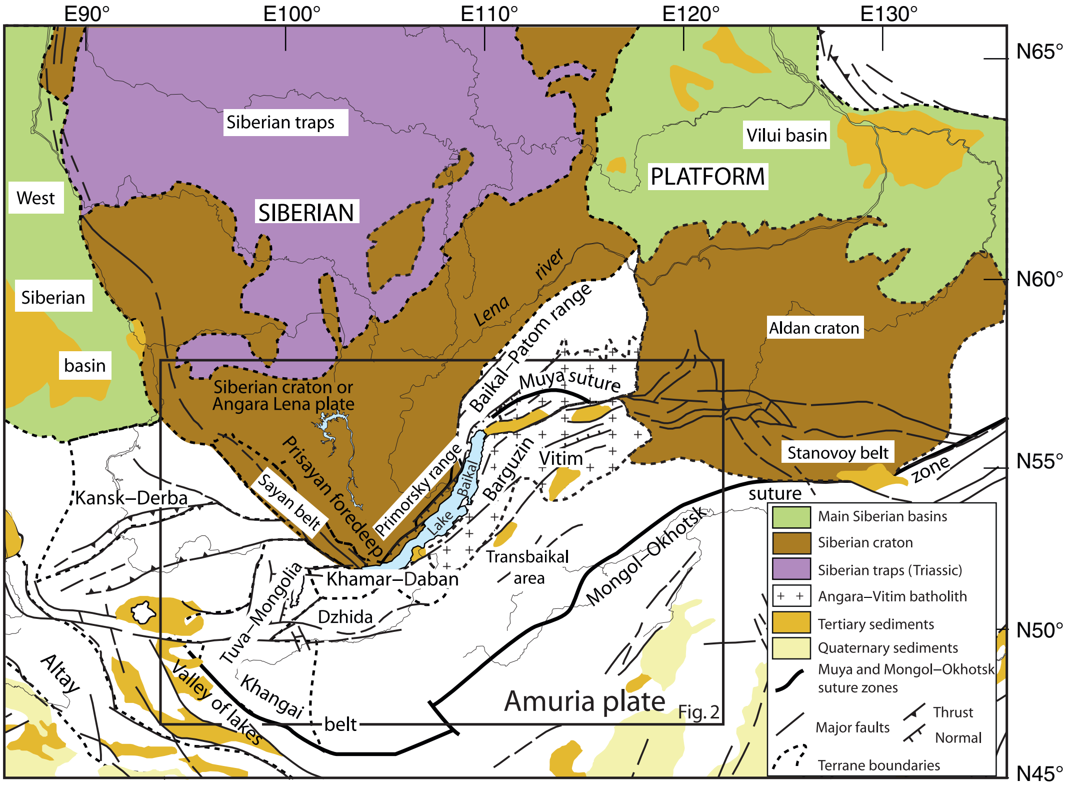

Fig. 1. General map of the various terranes and tectonic structures of eastern Siberia and northern Mongolia. The position of the various terranes and structures are based on Delvaux et al. (1995, 1997), Gusev and Khain (1996) and Malitch (1999). The black box indicates the localtion of Fig. 2. Modified after Jolivet et al. (2009).

largely debated as being either an asthenospheric diapir acting on the base of the lithosphere (e.g. Logatchev and Zorin, 1987; Windley and Allen, 1993; Kulakov, 2008), or extrusion of the Amuria plate driven by the India–Asia collision to the south (e.g. Molnar and Tapponnier, 1975; Hus et al., 2006; Petit and Déverchère, 2006). However, all available models recognise that the opening of the BRS and the peculiar stress field at the southern end of the rift are largely controlled by inherited crustal and lithospheric structures (e.g. Delvaux et al., 1997; Chemenda et al., 2002; Polyansky, 2002; Zorin et al., 2003; Hus et al., 2005, 2006; Petit and Déverchère, 2006; Jolivet et al., 2009; Mats and Perepelova, 2011). Some authors also consider the BRS as a major strike–slip system connecting the Tunka–Mondy strike–slip fault with the North Baikal basins system (Polyansky, 2002; Seminsky, 2009; Rasskazov et al., 2010; Mats and Perepelova, 2011) (Fig. 2).

In this contribution, we provide a comprehensive review of the various models and data relating to the tectonic deformation and crustal structure of a wide region encompassing northwestern Mongolia, the Sayan ranges, the BRS, the Transbaikal region, the North Baikal basins system and the southern edge of the Aldan craton (Fig. 2). This review is supplemented by our own field observations and mapping obtained in the Transbaikal area. Our investigation reveals that the faults that controlled the formation of the Mesozoic grabens have been reactivated not only close to the BRS but also to a considerable distance (several 100 km) away from the active extension zone. This implies that extension in the northwestern part of the Amuria plate is not restricted to the narrow rift structure of the BRS, but affects a significantly wider area.

Based on our newly-obtained data and existing published models, we construct a revised model for the tectonic evolution of the BRS and adjacent areas. We show that both transpressive deformation in the Sayan ranges and extension in the BRS and Transbaikal area can be explained by movements along major strike–slip systems linking the Altay range to the Stanovoy ranges. As already suggested by Tapponnier and Molnar (1979), therefore, Late Cenozoic extension in the BRS and the Transbaikal is produced solely by the interaction between these strike–slip systems and the inherited crustal and lithopsheric structural architecture. Finally, we show that the onset of extension in the BRS occurred prior to the onset of the India–Asia collision, and could be related to the geodynamic event that generated the Mesozoic Transbaikal basins.

2. Geological setting: the pre-Neogene tectonic history of SE Siberia

The crustal and lithospheric structure of southeast Siberia and north Mongolia are complex (Fig. 1), resulting from a series of collision and rifting events. A synthesis of this evolution is given in Jolivet et al. (2009) and only the main episodes are summarised below.

The basement of the Siberian craton grew during the Palaeoproterozoic (1900–1700 Ma) by accretion of various continental blocks, inducing compressive deformation, metamorphism and plutonism (e.g. Khain and Bozhko, 1988; Dobretsov et al., 1992; Delvaux et al., 1995; Gusev and Khain, 1996). A rifting phase during the Ectasian to Tonian (Riphean), associated with the opening of the Palaeo-Asian Ocean, led to the formation of intracontinental basins and a passive margin along the southern edge of the craton (Zonenshain et al., 1990a,b; Dobretsov et al., 1992; Belichenko et al., 1994; Delvaux et al., 1995; Gusev and Khain, 1996). Inversion of that margin initiated during the Tonian (Late Riphean) (Gusev and Khain, 1996), followed by accretion of various terranes such as the Barguzin and Khamar Daban blocks (Fig. 1) during the Ediacaran (Vendian)–Early Cambrian (e.g. Berzin and Dobretsov, 1994; Delvaux et al., 1995; Gusev and Khain, 1996) and the formation of large foredeeps along the southern margin of the Siberian craton. These foredeeps accumulated sediments until the Early Silurian (Melnikov et al., 1994). In the Late Cambrian–Early Ordovician (530–485 Ma), the Tuva–Mongolia block (Fig. 1) collided with the

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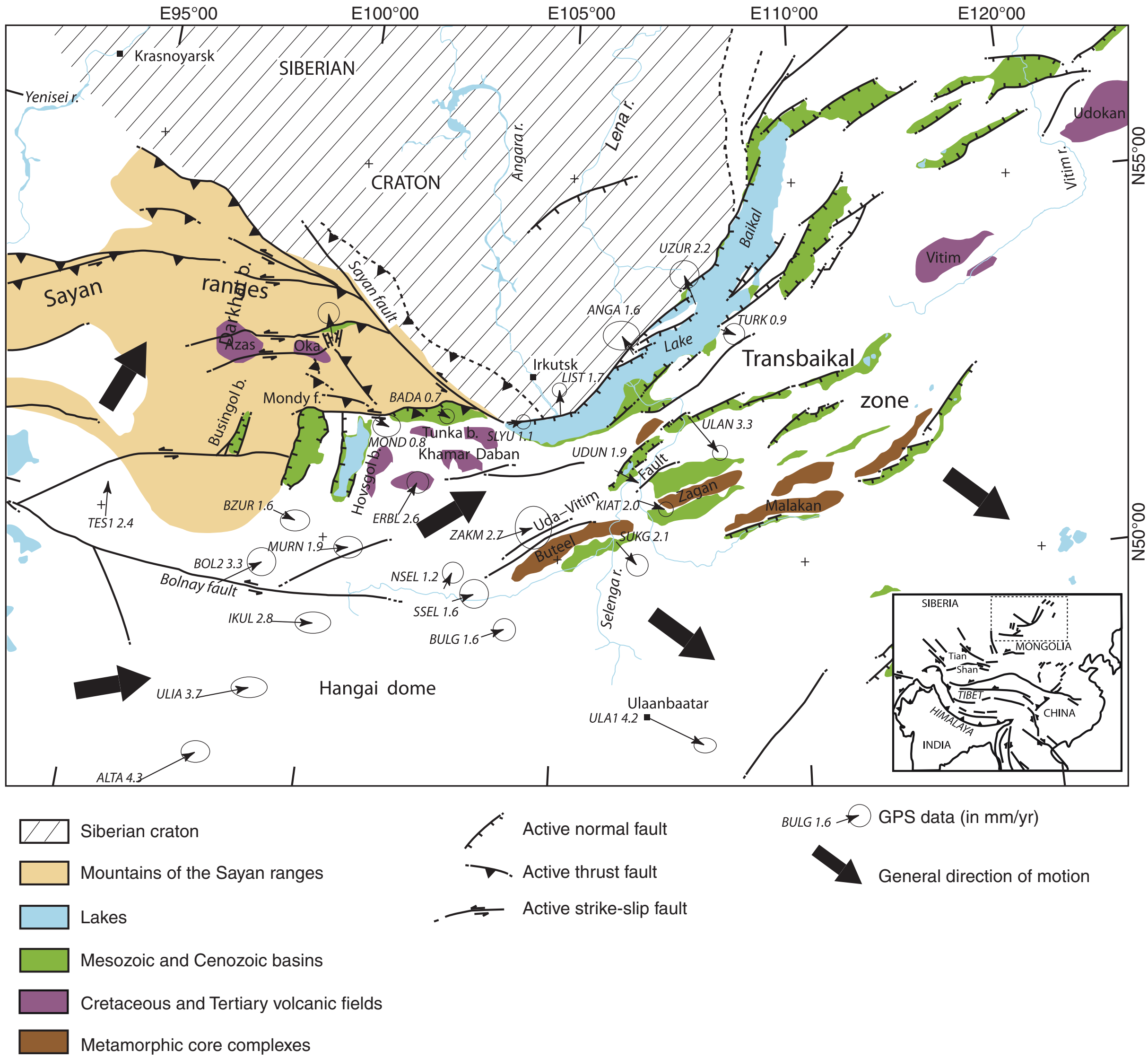

Fig. 2. Geodynamic and tectonic setting of the Sayan–Transbaikal area. Mapping of the general tectonic structures highlights the opposite deformational styles in the compressional Sayan ranges to the west and the extensional Transbaikal area to the east. Small arrows with confidence circles are GPS velocities. Large black arrows indicate the mean regional direction of lateral motion in the crust estimated from GPS data and fault geometry and kinematics. GPS data are from Calais et al. (2003). b. stands for basin, f. for fault and r. for river. Modified from Jolivet et al. (2011).

Siberian craton, producing a large regional metamorphic event along its eastern margin (Bibikova et al., 1990; Bukharov et al., 1992; Sryvtsev et al., 1992; Fedorovskii et al., 1993). During the same period (Ediacaran–Early Cambrian), island-arcs formed to the west in the Gorny Altay (the northwestern termination of the Altay range) and were either subducted or accreted to Siberia during the Late Cambrian–Early Ordovician (480–490 Ma) via subduction of the Palaeo-Asian Ocean (Glorie et al., 2011). To the east, the Late Cambrian–Early Ordovician phase was followed by a Late Silurian–Early Devonian deformation phase, affecting the southeastern and eastern margins of the craton and probably related to the collision of the Dzhida island arc (Fig. 1). The absence of post-Silurian sediments makes it very difficult to describe the evolution of the Siberian craton during the Middle–Late Palaeozoic. However, since Ordovician and Silurian deposits are only weakly deformed, the craton must have remained relatively stable (e.g. Gusev and Khain, 1996; Cocks and Torsvik, 2007). Palaeozoic apatite fission track ages obtained along the eastern margin of the craton confirm this long-lasting stability (Jolivet et al., 2009).

Within the Transbaikal region (Figs. 1 and 2), final subduction of the Palaeo-Asian Ocean generated the 339–285 Ma Angara–Vitim granite batholith emplaced in the Dzhida, Khamar Daban, Barguzin and Stanovoy regions (Delvaux et al., 1995; Yarmolyuk et al., 1997) (Fig. 1). The closure of the Palaeo-Asian Ocean was also responsible for the development of the main tectonic structures in the Baikal–Patom and Zhuya fold and thrust belts (Fig. 1).

During the Early Permian, western Mongolia collided with Siberia, marking the beginning of the closure of the Mongol–Okhotsk Ocean (Fig. 1) (Nie et al., 1990; Zonenshain et al., 1990a,b). Until the Early Jurassic, continuous northward subduction of the Mongol–Okhotsk oceanic crust is attested by granitoid magmatic activity (Filippova, 1969; Zorin et al., 1990). However, the age of the final closure of the ocean and the occurrence of a Mongol–Okhotsk collisional range in the Transbaikal area are still largely debated (see Jolivet et al., 2009 and references therein). While the switch from marine to continental sedimentation in the Transbaikal region (Mushnikov et al., 1966; Ermikov, 1994) and recent low-temperature thermochronology data

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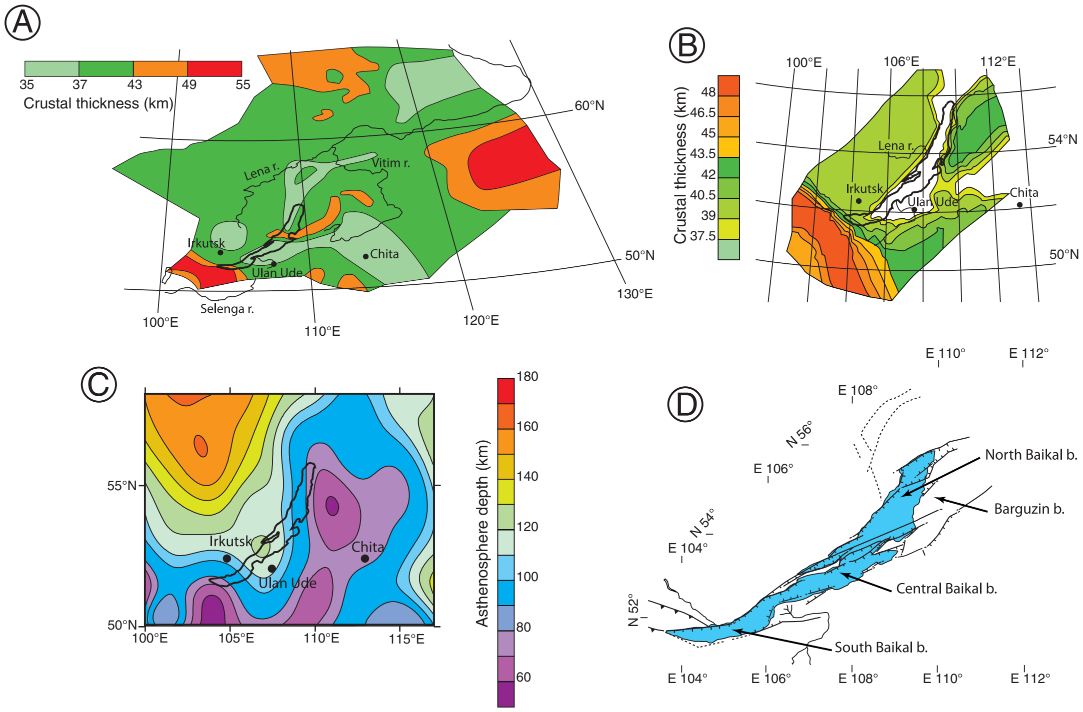

Fig. 3. Crustal and lithospheric thickness of the study area modelled from geophysical data. Figs. 3A (large scale map covering the Siberian craton, Baikal Rift System, Patom range and Aldan craton) and 3B (small scaler centered on the Baikal Rift System) are maps of the crustal thickness from Suvorov et al. (2002) and Polyansky (2002), respectively. Fig. 3C is a map of the lithospheric thickness from Petit and Déverchère (2006) covering the Siberian craton, Baikal Rift System and the Transbaikal area. The outiled shape on the three maps is Lake Baikal. r. stands for river. Fig. 3D is a simplified (from Fig. 2) map of Lake Baikal indicating the three main basins forming the Baikal depression (b. stands for basin).

(Jolivet et al., 2009, 2011) seem to indicate final closure during the early Middle Jurassic, palaeomagnetic data advocate an Early Cretaceous closure (Kravchinsky et al., 2002; Metelkin et al., 2004, 2007, 2010). Furthermore, there is no reported metamorphic event that could be associated with the formation of a collisional range following oceanic closure. If they existed, the sediments derived from erosion of such a range have been exported away from the Transbaikal area.

During the Late Jurassic–Early Cretaceous (possibly contemporaneous with the final closure of the Mongol–Okhotsk Ocean to the east), extensional basins and metamorphic core complexes formed in a vast region covering the southern margin of the Baikal–Vitim terrane (Figs. 1 and 2), the Transbaikal area, southern Mongolia and northern China (e.g. Zheng et al., 1991; Davis et al., 1996, 2001, 2002; Van der beek et al., 1996; Webb et al., 1999; Zorin, 1999; Darby et al., 2001; Fan et al., 2003; Meng, 2003; Wang et al., 2006; Donskaya et al., 2008; Daoudene et al., 2011). Extension was associated with intensive volcanic and intrusive activity (e.g. Tauson et al., 1984; Rutshtein, 1992; Gusev and Khain, 1996; Chen and Chen, 1997; Graham et al., 2001; Daoudene et al., 2011). The mechanism that drove this extension is still largely debated (Jolivet et al., 2009 and references therein; Daoudene et al., 2011). One possible explanation involves the orogenic collapse of the crust thickened by the collision between Siberia and Mongolia–North China. However, as indicated above, there is no evidence of this potential thickening event, including associated metamorphism or evidence of synorogenic sedimentation. Furthermore, the delay between the oceanic closure and the initiation of extension is either extremely short or non-existent, limiting the possibility of a strong thickening event.

Jolivet et al. (2009) suggested that there has been a continuum of deformation between the Mesozoic extension phase and the initiation of the Tertiary extension in the Baikal Rift Zone. For example, extension in the South Baikal depression (the proto South Baikal basin (Fig. 3)), the Barguzin basin and possibly the Tunka basin (Fig. 2) initiated during the Late Cretaceous–Palaeogene (Logatchev et al., 1996; Yarmolyuk and Ivanov, 2000; Mats et al., 2001; Logatchev, 2003; Tsekhovsky and Leonov, 2007; Jolivet et al., 2009).

Contemporaneously with extension to the east, a large planation surface developed in Central Asia during the Mesozoic. Planation was effective from the Early Jurassic around the Gobi Altay–Tian Shan area (De Grave et al., 2007; Jolivet et al., 2007, 2010; Vassallo et al., 2007; Buslov et al., 2008) but only from the Late Jurassic–Early Cretaceous in the East Sayan ranges that form the eastern part of the Sayan ranges, close to Lake Baikal (Jolivet et al., 2011). Furthermore, several authors indicate that complete planation might have been reached only during the Late Cretaceous–Palaeogene in the West Sayan ranges (along the southeastern edge of the West Siberian basin (Fig. 1) (De Grave and Van den haute, 2002; De Grave et al., 2008, 2009, 2011; Glorie et al., 2012). The geodynamic and geomorphic interpretation of this surface is still uncertain. The varying ages proposed for complete planation within the regions where remnants of the surface have been observed indicate that its formation was largely governed by local tectonic activity. In that respect, the Central Asia Planation Surface was probably a low-altitude peneplain resulting from progressive erosion of the various ranges that developed in Central Asia during the Mesozoic. Its altitude could have been around 1000 m, similar to the actual altitude of the Valley of Lakes in Mongolia (Fig. 1). However, in areas such as the Prismorsky range or the Patom range (Fig. 1), the surface is now clearly higher than the surface of the nearby Siberian craton without evidence of a corresponding Tertiary uplift (Jolivet et al., 2009). In this case, the surface could have formed as an elevated plateau built up during the Mesozoic orogeny and preserved from strong erosion since then.

In Siberia, general dismembering of this planation surface initiated in the Late Oligocene–Miocene. To the east and south, extension occurred in the Tunka, South Baikal, Central Baikal and Barguzin basins (Figs. 2 and 3) (e.g. Mats, 1985, 1993; Logatchev, 1993, 2003; Mats et al., 2001; Jolivet et al., 2009). To the west in the Sayan ranges, compression and transpression probably initiated also during the Oligocene (Vdovin,

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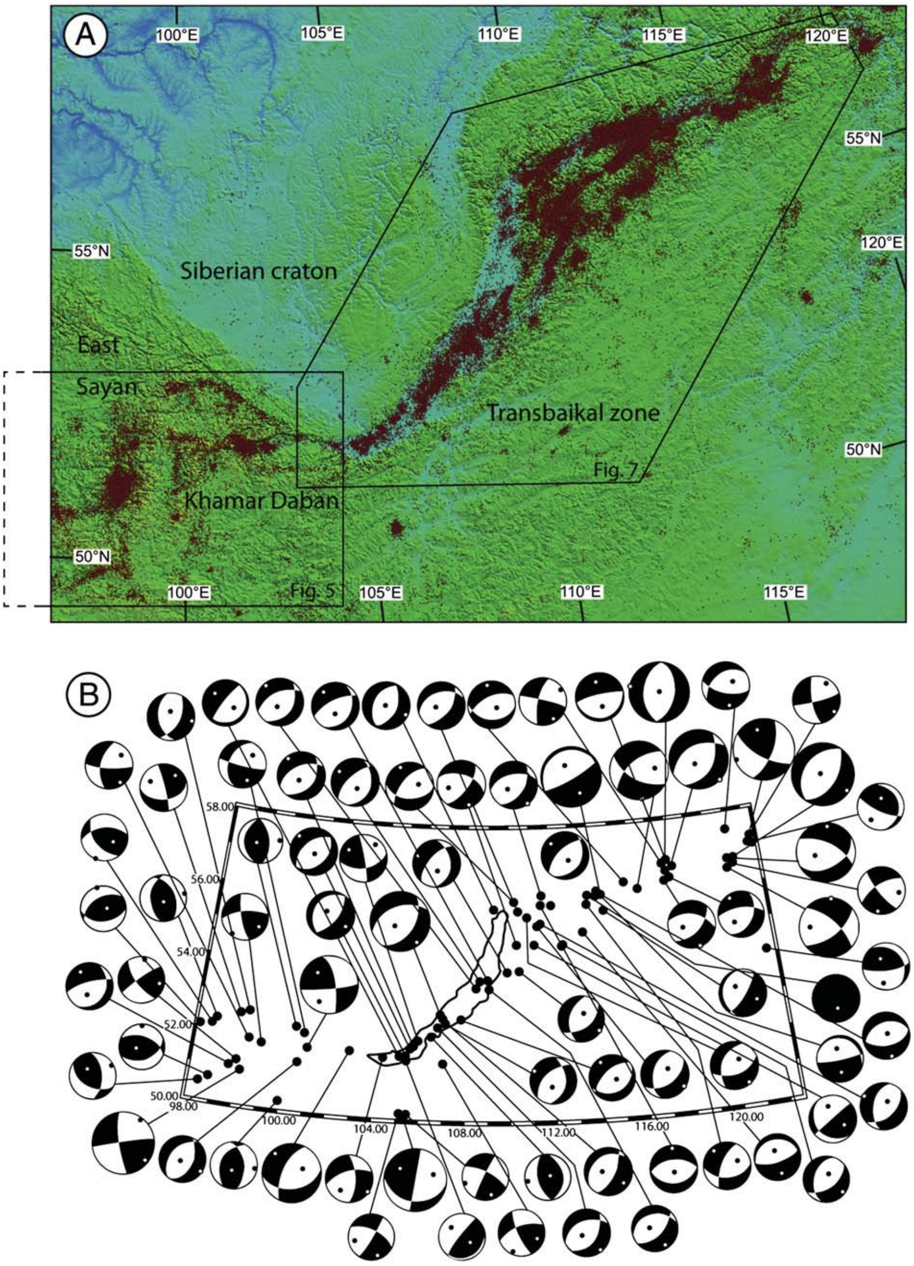

Fig. 4. A: earthquake epicenters in the Baikal rift zone and adjacent areas over the period 1950–2011 based on the data from the Baikal Branch of Geophysical Survey Siberian Branch RAS (http://www.seis-bykl.ru). The earthquake epicenters are not scaled in magnitude. B: focal mechanisms of the M > 5 earthquakes in the BRS and adjacent areas over the period 1950–1998, lower hemisphere projection. Black areas in the diagrams show compression waves and white areas indicate refraction waves. Light and dark points therein correspond to the principal axes of tensile (T) and compressive (P) stresses. The figure was drawn by V.I. Melnikova based on Solonenko et al. (1993) and Melnikova and Radziminovich (1998).

1976), but active relief building only occurred since Late Miocene–Pliocene (Arzhannikova et al., 2011; Jolivet et al., 2011).

3. Geophysical data: modelling the present-day structure and deformation of the crust and lithosphere

The complex tectonic history of southeast Siberia and north Mongolia has produced a highly variable crustal and lithospheric structure. Several estimates of the crustal thickness around the BRS have been obtained based on seismic and gravity data (e.g. Krylov et al., 1981; Zorin et al., 1986, 1989; Suvorov et al., 1999, 2002; Polyansky, 2002; Petit and Déverchère, 2006; Mordvinova and Artemyev, 2010) (Fig. 3). These models differ slightly and it is beyond the scope of this work to discuss them. The mean crustal thickness of the Siberian craton is 40–46 km (Suvorov et al., 2002). Within the Baikal depression, the depth of the Moho varies between 34 km (Central Baikal basin) and 48 km (North Baikal basin) (Fig. 3). All the models agree that a very thick crust (up to 50 km) is present below the East Sayan ranges, the Tunka and Hovsgol basins and the western part of the Khamar Daban ranges (Fig. 2). This sharp change in crustal thickness implies ~5–10 km uplift of the Moho across the Sayan fault that separates the Siberian craton from the Sayan ranges, suggesting that the Sayan fault is a lithospheric-scale tectonic structure (Fig. 3). In the Transbaikal area, the overall crustal thickness appears very heterogeneous but the variations in crustal thickness are not clearly correlated to the various Mesozoic basins.

Numerous models also exist for the structure of the lithosphere, based on tomography, seismic and gravity data (e.g. Petit and Déverchère, 2006 and references therein; Kulakov, 2008; Kulakov and Bushenkova, 2010; Mordvinova and Artemyev, 2010). Lithospheric

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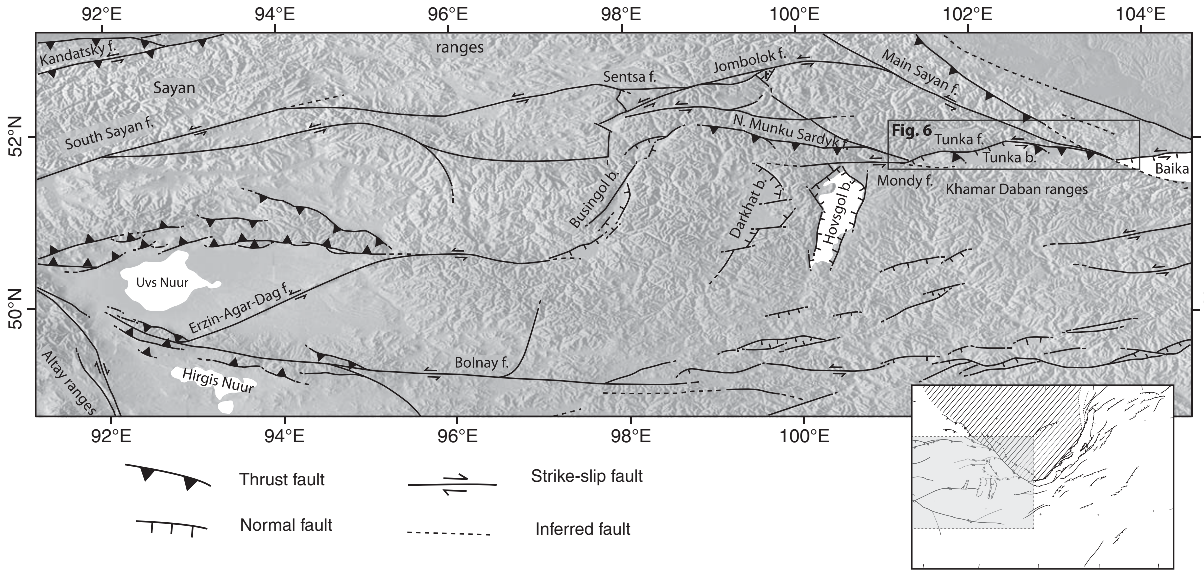

Fig. 5. General tectonic map of northwestern Mongolia and southern Siberia showing the Busingol–Darkhat–Hovsgol–Tunka (BDHT) and Bolnay strike–slip fault systems. b. stands for basin and f. for fault. The small inset map (simplified from Fig. 2) shows the area covered by Fig. 5 in the study area.

thickness decreases progressively from 180 km under the Siberian craton to ~60–70 km below the Vitim–Transbaikal and Khamar Daban areas (Fig. 1), apparently in conjunction with the occurrence of large Cretaceous–Tertiary volcanic fields (Figs. 2 and 3) (see also Lebedev et al., 2006; Kulakov, 2008; Ivanov and Demonterova, 2010).

Seismic activity within the studied area is strongly correlated with the BRS (Fig. 4). However, some intense activity is also recorded in the East Sayan ranges along the Tunka–Mondy faults, in the Darkhat basin and along the Jombolok fault (Figs. 2, 4 and 5). Towards the north of the BRS, the seismic activity is spread on a wide zone encompassing the Barguzin and North Baikal basins (Fig. 4). In the Central Baikal and Barguzin basins, focal mechanisms indicate NW–SE extension with almost no strike–slip movement. The North and South Baikal basins show both extension and a small left-lateral strike–slip displacement. Finally, the Transbaikal region and NW Sayan ranges are virtually free of seismic activity, except for some localised spots potentially associated with volcanic activity.

Global Positioning System (GPS) data recorded both locally within Mongolia and SE Siberia and more regionally within Asia provide valuable information on the present day crustal deformation (e.g. Calais et al., 1998, 2003, 2006; Polyansky, 2002; San'kov et al., 2003, 2004; Petit and Fournier, 2005; Vergnolle et al., 2007). Fig. 2 shows displacement vectors from the various GPS stations in the studied area with respect to stable Eurasia. To the SW in Mongolia, the general motion is directed ENE and is accommodated by large E–W-directed sinistral strike–slip faults such as the Bolnay fault (Fig. 2). The few data available in the West Sayan ranges indicate movements towards the NNE, compatible with the observed compressional deformation in this area. The general crustal motion turns towards the NE in the Khamar Daban region, south of the Tunka basin. Further east, GPS data indicate a sharp turn from a NNE to SE direction. This rotation corresponds to the switch between the mainly transpressive deformation in the East Sayan ranges to the NW–SE extension in the Transbaikal area.

Based on geophysical and tectonic data such as fault plane measurements, satellite imaging of the main structures and stress-field components derived from microtectonics measurements, several numerical and analogue models have been proposed to explain the present-day deformation of Asia and the BRS. Most of these models acknowledge that if continued convergence between India and Asia is the main driving force for crustal deformation throughout Asia, the dynamics of the subduction zones that border the eastern edge of Asia must play a significant role in the observed deformation patterns (e.g. Calais et al., 2003, 2006; Vergnolle et al., 2007). Buoyancy forces may also explain part of the deformation (e.g. Flesch et al., 2001;

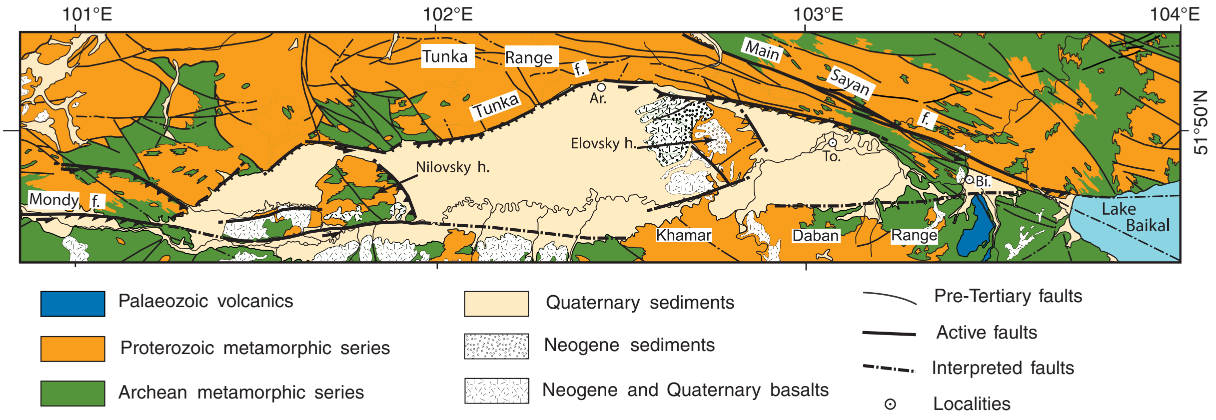

Fig. 6. Tectonic and geological map of the Tunka basin. The geology is based on the Russian geological map (VSEGEI, 1975). Ar., Arshan village; To., Tory village; Bi., Bistraya village. Elovsky h. and Nilovsky h. are both basement highs (h.). f. stands for fault.

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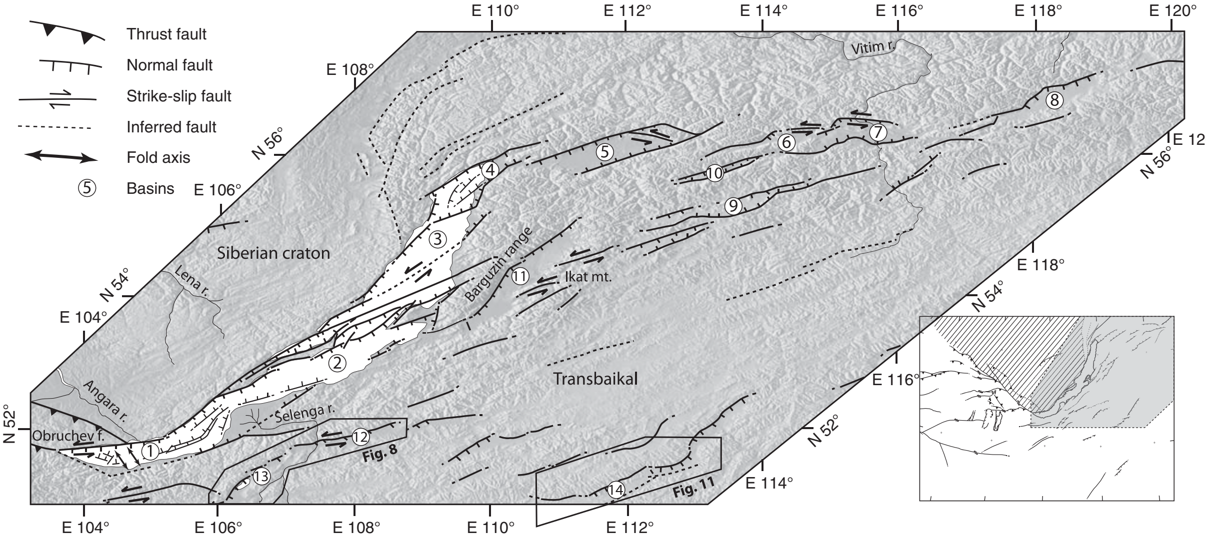

Fig. 7. General tectonic map of the Baikal Rift System and the Transbaikal region showing the main active faults. Numbers indicate the different active basins: 1. South Baikal; 2. Central Baikal; 3. North Baikal; 4. Kichera; 5. Upper Angara; 6. Muyakan; 7. Muya; 8. Chara; 9. Tsipa-Baunt; 10. Upper Muya; 11. Barguzin; 12. Ulan-Ude; 13. Gusinoozersky; 14. Ingodinskaya. f. stands for fault, mt. for mountains and r. for river. The two boxes indicate the location of Figs. 8 and 12. The inset map (simplified from Fig. 2) shows the general location (grey shaded area) of Fig. 7 within the study area (see also box in Fig. 4). The dashed area in the inset indicates the Siberian craton.

England and Molnar, 2005; Calais et al., 2006), but only in areas of high relief such as the Tibetan plateau and not in our region of interest (Vergnolle et al., 2007).

Using numerical models, Petit and Fournier (2005) demonstrated that the present-day stress and velocity fields in the Amuria plate could be explained by the conjunction of NE-directed compression (due to the India–Asia collision) and SE-directed extrusion (driven by Pacific subduction to the east). These authors showed that the Amuria plate mostly behaves as a rigid block extruded towards the SE. Polyansky (2002) further demonstrated that the GPS and fault

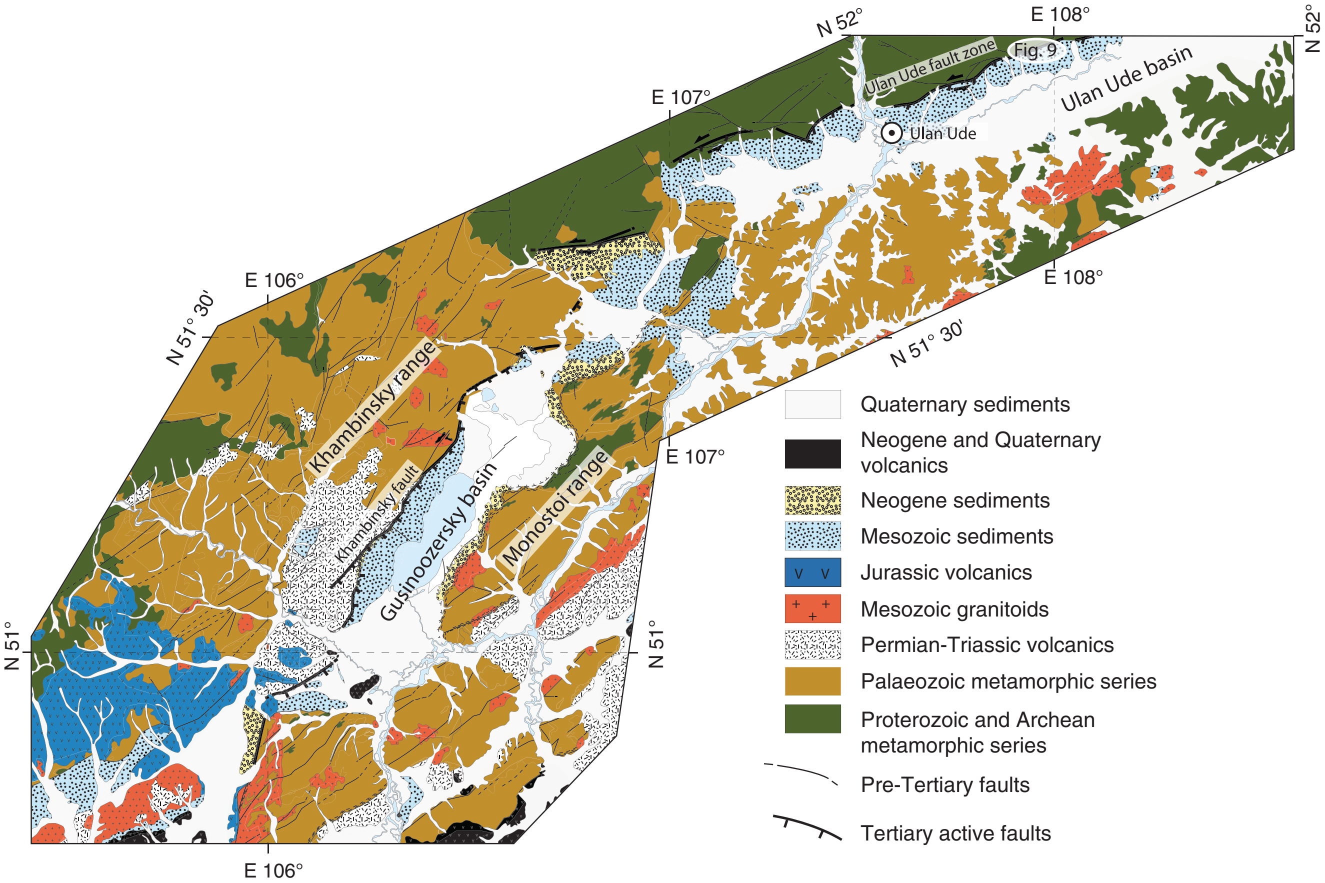

Fig. 8. Simplified geological map of the Gusinoozersky and Ulan Ude basins showing the main active faults (thick lines). Modified from the 1:200,000 West-Transbaikal geological maps (VSEGEI, 1958, 1960, 1961a,b, 1962, 1964). The white ellipse labelled Fig. 9 near Ulan Ude indicates the approximate location of Fig. 9.

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kinematics data around the BRS are in good agreement with a model associating NE-directed compression between both tips of the Baikal rift with a SE-directed extension applied to the whole Transbaikal region. In particular, this model seems to explain the switch between transpression and extension and the sharp bend in the direction of the GPS-derived velocity field south of the BRS.

Chemenda et al. (2002) challenge this explanation of the formation and evolution of the BRS using a lithospheric-scale analogue model. Their model combines the effect of the shape of the Siberian craton with an initial weakness zone in the lithosphere along the edge of the craton associated with E–W-directed extension. Similarly, using an analogous setting involving a pre-existing curved (in map view) structural lineament running along the long axis of the BRS, Seminsky (2009) describes the evolution of the BRS through a pure left-lateral strike–slip zone. However, this 2D model does not take into account the tectonic deformation in the Transbaikal area, nor in the Sayan ranges. For example, the uplift of the Sayan ranges is considered as passive, driven by an anomalous mantle acting below the ranges. However, as for several other proposed models, a significant conclusion of Seminsky (2009) is that the deformation that affected the BRS is older than the India–Asia collision and thus cannot be entirely driven by this mechanism. The forces that drove the initial extension in the BRS are thus different from those that drive it in the present day, and the question remains whether they are still active and superimposed on the India–Asia collisional stress field.

4. The Busingol–Darkhat–Hovsgol–Tunka system

The Busingol–Darkhat–Hovsgol–Tunka (BDHT) system is a major strike–slip fault network with extensional relay zones, connecting the Erzin–Agar Dag strike–slip fault to the west (Arzhannikov and Arzhannikova, 2009) with the Mondy and Sayan faults to the east (Larroque et al., 2001; Chipizubov et al., 2003; Arjannikova et al., 2004; Arzhannikova et al., 2005) (Figs. 1 and 5). Little is known about the initiation and tectonic evolution of the N–S-oriented Busingol, Darkhat and Hovsgol basins, which have mainly been investigated for palaeoclimate studies (e.g. Hövsgöl Drilling Project Group, 2007). However, subsidence in the Hovsgol basin initiated during the Late Miocene around 8–7 Ma (Ivanov and Demonterova, 2009) and probably later during the Pliocene in the Darkhat basin (Ufland et al., 1969; Spirkin, 1970). All three basins are E–W extending structures bordered by segmented normal faults (Fig. 5). In the Darkhat basin, the normal faults seem restricted to the eastern side of the basin, while in the Busingol and Hovsgol basins, both sides are affected. The Busingol basin is connected to the north with the North Munku–Sardyk reverse fault system by a series of N–S sinistral strike–slip faults and N–S to NE–SW transtensive faults (Fig. 5).

Towards the east, the ~80 km long Mondy fault connects the Hovsgol system with the Tunka basin (e.g. Treskov and Florensov, 1952; Arjannikova et al., 2004). The Mondy fault is mainly a left-lateral strike–slip structure with a small reverse component on some of its segments. Reverse movements are also observed on the Ikhe–Ukghun fault that connects with the Mondy fault in its eastern termination (Fig. 5). The present-day kinematic regime on the Mondy fault has prevailled since at least the Pleistocene and replaced the transtensional regime that was active since the Oligocene (Arjannikova et al., 2004).

The Tunka basin develops along an E–W axis parallel to the general direction of the BDHT strike–slip system (Figs. 5 and 6). The older sediments recognised within the Tunka basin are Late Cretaceous–Palaeogene (Mats, 1993; Scholz and Hutchinson, 2000) and probably correspond to an initial limited phase of basin development through the reactivation of inherited faults, contemporaneous with the initial formation of the South Baikal depression (Tsekhovsky and Leonov, 2007) and the Barguzin basin (Figs. 2 and 7) (Jolivet et al., 2009). However, the Late Cretaceous–Palaeogene sediments are covered by a thick lateritic–kaolinic weathering crust corresponding to a phase of non-sedimentation and thus probably of tectonic quiescence (Mazilov et al., 1972; Mats, 1993; Kashik and Masilov, 1994; Dehandschutter et al., 2002; Logatchev et al., 2002; Jolivet et al., 2011). Strong subsidence of the basin initiated in the Late Oligocene–Early Pliocene (Logatchev and Zorin, 1987; Mazilov et al., 1993) and the present general contours of the basin were probably established by the Late Pliocene (Mazilov et al., 1972). Finally, inversion of the basin initiated during the Late Pliocene–Early Pleistocene (Larroque et al., 2001; Rasskazov et al., 2010; Arzhannikova et al., 2011) contemporaneously with the inversion observed along the Mondy fault. This inversion was most likely controlled by a change in the regional stress field due to significant strengthening of India–Asia compression that has been affecting the region since the Late Miocene (e.g. Arzhannikova et al., 2011; Jolivet et al., 2011).

From Mondy to the west to Bistraya in the east, the Tunka basin is formed by a series of sub-basins separated by basement highs (Fig. 6). Two dominant ridges, the Nilovsky high to the west and the Elovsky high to the east, are uplifted along N130°-trending, SW-verging thrust faults bordering their eastern edges (Fig. 6). North of the basin, the Tunka range presents an alpine-style dissected topography with summits up to 3300 m. The range is separated from the basin by the complex Tunka fault system that developed mostly through reactivation of Proterozoic and Palaeozoic inherited structures (e.g. Sherman et al., 1973; McCalpin and Khromovskikh, 1995; Larroque et al., 2001; Arzhannikova et al., 2011; Zhimulev et al., 2011). In the western part of the basin, the Tunka fault is composed of a series of E–W sinistral strike–slip fault segments relayed by NE–SW normal fault segments (e.g. Larroque et al., 2001). Further to the east, the Tunka fault turns towards E–SE and merges with the lithospheric-scale Sayan fault system. In that region, between Arshan and Tory, the Tunka fault is formed by left-lateral reverse segments roughly parallel to the thrust faults bordering the eastern edge of the Nilovsky and Elovsky basement highs in the basin (Larroque et al., 2001; Chipizubov et al., 2003).

The Tunka basin is bordered to the south by the 2200–2500 m high flat-topped Khamar Daban range. Several authors indicate the existence of a sinistral strike–slip fault separating the basin from the range, reported to have initiated during the Pliocene (Delvaux et al., 1997; Lunina and Gladkov, 2004a,b). However, field evidence for a continuous structure along the southern margin of the Tunka basin is lacking (e.g. Arjannikova et al., 2004), except towards the west where a fault segment merges with the Mondy fault.

The Tunka basin can thus be interpreted as a pull-apart structure (Sherman and Levi, 1978) along a major E–W strike–slip system linking the Erzin–Agar Dag strike–slip fault with the southern tip of Lake Baikal. The basin opened as a series of sub-basins separated by compressional ridges. The present tectonic regime is clearly linked to the NW-directed motion of the Mongolian lithosphere induced by the India–Asia collision. The Late Pliocene–Pleistocene inversion event recorded in the Tunka fault system could correspond to the final connection between the western part of the BDHT system and the Mondy strike–slip fault. However, the Oligocene onset of slow subsidence in the Tunka basin, as well as the Late Miocene and Pliocene subsidence in the Hovsgol and Darkhat basins, remains to be explained. The Late Miocene–Pliocene initiation of the Hovsgol and Dakhat basins is contemporaneous with the first evidence of relief building in the East Sayan ranges (e.g. Arzhannikova et al., 2011), and both mechanisms could be linked to the initial poorly-expressed effects of the India–Asia collisional stress field. As suggested by Tsekhovsky and Leonov, 2007 and Jolivet et al., 2009 the Oligocene subsidence of the Tunka basin could be linked to a second (or effectively first) stress field generated by far-field effects of the deformation along the Pacific subduction zone, or alternatively by thermal perturbations in the mantle due to the occurrence of subducted mantle slabs (the Pacific slab or the Mongol–Okhotsk slab). By Late Pliocene–Pleistocene times, both

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stress fields would have been superimposed, the India–Asia collision mechanism inducing most of the tectonic deformation.

5. The Baikal Rift System

5.1. The Baikal basins

The BRS (Figs. 1 and 7) is a key feature of the tectonic evolution of Asia and has been studied by a number of authors. A complete summary of the formation of the BRS is given in Petit and Déverchère (2006) and Jolivet et al. (2009) and only the main points will be discussed below. Two main hypotheses are proposed for the opening of the BRS (Sengör and Burke, 1978): (1) the “active rift hypothesis” considers that rifting is driven by a wide asthenospheric diapir acting on the base of the crust beneath the rift axis (e.g. Logatchev and Zorin, 1987; Windley and Allen, 1993; Kulakov, 2008); (2) the “passive rift hypothesis” considers that the BRS is a pull-apart basin opening in response to eastward extrusion of the Amuria plate, driven by the India–Asia collision to the south (e.g. Molnar and Tapponnier, 1975; Cobbold and Davy, 1988; Petit et al., 1996; Petit and Fournier, 2005; Petit and Déverchère, 2006). Recent geophysical data tend to demonstrate that there is no hot mantle plume beneath the rift axis and that the “active rift hypothesis” is probably not valid (e.g. Tiberi et al., 2003; Ivanov, 2004; Petit et al., 2008). However, it also appears that if the “passive rift hypothesis” is favoured, the occurrence of a strong structural inheritance within the crust and lithosphere is required (e.g. Delvaux et al., 1997; Chemenda et al., 2002; Jolivet et al., 2009). Finally, several studies have explored the possibility of interactions between the compressional stress field generated by the India–Asia collision and an extensional stress field produced by the western Pacific subduction (e.g. Kimura and Tamaki, 1986; Davy and Cobbold, 1988; Fournier et al., 1994, 2004; Logatchev, 2003).

The chronology of BRS opening is mostly based on sedimentological data, and is generally divided in two phases: (1) a “slow rifting” phase during the Late Oligocene to the Late Pliocene, followed by (2) a “fast rifting” phase from the Late Pliocene to the present (e.g. Logatchev and Zorin, 1987; Logatchev, 1993, 2003; Mats et al., 2001; Hus et al., 2006; Petit and Déverchère, 2006). However, as explained above, recent apatite fission track studies (Jolivet et al., 2009) and sedimentological data (e.g. Mats, 1993; Scholz and Hutchinson, 2000; Tsekhovsky and Leonov, 2007) suggest that extension in the South Baikal basin initiated in the Late Cretaceous–Palaeogene. During the Early Pliocene, the BRS propagated towards the north and the North Baikal basin started to form in association with renewed tectonic activity within the Barguzin basin (e.g. Hutchinson et al., 1992; Delvaux et al., 1997; Petit and Déverchère, 2006; Jolivet et al., 2009; See below). This northward propagation of the BRS is contemporaneous with a general increase in deformation rates throughout the whole of Central Asia (e.g. De Grave et al., 2007; Vassallo et al., 2007; Jolivet et al., 2011).

A large majority of the faults controlling the three main Baikal basins have a purely normal component (e.g. Delvaux et al., 1997; Levi et al., 1997; Petit and Déverchère, 2006). However, strike–slip components and compressional folding are also observed (Fig. 7). In the South Baikal basin, the N80°-trending Obruchev fault, which is directly connected to the Main Sayan fault and limits the basin to the west, is a left-lateral transtensive structure (e.g. Radziminovitch et al., 2006). The orientation of the Obruchev fault is similar to the orientation of the E–W left-lateral strike–slip fault segments along the northern edge of the Tunka basin. Seismic profiles within the southern tip of the South Baikal basin also show some 1 km amplitude and 7–8 km wavelength folds with NW-trending axes. These folds are covered by non-deformed Middle Pleistocene–Holocene sediments (Levi et al., 1997). Similar folding has been reported by Voropinov (1961) in Miocene coal-bearing series of the southern margin of the South Baikal basin. Sherman and Levi (1978) and Levi et al. (1997) interpret the folding as resulting either from short-term tectonic inversion through NE–SW directed compression, or from sinistral strike–slip movements in the basement. Towards the north of the South Baikal basin, seismic profiles also show evidence of lateral movements on an axial basement fault (Levi et al., 1997) that may represent the northeastward continuation of the Obruchev fault. The Central Baikal basin is again affected by transtensive faults trending towards N60°, transverse to the axis of the basin (Levi et al., 1997). Some of these faults affect the southern reach of the Barguzin basin (Fig. 7).

5.2. The North Baikal basins system

NE of the North Baikal basin, the BRS continues through a series of NE–SW-oriented, en echelon and generally asymmetric basins, namely the Kichera, Upper Angara, Muyakan, Upper Muya, Muya, Tsipa-Baunt and Chara basins (Fig. 7). This large-scale structure has been interpreted either as a major shear zone (or transform fault) linking the BRS with the Stanovoy strike–slip zone to the east (Sherman, 1992), or as a zone of oblique extension (San'kov et al., 2000). The faults that define the basin margins are strongly controlled by inherited crustal structures (San'kov et al., 2000; Petit and Déverchère, 2006). West of 111°E, the N20°-trending normal fault that bounds the western side of Lake Baikal is nearly parallel to the edge of the Siberian craton (Logatchev and Zorin, 1992; Logatchev, 1993; San'kov et al., 2000). Between 111°E and 115°E, the various fault segments that bound and link the basins form an en echelon structure corresponding to a general E–W-directed left-lateral strike–slip fault zone (San'kov et al., 2000). Within this array, the main faults trend towards N60°, parallel to the general inherited structural direction of the basement (San'kov et al., 2000; Lunina and Gladkov, 2008). Finally, east of 115°E the faults have a more dispersed orientation ranging from E–W to N–S.

The exact timing of formation of the North Baikal basins is still debated, but ranges between the Late Miocene and the Pliocene depending on the location of the individual basin (e.g. Hutchinson et al., 1992; Logatchev and Zorin, 1992; San'kov et al., 2000; Petit et al., 2009). Their initiation is thus very similar in age to the BDHT basins, and the contemporaneous opening of these basin systems at both ends of the BRS might correspond to a major development phase of the rift. However, within the North Baikal basins, no clear eastward propagation of deformation is observed that would correspond to an eastward propagation of the BRS. On the contrary, San'kov et al. (2000) report a largely homogeneous displacement field (in terms of magnitude) over the whole rift in both space and time.

5.3. The Barguzin basin

Unlike the other basins in the east of the Transbaikal area, the Barguzin basin is connected to the Central Baikal basin, and its evolution is closely related to the evolution of the BRS (Fig. 7). With a surface ~500 m high, it is separated from the North Baikal basin by the ~2600 m high Barguzin range. It is bounded to the NW by a series of NE–SW-directed en echelon active normal faults (Fig. 7) (e.g. Florensov, 1960; Solonenko, 1968; Delvaux et al., 1997; Lunina and Gladkov, 2007). While initial studies reported strike–slip motion on these faults (Solonenko, 1968, 1981), recent palaeoseismological and geomorphological investigations of the fault scarps only reported normal motion (Chipizubov et al., 2007).

To the east, the Barguzin basin is separated from the Ikat mountains by NE–SW-striking normal faults that seem to have only limited offset (Fig. 7). A set of E–W left-lateral strike–slip faults is also observed mainly in the Ikat mountains, some of which reach the basin (Lunina and Gladkov, 2007). Like most of the Transbaikal basins, the infill is asymmetric, the greatest thickness (2500 m) being observed along the Barguzin range to the west (Fig. 7) (e.g. Nevedrova and Epov, 2003; Epov et al., 2007). Furthermore, like the Tunka basin, the Barguzin basin is divided into three sub-basins by basement highs

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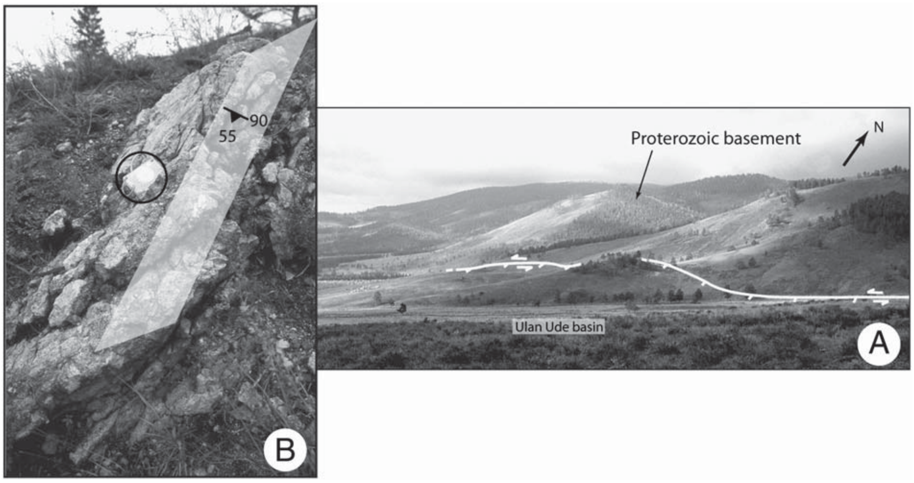

Fig. 9. A: General view of the active northern margin of the Ulan Ude basin, highlighting the position of the left-lateral transtensive Ulan Ude fault zone. B: picture showing the inherited ductile fault planes in the Palaeozoic metamorphic series. Scale is provided by the compass in the black circle.

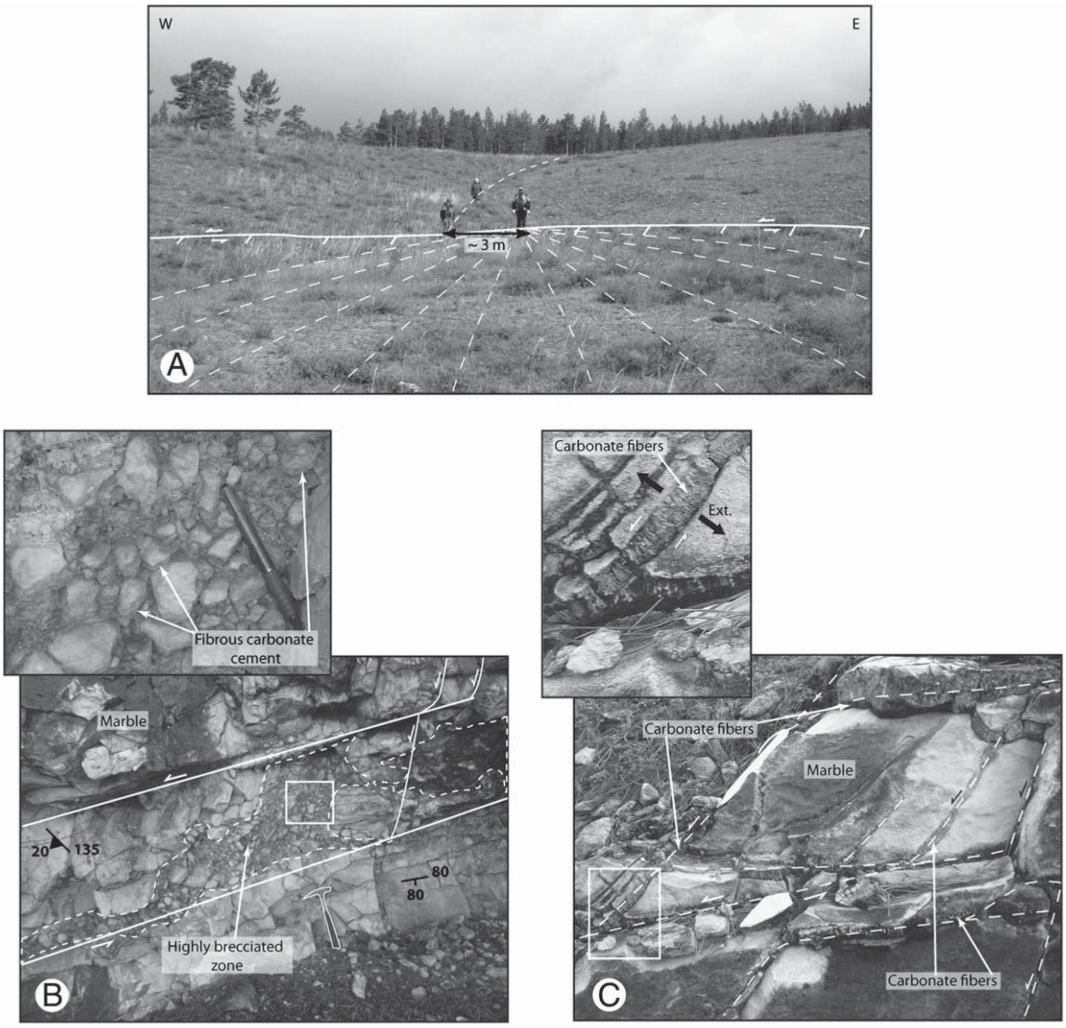

Fig. 10. Northern margin of the Ulan Ude basin. Fig. 10A shows an offset gully and its corresponding alluvial fan. The total displacement between the apex of the fan and the gully is estimated at ~3 m (black arrow). Fig. 10B shows left-lateral transtensive fractures and associated breccias in the Proterozoic marbles outcropping along the path of the active fault. Fig. 10C shows transtensive fractures filled with carbonate fibers developing in the same Proterozoic marbles. This deformation clearly occurred in more superficial levels than the ductile fault planes observed in Fig. 9. The fault direction is compatible with the direction of the active fault as given by the fault scarp on Fig. 10B. There is no age constraint on the structures observed in the marbles, and they may be either Mesozoic (during the initial formation of the basin) or Tertiary (corresponding to the actual phase).

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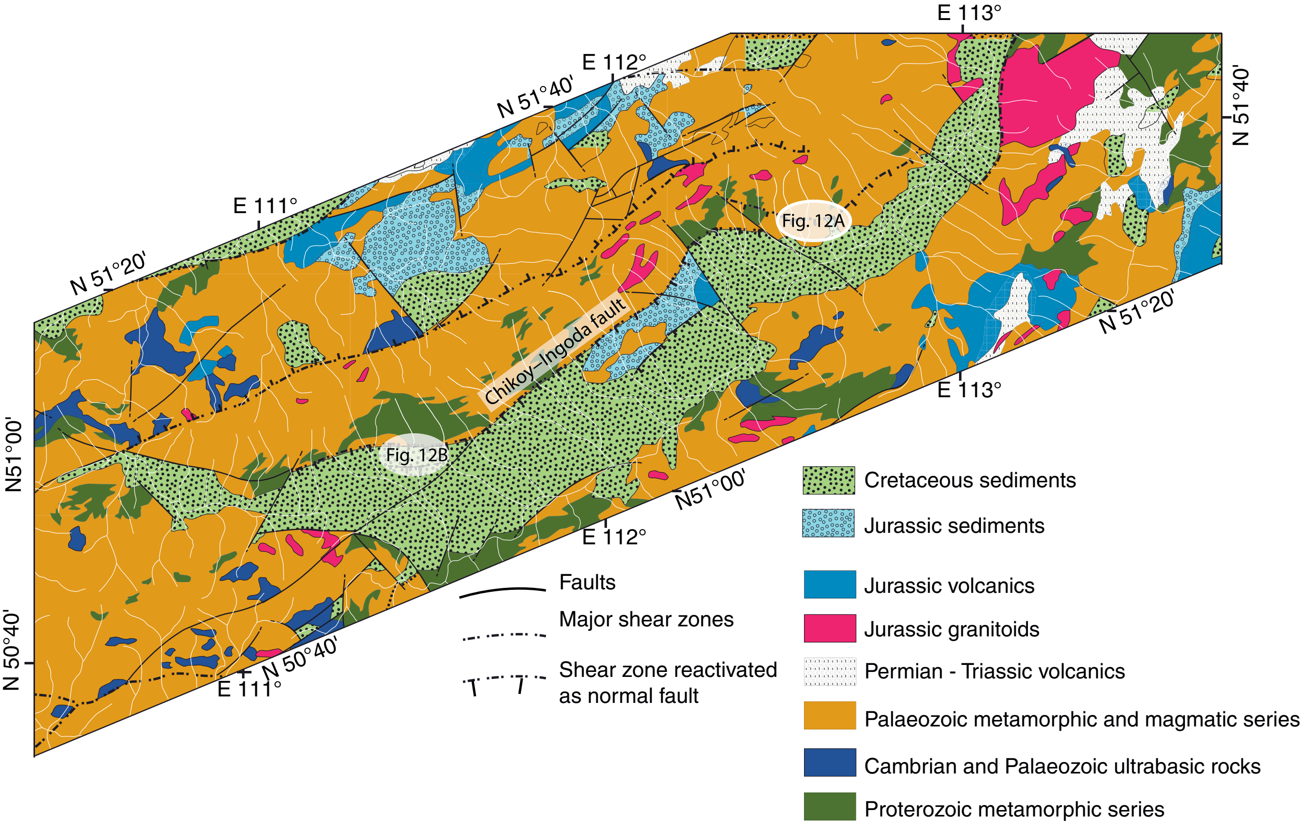

Fig. 11. Simplified geological map of the Ingodinskaya basin highlighting the main active faults. These faults are superimposed on major Proterozoic or Palaeozoic shear zones. Note the large amount of Jurassic magmatism characteristic of the Transbaikal basins. The white ellipses indicate the area of Fig. 12A and B. Modified from the 1:1,000,000 Russian geological map (VSEGEI, 1977).

separated by NE-striking faults (Florensov, 1977; Epov et al., 2007). Based on sedimentological data, the evolution of the Barguzin basin initiated during the Middle Pliocene, contemporaneous with the onset of the "fast rifting" stage of BRS development (Florensov, 1982). However, although confirming a Pliocene increase in exhumation of the Barguzin range, recent apatite fission track work demonstrated that this exhumation initiated during the Late Cretaceous or the Early Palaeogene (Jolivet et al., 2009). This timing is compatible with both the Late Cretaceous–Palaeogene subsidence observed in the Tunka basin (Mats, 1993; Scholz and Hutchinson, 2000) and the initiation of the South Baikal basin (Tsekhovsky and Leonov, 2007) resulting from a geodynamic mechanism independent of the India–Asia collision (Jolivet et al., 2009 and references therein).

6. The Transbaikal extension zone

6.1. The Gusinoozersky and Ulan Ude basins

The NNE–SSW-elongated Gusinoozersky basin (Figs. 7 and 8) is an asymmetrical Mesozoic graben predominantly filled by Late Jurassic to Early Cretaceous detrital sediments (conglomerates and sandstones interbedded with coal layers), and covered by a thin layer of Neogene to Quaternary river sediments (VSEGEI, 1958, 1960, 1961a,b, 1962, 1964; Bulnaev, 2006; Tsekhovsky and Leonov, 2007). The Mesozoic sequence is much thicker (up to 2500 m) along the Monostoi normal fault that bounds the basin to the east than along the Khambinsky normal fault to the west, implying that the eastern edge of the basin was more active during the Mesozoic (Bulnaev, 2006; Lunina and Gladkov, 2009). However, our field investigations show that the Tertiary reactivation only occurred along the Khambinsky fault, which shows a well defined fault scarp and is associated with at least two seismic rupture events at 5290 ± 100 years and 2680 ± 60 years BP (Chipizubov et al., 2002). The active Khambinsky fault appears divided into several segments (Fig. 8) whose location is probably governed by structural inheritance. Movement on the Khambinsky fault is interpreted as purely normal, because no evidence for strike–slip components have been observed based on satellite images, field investigations or previously published work (Chipizubov et al., 2002). Several other normal faults affect the basin but do not show any evidence of Quaternary ruptures.

To the south, the Gusinoozersky basin is linked to the major Bolnay strike–slip fault system (Figs. 2 and 5) through the Uda–Vitim Fault zone (e.g. Delvaux et al., 1997) (Fig. 2). The left-lateral strike–slip movement on the Bolnay fault is transferred in the Transbaikal region through a series of small en echelon pull-apart structures linked by strike–slip segments (Fig. 8). The fault system progressively curves from the E–W direction of the Bolnay fault to an ENE–WSW-directed left-lateral transtensional fault zone south of the BDHT system, before reaching the NNE–SSW-directed purely extensional Gusinoozersky basin.

North of the Gusinoozersky basin, the ENE–WSW-elongated Ulan Ude basin (Figs. 7 and 8) is a Mesozoic half-graben also filled with Late Jurassic–Early Cretaceous detrital sediments. It is separated from the Gusinoozersky basin by a basement high composed of Palaeozoic and Proterozoic metamorphic rocks (Fig. 8). The basin is bounded to the north by a series of E–W to ENE–WSW trending en echelon normal faults (the Ulan Ude fault zone). Like the margins of the Gusinoozersky basin, the Ulan-Ude fault zone is largely controlled by an inherited ductile shear zone that developed in the basement probably during the Palaeozoic (Fig. 9). Fault planes were measured as striking N70–90°, with dips of 55–75° S. To the south, no fault can

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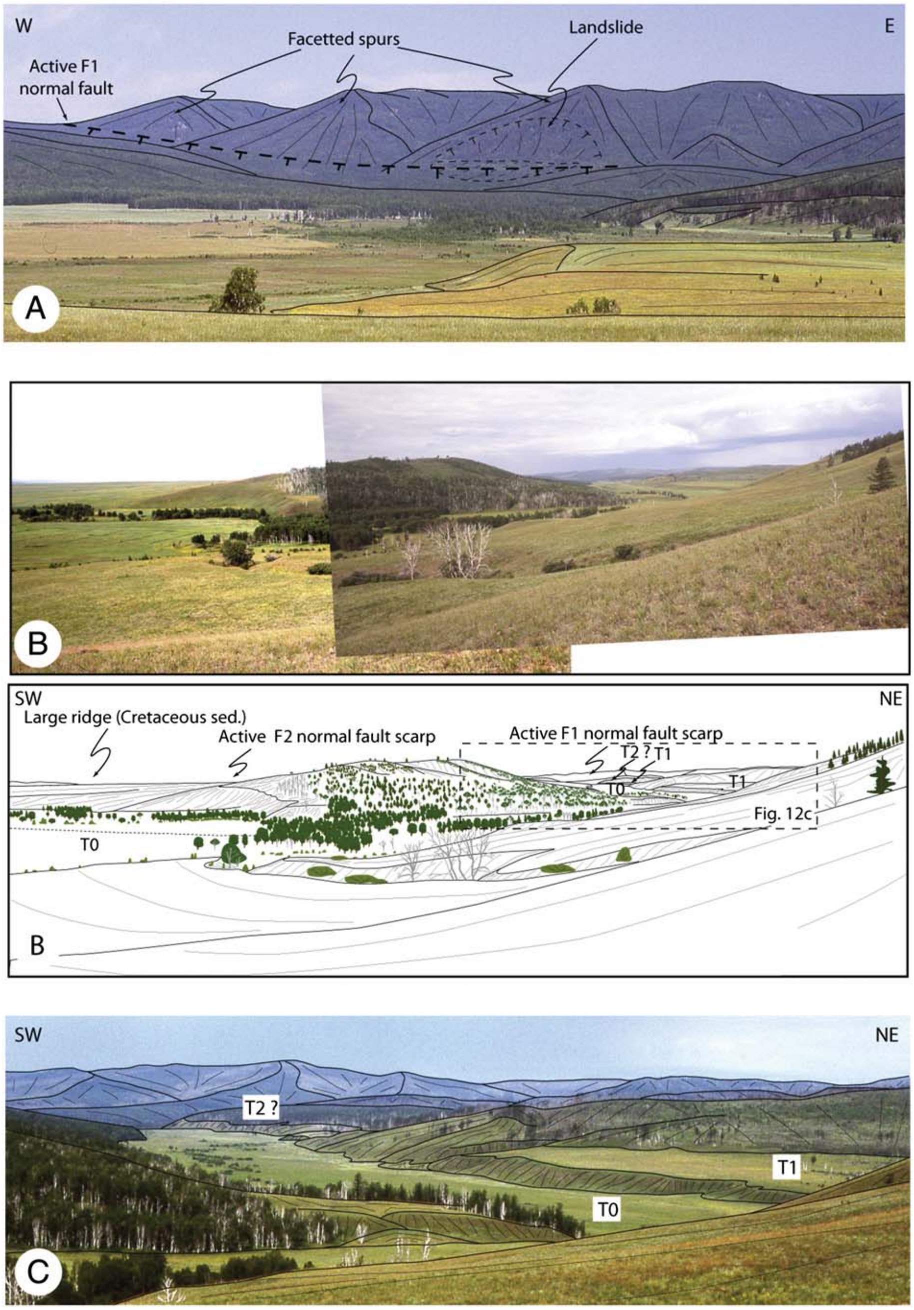

Fig. 12. Features of the active northern edge of the Ingodinskaya basin. Fig. 12A shows an overview of the normal fault scarp bordering the basin. Well-developed facetted spurs, some of them affected by landslides, indicate recent or active normal faulting. Fig. 12B shows a section of the basin situated between two parallel normal faults (F1 and F2) and displaying a series of alluvial terraces detailed in Fig. 12C. Fig. 12C is a close view of the alluvial terraces developed between faults F1 and F2. The terraces T0 and T1 are well recognised, and a third terrace (T2?) might exist in the background. However, we were not able to clearly distinguish between this possible T2 level and the well-developed T1 terrace. Both T2(?) and T1 disappear south of the F2 fault, implying recent activity along this structure.

be observed, suggesting that all the Mesozoic deformation occurred on the northern side of the basin. The Ulan-Ude fault zone also shows evidence of Quaternary deformation: gullies are offset by several meters (Fig. 10A), and fault planes developed in the Proterozoic basement marbles indicate a strongly left-lateral transtensional movement on most of the fault segments (Figs. 10B and C). However, no direct age constraints are available.

Like the Tunka system, the Gusinoozersky and Ulan Ude basins can be interpreted as pull-apart structures along a major strike–slip system. This second system initiates to the west with the Bolnay fault and progressively turns towards the north when reaching the Transbaikal area.

6.2. The Ingodinskaya basin

The NE–SW-elongated Ingodinskaya basin (Figs. 7 and 11) is located 400 km east of Lake Baikal, near the city of Chita. The basin is again a Mesozoic graben filled by Late Jurassic to Early Cretaceous detrital sediments (VSEGEI, 1977). Both the northern and southern margins of the basin are faulted. Like the Gusinoozersky and Ulan Ude basins, the Mesozoic faults are superimposed on an inherited Palaeozoic ductile shear zone, forming the Chikoy–Ingoda fault zone (Zorin, 1999). The bordering faults are segmented, with two sets of en echelon E–W and NE–SW-oriented segments. From relationships evident in geological maps and additional field investigations, the

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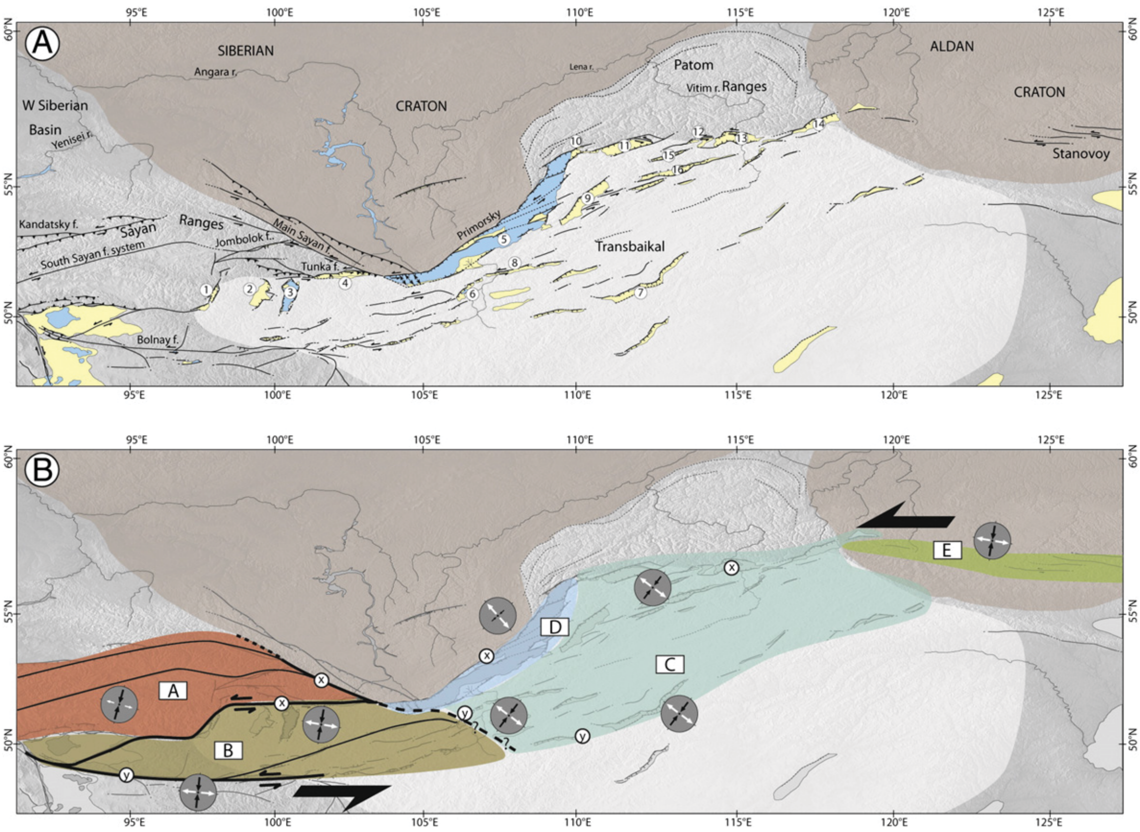

Fig. 13. General model for deformation in the Baikal Rift System (BRS). A: general tectonic map of the studied area, showing the main active faults and Tertiary basins (in yellow). Numbers indicate the major basins: 1. Bunsingol; 2. Darkhat; 3. Hovsgol; 4. Tunka; 5. Baikal; 6. Gusinoozersky; 7. Ingodinskaya; 8. Ulan-Ude; 9. Barguzin; 10. Kichera; 11. Upper Angara; 12. Muyakan; 13. Muya; 14. Chara; 15. Upper Muya; and 16. Tsipa-Baunt. The general limits of the Siberian and Aldan cratons are shaded in brown. The Transbaikal crust is shaded in grey. B: tectonic model for the formation of the BRS and Tertiary Transbaikal basins: the deformation is driven by two major strike–slip systems: (x) the BDHT–Baikal–Stanovoy system, and (y) the Bolnay system. Left-lateral movement along both fault zones generate different local stress fields as indicated by the stress symbol. White arrows correspond to extensional stress and black arrows to compressional stress (e.g. Delvaux et al., 1997). In the Sayan ranges (Zone A in red), NNE–SSW transpression dominates, building positive topography along large transpressive thrust faults. In the BDHT area (Zone B in brown), E–W transtension dominates, leading to the formation of the various BDHT basins. In the Transbaikal crust (Zone C in green), NE–SW transtension dominates, with a major NW–SE extensional component developed. This diffuse deformation is driven by the Bolnay strike–slip system associated with the strong tectonic inheritance of the Transbaikal crust. Around Lake Baikal (Zone D in blue), NE–SW extension dominates, mainly localised along the edge of the Siberian craton. Finally, the BDHT–Baikal–Stanovoy system ends in the Stanovoy strike–slip fault system along the southern limit of the Aldan craton (Zone E in green). See text for detailed discussion. (For references to colour in this figure legend, the reader is referred to the web version of this article.)

northern fault zone appears active, whereas no evidence of activity is found along the southern fault zone. The former features a distinct fault scarp that displays a series of well-developed facetted spurs indicative of recent normal faulting (Fig. 12A). Recent erosion of those spurs by landslides might also indicate active deformation. Rock exposures in the field are extremely restricted due to the thick forest cover, but preliminary observations from fault scarps seem to indicate purely normal movement on the NE–SW directed fault segments. Finally, active faulting is confirmed by the incision of Late Quaternary alluvial terraces along rivers that cross the tectonic structures (Figs. 12B and C). Satellite images do not show any evidence of strike–slip displacement of these rivers, consistent with the field observations.

Zorin et al. (2003) inferred a link between the Chitoy–Ingoda fault zone and the Bolnay strike–slip fault system. Although more fieldwork should be done to clearly assess this connection, such a link would confirm the strongly partitionned deformation on the numerous faults that form the eastern termination of the Bolnay fault, from the Gusinoozersky–Ulan Ude basins to the Ingodinskaya basin. The extension observed in the eastern-most basin (the Ingodinskaya basin) would still be driven by left-lateral motion on a major strike–slip system. This supports the notion that the inherited crustal and/or lithospheric structure of the Transbaikal region plays a major role in the localisation and evolution of deformation. However, unlike in the BRS where the deformation is strongly localised in a single deformation zone along the edge of the Siberian craton, the deformation in the Transbaikal region is largely distributed within the various Mesozoic grabens.

7. Discussion

Most of the present-day deformation patterns within the BRS and the Transbaikal area can be explained by motion along major strike–slip fault networks that reactivate a complex crustal and lithospheric structural architecture

Within the West Sayan ranges, ENE–WSW transpressional fault systems such as the Kandatsky and South Sayan faults affect the thick Mongolian crust, driving it towards the rigid Siberian craton (Fig. 5). All of these faults merge to the NE into the Main Sayan fault that acts as a transform between the Siberian craton and the Mongolian lithosphere (Figs. 1 and 13). Displacement along this fault is controlled by motion along the transpressive faults in the Sayan ranges and should gradually increase towards the SE.

Further to the south, the E–W BDHT strike–slip system is similar to the West Sayan fault systems, except that it reaches the Siberian craton close to its southern tip. This particular geometry allows the crustal (and potentially lithospheric) material moving eastward to be diverted towards the SE. In the Tunka basin, this local change in the stress field allows extension along the NE–SW fault segments, while the general pattern remains that of a strike–slip fault system (Fig. 13). Localisation

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of the extension in the Tunka basin is also probably linked to a pre-existing (Late Cretaceous–Early Palaeogene) proto-Tunka basin, as indicated by yet poorly dated pre-Tertiary sediments localised along Proterozoic inherited shear zones (e.g. Mats, 1993; Scholz and Hutchinson, 2000; Jolivet et al., 2009; Mats and Perepelova, 2011). Basement structures oriented NW–SE, such as the faults that border the eastern edge of the Nilovsky and Elovsky basement highs, are reactivated as thrust faults (Fig. 6).

Motion along the BDHT system is then transferred and strongly localised along the eastern edge of the Siberian craton, opening up the BRS (e.g. Rasskazov et al., 2010; Mats and Perepelova, 2011). The fault pattern in the South Baikal basin is similar to the one in Tunka, associating E–W left-lateral transtensive faults (e.g. the Obruchev fault) with NW–SE compressional folding and NE–SW extensional faults. Within the Central Baikal basin, faults are again transtensive (oriented ENE–WSW) or normal (oriented NE–SW).

Within the North Baikal basins, the fault system leaves the edge of the Siberian craton and turns towards the east to connect with the Stanovoy strike–slip system (Fig. 13). The fact that deformation deviates from the strong mechanical heterogeneity associated with the edge of the Siberian craton is probably linked to the occurrence of the roughly E–W-oriented Proterozoic structures of the Muya collision zone. These structures, parallel to the general direction of the strike–slip system, allow the BDHT–Baikal system to be connected with the Stanovoy system to the east. The North Baikal basins system is interpreted as a large-scale shear zone within which inherited crustal structures oriented NE–SW induce local extension (Sherman, 1992; San'kov et al., 2000; Petit and Déverchère, 2006).

The transtensive faults in the Central Baikal basin connect to the east with another transtensive fault zone extending from the Barguzin basin to the Tsipa-Baunt basin, and continue into the Chara basin to the NE (Fig. 12). However, it is noted that the GPS and earthquake focal mechanisms in this region show almost pure orthogonal extension (Figs. 2 and 4). This large NE–SW shear zone is superimposed on the zone of thick crust observed below the Barguzin basin and to its NE (Polyansky, 2002; Suvorov et al., 2002; Petit and Déverchère, 2006) (Fig. 3). The Barguzin–Tsipa-Baunt shear zone is probably the eastern continuation of the Palaeozoic Primorsky shear zone west of Lake Baikal (Delvaux et al., 1997), implying that the Tertiary deformation is again strongly controlled by pre-existing tectonic structures. The occurrence of the Barguzin–Tsipa-Baunt fault zone broadens the general shear system encompassing a large triangular zone (Fig. 13). This zone corresponds to that of diffuse strain imaged by the distribution of seismicity (e.g. Sherman et al., 2004; Petit and Déverchère, 2006; Petit et al., 2008) (Fig. 4).

The Bolnay strike–slip fault system is parallel to the BDHT–Baikal–Stanovoy shear zone discussed above. Nevertheless, the northeastern termination of the Bolnay system is highly distributed within the Transbaikal crust. The localisation of deformation in the BDHT–Baikal–Stanovoy system is strongly governed by the huge contrast in crustal and lithospheric mechanical properties between the Siberian craton and the Mongolia–Transbaikal lithosphere (e.g. Chemenda et al., 2002; Polyansky, 2002; Petit and Déverchère, 2006; Petit et al., 2008). The crust in the Transbaikal area is also affected by a strong structural inheritance mostly generated by the Mesozoic closure of the Mongol–Okhotsk Ocean and the extension phase that followed (e.g. Zheng et al., 1991; Davis et al., 1996, 2001, 2002; Van der beek et al., 1996; Webb et al., 1999; Zorin, 1999; Darby et al., 2001; Fan et al., 2003; Meng, 2003; Wang et al., 2006; Donskaya et al., 2008; Daoudene et al., 2011). As demonstrated for the Gusinoozersky, Ulan Ude and Ingodinskaya basins, the Tertiary faults are superimposed on Mesozoic faults, themselves sometimes superimposed on Palaeozoic shear zones (Figs. 8, 9 and 11). This inherited crustal structure allows the distribution of sinistral strike–slip deformation along the Bolnay fault into a number of smaller faults, creating a transtension zone several hundreds of kilometres wide in the Transbaikal area. This transtension zone will probably ultimately connect to the North Baikal–Stanovoy shear zone through strike–slip systems such as the Uda–Vitim dislocation zone (Fig. 13). In that respect, while the BRS can still be considered as a narrow rift the general extension framework in the northwestern region of the Amuria plate is better described by a wide rift model (Buck, 1991).

It is interesting to contemplate the effects of these two deformation styles on the structure of continental margins where extension evolves to true rifting. For example, a mechanism involving interaction between extension and a strong pre-existing tectonic structure could explain the formation of the numerous small continental blocks that rifted from the northeastern margin of Gondwana during the Jurassic and travelled across the Eastern Mediterranean Neotethys until their accretion to Eurasia (e.g. Robertson et al., 1991; Mackintosh and Robertson, 2009; Robertson and Ustaömer, 2009). Robertson et al. (1991) describe the Middle Jurassic Eastern Mediterranean Neotethys as being composed of a series of small micro-continents, tens to hundreds of kilometres wide, separated by small-scale oceanic basins. Rifting of such small pieces of continental crust could correspond to the last evolutionary stage of a BRS–Transbaikal system, with the small oceans and continents forming through distributed extension along inherited structural networks.

If the model presented above satisfyingly describes the present-day deformation pattern, the Late Cretaceous–Early Palaeogene slow extension in the Tunka, South Baikal and Barguzin basins remains to be explained. Many authors suggested that the Late Oligocene–Early Pliocene extension phase was driven by far-field effects of the Pacific–East Asia active subduction (e.g. Rassakov, 1993; Delvaux et al., 1997; Polyansky, 2002), and that this mechanism might still be active (e.g. Calais et al., 2003, 2006; Vergnolle et al., 2007). Using apatite fission track analysis on the Barguzin range that separates the Barguzin and North Baikal basins, Jolivet et al. (2009) showed that the movements on the bordering normal faults initiated at 65–50 Ma. This initial extension is clearly neither related to the effects of the India–Asia collision, nor to asthenospheric upwelling (Jolivet et al., 2009). Furthermore, it potentially created a continuum of deformation between the Mesozoic extension in the Transbaikal area and the onset of rifting in the BRS. The addition of stress generated by the India–Asia collision only accelerated the deformation, potentially creating new faults but mainly reactivating inherited Palaeozoic and Mesozoic tectonic structures within and along the boundaries of the Amuria plate.

Finally, the eastward termination of the Main Sayan fault must be addressed. This fault is generally considered to end along the southwestern tip of Lake Baikal, the strike–slip motion along the fault being relayed by extension within the rift (e.g. Delvaux et al., 1997; Petit and Déverchère, 2006; Mats and Perepelova, 2011). However, in the present model, the crust of the Transbaikal area (Zone C in Fig. 13) is extending towards the SE, while the crust south of the BDHT system extends towards the east (Zone B in Fig. 13). This difference in motion could potentially create a shear zone accommodating the eastward expansion of the Transbaikal lithosphere and propagating the transform Main Sayan fault towards the SE (Fig. 13).

8. Conclusion

The Tertiary opening of the Baikal Rift System and extension in the Transbaikal region can both be explained by strike–slip movements on major shear zones strongly influenced by inherited crustal and lithopsheric structures.

The triangular shape of the rigid Siberian craton as well as the strong mechanical contrast between Siberian and Mongolian lithospheres tightly control the tectonic structure of the eastern termination of the Busingol–Darkhat–Hovsgol–Tunka system. When reaching the southern tip of the craton, the Tunka fault merges with the Main Sayan fault and all movement is transferred along the eastern edge of the craton into the Baikal Rift System. This is not the case for the eastern termination of the

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Bolnay–Uda–Vitim fault system. When reaching the Transbaikal region, this major fault zones encounters crust that has been strongly dissected by the Palaeozoic and the Mesozoic deformation. The Tertiary deformation reactivates some of these inherited structures and distributes itself over a wide area encompassing the Gusinoozersky basin as well as others towards the east (Fig. 13).

The Main Sayan fault acts as a transform between the thick Mongolian crust and the Siberian craton. Movement on this fault is governed by activity along the major strike–slip systems of the West Sayan ranges and should increase towards its southern termination. The southeastward crustal extension in the Transbaikal area could potentially induce the eastward propagation of a shear zone prolongating the Main Sayan fault towards the SE (Fig. 13). However, due to its position along the mechanical discontinuity between the Mongolian crust and the Siberian craton, the Main Sayan fault is a strongly localised structure. In the Transbaikal crust, strike–slip deformation would not be influenced by such a mechanical contrast. This would potentially lead to the formation of a much wider shear zone with NW–SE strike–slip faulting distributed among several parallel faults.

Finally, if the Baikal Rift System by itself can be considered as a narrow rift, the entire extensional system in the NW Amuria plate (the Baikal and Tranbaikal area) corresponds to a wide rift system. The Amuria plate is therefore not behaving as a rigid block but rather as a highly distributed extensional zone.

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