Velocity structure around the Baikal rift zone from teleseismic and local earthquake traveltimes and geodynamic implications

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Abstract

We present new results on the velocity structure of the Baikal rift zone, Asia, deduced from a comparative teleseismic and local tomography analysis. The aim of this paper is to better identify the role of deep mantle processes versus that of far-field tectonic effects on the occurrence of extensional tectonics within a continental plate. We use 36000 traveltimes of P-refracted waves from the ISC catalogues and Pg and Pn traveltimes of 578 earthquakes recorded by the Russian regional network to determine a velocity model by the use of local and teleseismic inversion procedures. The models show that some velocity patterns are continuous from the surface down to at least 400 km. Among them, a narrow negative anomaly goes through Mongolia and follows the southern and eastern margins of the Siberian craton: this structure is interpreted as a thin mantle plume rising beneath the rift axis. However, our results do not evidence any wide asthenospheric upwarp at this place. Other velocity anomalies observed near the surface are not deeply rooted. In particular, a negative anomaly is observed at shallow levels (48 km) beneath the northern third of Lake Baikal, which is disconnected from deeper structures. It may be explained by the existence of underplated magmatic material at the bottom of the crust. By comparing the geometry of deep-rooted anomalies to the present-day stress field patterns, we conclude that the sub-lithospheric mantle dynamics is not the main factor controlling extensional processes in the Baikal rift. However, it does contribute to a thermal weakening of the lithosphere along a mechanical discontinuity bounding the Siberian shield. We finally conclude that three favourable conditions are gathered in the Baikal area to generate extension: far-field extensional stress field, mechanical inherited lithospheric weakness and heat supply. Further studies should help to precise the genetic link between these three factors. © 1998 Elsevier Science B.V. All rights reserved.

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1. Introduction

During its geological history, a continental plate may be partly or totally dislocated, giving birth to new plate boundaries. Continental rifting is thus an important stage of the evolution of a continental lithosphere. Deep mantle processes such as asthenospheric diapirism and plume genesis are known to play an important role during the initiation and development of rifting (Nicolas et al., 1994). However, in several cases, the relative importance of man-
tle processes with respect to that of plate tectonics effects is not clearly determined, mainly because rifting often results from the combination of both factors (see a recent review in Ruppel, 1995). It is also known that the lithospheric fabric inherited from previous orogenic stages strongly influences the development of elongated rift systems (Vauzech et al., 1997).

Present-day continental rifts offer the opportunity to understand what happens during the early stages of continental break-up. In this paper, we intend to investigate the mantle structure over a part of the wide Asian continent where continental rifting is currently observed. Our aim is to contribute in shedding light on the respective roles of asthenosphere upwarping and far-field boundary conditions in the development of intracontinental rifts. Because of its entirely intracontinental position, far away from any plate boundary, the Baikal rift zone (BRZ) appears particularly well suited to address this point.

We first present hereafter compared results of teleseismic and local earthquake tomography in order to image the upper mantle and lower crust structure beneath the BRZ and Mongolian regions. Then we compare this velocity distribution with previous analyses of the present-day stress field patterns in order to propose an interpretation in terms of mechanisms of rift opening. The investigation of mantle and crust structures is performed by the application of two different schemes using different seismological data. The first one, the so-called ‘inverse teleseismic scheme’, uses teleseismic data for studying the velocity variations at a depth range of 100–600 km. The second approach (‘local scheme’) is based on the inversion of local earthquake travel times recorded by the regional seismological network: source–receiver spatial distribution within the study region, at epicentral distances up to 1500 km, enables investigation of the velocity structure at depths up to 70 km.

2. General setting and debated questions

The first important observation is that the BRZ develops along (or close to) a major suture zone delimiting two structural entities: the Palaeo–Mesozoic Baikal folded system and the Archaean Siberian craton (Fig. 1). Continental accretion and collage around the border of the Siberian shield have ended at late Mesozoic time (Melnikov et al., 1994; Delvaux et al., 1995). Therefore, the inherited tectonic fabric of the continental lithosphere seems to primarily control the location and propagation of the Baikal rift. A second major fact is that extensional processes have been very active for 3 m.y. and are still currently observed in the BRZ, as evidenced by a high seismic activity (Golenetsky, 1990) (Fig. 1) and significant crustal stretching resulting in a young and contrasted topography controlled by normal faulting (Van der Beek, 1997).

Previous studies in the BRZ have led to contrasted interpretations regarding its deep structure and the onset of rifting. Several authors assume that a wide asthenospheric diapir reaches the depth of 40–50 km beneath the Lake Baikal and drives extension. Deep seismic sounding data have evidenced the presence of a low-velocity zone (7.8 km/s) beneath the Moho close to the rift axis (Puzyrev, 1981; Krylov and Mishenkin, 1984). Analyses of gravity anomalies (Zorin, 1981; Logatchev et al., 1983), and seismological data (e.g., Gao et al., 1994a) also led these authors to propose that the lithosphere in the BRZ is reduced to the crust thickness, i.e. 40–50 km, over a width of more than 300 km beneath the rift axis. On the other hand, on the basis of plate kinematic analyses (Zonenshain and Savostin, 1980) and physical models of the deformation in the Asian continent (Davy and Cobbold, 1988; Peltzer and Tapponnier, 1988), several authors suggest that the India–Asia collision may have triggered extension in the Baikal region. Indeed, recent analyses of the present-day stress field and numerical strain modelling indicate that the current deformation is mainly linked to the effects of the India–Asia collision (Petit et al., 1996; Lesne et al., 1998). Heat flow data (Lysak, 1992) depict a globally ‘cold’ geotherm (40–60 mW/m²) with only locally elevated values (80 to more than 100 mW/m²) inside the rift basins. Moreover, analyses of mantle xenoliths (Ionov et al., 1995), as well as gravity models of the mechanical behavior of the lithosphere (Ruppel et al., 1993; Burov et al., 1994; Van der Beek, 1997) show that the lithosphere in this region cannot be significantly thinned, as would predict the hypothesis of a wide, hot, asthenospheric diapir.
Fig. 1. Seismicity and active faults of the Baikal area. Open circles are epicentres of $M > 3$ events recorded by the regional network between 1960 and 1996. Thick lines are active faults. The thin rectangle locates the cross-section of Fig. 9. The Siberian craton is colored in grey. Note that the Siberian craton is nearly aseismic, whereas the Baikal folded zone depicts a relatively high background activity. The rift system is underlined by the highest epicentre density.

Are these characteristics comparable with those observed in other major rift systems like the East African rift (EAR)? Results of regional tomography (e.g. Achauer, 1992; Davis et al., 1993) and recent global high-resolution tomographic models (Zhang and Tanimoto, 1993; Grand et al., 1997; Van der Hilst et al., 1997) show that the seismic structures beneath the Baikal rift and the eastern branch of the EAR are significantly different: the Kenya rift area is obviously connected with a large negative velocity anomaly, which is observed throughout the upper and lower mantle, whereas nothing similar is observed beneath the Baikal rift. However, these global-scale studies do not allow to disregard the fact that small-scale mantle perturbations could take place in the BRZ and play a significant role in the rift opening, as proposed by Gao et al. (1994a,b).

3. Velocity structure beneath the Baikal region from teleseismic data

3.1. Data and method

The presented model is obtained by applying the so-called ‘inverse teleseismic scheme’ (ITS) (Koulakov, 1998). Unlike conventional teleseismic schemes which use the information of far-distance sources recorded by the regional seismological network of the study region, the ITS uses earthquakes located within the study region and recorded by stations of the worldwide seismological network. The ITS ray paths have been previously used in several regional tomographic inversions (e.g., Spakman et al., 1993; Puspito et al., 1993), but they represent a minor part of the data used in these studies, and they did not undergo any specific processing. Here and in
Koulakov (1998), the ITS procedure is applied for the first time in its pure form. Common assumptions are shared by the ITS and conventional schemes: in particular, all possible factors influencing the travel-time of a ray outside the study volume are included into two constants which are the station and source corrections. But the values and signs of the corrections applied in both procedures are different. For instance, the influence of the lower mantle is included into source corrections in the first procedure, and into station corrections in the ITS; therefore, ITS requires specific statistical procedures to compute relevant residuals (Koulakov, 1998).

We summarize here the reasons why the ITS is more likely to provide a more accurate velocity model than the conventional scheme, especially when applied to our region of interest. Both approaches have been previously applied to the Altai–Sayan region: the model TSA-94 (Kulakov et al., 1995) was obtained using a conventional teleseismic scheme, whereas another one was obtained using the ITS (Koulakov, 1998). First of all, the ITS provides a high number of station–event couples that enables to compute more accurately source and station corrections, which are very important for residuals estimation. For instance, the average number of stations recording one given event was no more than 8–10 in the model TSA-94, whereas it is higher than 30–40 in the ITS model. Second, the number of data available for ITS is significantly higher than for the conventional scheme (36000 and 6000, respectively). Third, a denser distribution of sources compared to stations (Fig. 2A) enables us to study relatively shallower levels: the average distance of 200–300 km between stations in the model TSA-94 provides enough ray intersects at depths of about 200–300 km, whereas the same number of intersects is already obtained at the depth of 100 km in the ITS. Finally, the information of the worldwide seismological network is stored on international databases and can be easily extracted for any tectonically active region. As an example, let us recall that several years of hard manual work were necessary to get the 6000 traveltimes required for the model TSA-94 (Kulakov et al., 1995), whereas data from the ISC file service were recovered nearly instantaneously for the application of the ITS (Koulakov, 1998). In addition, the quality of ISC data, once carefully selected, is generally higher since the worldwide seismological network uses only the best stations from regional networks.

One drawback of the use of ITS is that epicentre locations can have a strong influence upon final results: indeed, it is well known that the accuracy of hypocentre determination by teleseismic rays is not high. In order to reduce the mislocation of seismic sources, a part of the earthquakes used in this study (i.e. most events occurring within the Russian territory; see Fig. 2A) have been relocated using regional traveltimes. Previous tests on the effects of source mislocations have shown that, in the case of a random shift of epicentre positions up to 20 km, the result of inversion is still stable (Koulakov, 1998). Another assumption of the teleseismic inversion is that the ray path is not highly dependent on perturbations of the reference model in its upper part (e.g., Hirahara, 1993). It allows us to apply a linearized approach using the rays constructed in a reference medium, which is the model PEM-C from Dziewonski et al. (1975). It was shown that if the anomalies are lower than 4–5%, a linearized approach can be applied with satisfying accuracy. We have to recall that the teleseismic inversion only provides relative velocity variations, and not absolute values.

For this study, we use traveltimes of about 500 earthquakes (Fig. 2A) recorded by approximately 1500 stations of the worldwide network in the time span 1961–1984 (36000 traveltimes of refracted P-waves from the ISC catalogues). We selected events recorded by at least 10 stations of the worldwide seismological network; most of them are recorded by 30 to 40 stations. Relocation of the events recorded by the regional network provides an estimate of the maximum error on epicentre location of 10–15 km. Events with high initial residuals (up to 10 s) are kept, since stations and/or source corrections may partly explain them. Inversion of the velocity structure is performed independently in 15 prisms of N–S and E–W orientations over 1500 km long, 200 km wide and 600 km depth, covering the study area and partially overlapping each other (Fig. 2B). It allows us to operate with fairly small matrices (300 × 3000 elements) which are inverted by the SVD (singular value decomposition) method. Furthermore, it provides an estimate of the model reliability by comparison of anomaly patterns in the overlapping areas, which were carefully checked.
The inversion results in a reduction of residuals of 65–75%, depending on the region.

The parametrization of the velocity model is performed by the vertex method, which is detailed in Kulakov et al. (1995). Velocity perturbations are computed at the nodes of a 3D grid. Vertices are located within the study volume according to the ray density, so that the resolution of the output model...
Fig. 3. Three-dimensional velocity model of the P-velocity anomalies in the upper mantle obtained by teleseismic tomography at depths of 100 (a), 200 (b), 400 (c) and 600 (d) km. Contour intervals are given every 1%. Dark areas depict negative anomalies. The position of nodes where velocity anomalies were calculated are shown by small circles.
Fig. 3 (continued).
is directly related to the density of input information (Fig. 3). The velocity distribution between these nodes is linearly interpolated using a 3D triangulation. Average velocity perturbations are obtained for the fifteen different prisms, and are presented in Fig. 3 on four ‘slices’ at depths of 100, 200, 400, and 600 km. It is worth noting that the vertex distribution is highly heterogeneous at shallow (100–250 km) levels, leaving large areas of low resolution. The nodes are more equally distributed at deeper levels.

3.2. Results

The resulting tomographic model evidences significant velocity perturbations beneath the Baikal–Mongolia region (Fig. 3). One of the most important features of the obtained model is a periodicity of N–S-oriented positive and negative perturbations (+3% and −3%, respectively) of about 200 km wide extending from the southern border of Mongolia up to the Siberian craton. This pattern is clearly observed down to a depth of at least 400 km (Fig. 3a–c). Two negative anomalies (−1 and −2%) are located beneath the depressions of the Ubsu-Nur and Hubsugul lakes (see Figs. 1 and 3). The easternmost one is deeply rooted, and extends in a N–S direction from the southern border of Mongolia through Lake Hubsugul, up to the Siberian craton. When reaching the southern border of the Siberian plate, it turns to a northeast direction beneath the western shore of Lake Baikal, thus following there the general rift trend. A large positive anomaly of similar shape is bounding the eastern side of the rift; two other positive anomalies are located east and west of the Ubsu-Nur depression.

It is therefore possible to correlate at shallow levels (100–200 km) the pattern of velocity anomalies with the structural fabric of the BRZ: both are trending in a SW–NE direction. This seems not to occur further south, i.e. in Mongolia, where the main active faults are E–W or NW–SE (Baljinnyam et al., 1993; Schlupp, 1996) and velocity anomalies are N–S (Fig. 3a,b).

At deeper levels, the velocity variations are of lower amplitude (1–2%). At 400 km, N–S-oriented positive and negative anomalies are cross-cutting the southern part of the rift; hence they are no more clearly connected with the surficial rift structure. A positive anomaly is observed beneath the northern half of Lake Baikal and east from it. The negative anomaly located beneath the southwestern border of the Siberian craton (and especially below Lake Hubsugul) becomes wider at greater depths (600 km), which is not the case below the Ubsu-Nur depression (Fig. 3d). Furthermore, it is worth noting that contrary to the latter region, the topography of the Hubsugul region is characterized by a wide, elevated doming shape: the Hangai dome (Windley and Allen, 1993).

From these results, we do not see any evidence for a large-scale velocity reduction linked to a major asthenospheric upwarp, but rather a pattern of relatively narrow (about 200 km) and continuous positive and negative anomalies, the most important one being located west of Lake Baikal, beneath the Siberian craton. Are these structures continuous at shallower (e.g., lithospheric and crustal) levels? The teleseismic approach alone does not allow us to enter upon this point. The second step of this study thus aims at determining the velocity structure at shallower depths (0–70 km) in order to compare it with the deep structures observed by teleseismic tomography.

4. Local scheme

4.1. Data and traveltime residual analysis

For this purpose, we use a database consisting of Pn arrival times of 1201 events of magnitude >3.3 recorded by the Russian local and regional network between 1966 and 1992, in an area located between 97°E and 110°E and between 48°N and 54°N. We have selected 374 events as good-quality data for the following study, according to the following criteria: epicentral distance <200 km; azimuthal gap <200°; number of weighted phases >7; hypocentre mislocation between two different velocity models <30 km; ratio of greatest versus lowest singular values of the partial derivatives matrix <100; last hypocentre adjustment <50 km; standard errors on hypocentre location <15 km; RMS (root mean square residual) <2.5 s. We add Pn arrival times of 204 events which were selected and used for a previous tomography analysis in the north Baikal rift (Petit and Deverchère, 1995) (Fig. 4).
Fig. 4. Map of the 374 earthquakes used for local inversion (dots). Triangles are stations of the regional network.

The data processing begins with the determination of the best-fitting 1D velocity model for the study region. As shown in Fig. 5, the dispersion of residuals at all epicentral distances is high, which seems to depict a-priori high heterogeneous structure in the study region. Since the 1D inverse solution is not very stable (in the sense that large variations of the velocity model may cause small variations of the traveltime curve), only a simplified representation of the 1D velocity model is significant. Our model is described by five parameters:

\[
\begin{align*}
p_1 &= z \text{ (Moho)} \\
p_2 &= v(z_1), \\
p_3 &= v(z_2), \\
p_4 &= v(z_3), \\
p_5 &= v(z_4),
\end{align*}
\]

\[
\begin{align*}
z_1 &= 0 \text{ km} \\
z_2 &= p_{1-2} \text{ km} \\
z_3 &= p_{1+2} \text{ km} \\
z_4 &= 300 \text{ km}
\end{align*}
\]

Between \(z_1, z_2, z_3\) and \(z_4\) the velocity distribution is approximated linearly. Calculation of velocity parameters is performed by a trial and error method. The best-fitting synthetic traveltime curve is found for an average crustal thickness of about 45 km (Fig. 5 and Table 1), although it is not clearly associated with a significant RMS (root mean square residual) decrease because of the strong dispersion of the residuals.

A preliminary analysis of the time residuals \((T_{\text{obs}} - T_{\text{cal}})\) computed with respect to this best-fitting 1D model is made in order to estimate the main patterns of velocity variation in the study region. In this first step, the Pn arrivals are interpreted as refracted waves, thanks to the use of velocity gradients in the 1D model. The rays are divided into four groups depending on the epicentre–station distance (and on
Fig. 5. Initial data for the local scheme. (A) Traveltime–distance plot. (B) Residuals calculated relatively to four models using different Moho depths of 35, 40, 45 and 50 km. The parameters of 1D velocity distribution were optimized in order to obtain the best-fitting model in each case (see text).
the depth of the central point). Each group contains approximately the same number of rays. The values of the residuals related to the central point are averaged within a grid of 100 × 100 km. This averaging approach provides a first-order estimation of velocity anomalies (see Fig. 6).

In the layers 1 and 2 corresponding to crustal levels, the residuals depict a slow area beneath Lake Baikal, probably related to the thick sediment and water infill (Hutchinson et al., 1992). Negative residuals \( T_{\text{obs}} < T_{\text{cal}} \) are found west of the island Olkhon in the central part of the lake, north and east of Lake Hubsugul, and at the eastern rift end (Fig. 6A,B). In the first upper mantle layer (layer 3), positive residuals are restricted to the northern third of Lake Baikal and to a large anomaly covering the southern part of the rift. Null and negative residuals occupy the central and eastern parts of the rift. Finally, the deepest layer depicts quite different features: the central part of the rift is there characterized by an area of positive residuals which extends east of the lake. The northern third of the lake, as well as the major part of its eastern side, is covered by an area of negative residuals (Fig. 6C,D).

4.2. Local tomography inversion

These data are then used for a 3D tomography inversion, using the routine developed by Abers and Roecker (1991) and Roecker et al. (1993), as follows.

The velocity model is divided into two layers and one half-space (Fig. 7). Layer 1 extends between 0 and 15 km; it is composed of only two meta-blocks, because of the poor ray density: one ‘peripheric’ meta-block and one ‘central’ meta-block covering the rift basins. Numerous tests are performed in order to determine the best parametrization for layers 2 and 3, especially concerning the depth of the layer 2–layer 3 interface which represents the Moho. The average crustal thickness is very high (48 ± 7 km); it is determined by fitting the traveltime–distance curve of Pn waves (Fig. 5). Determination of the most likely Moho geometry is based on previous results of gravity modelling (Petit et al., 1997). The selected model depicts the highest variance reduction and allows a good spatial coherence of the inverted velocities between the different blocks. West of the rift area, the crustal thickness is very high (55 km); the blocks covering the rift axis and its eastern side depict a thinner crust (46–43 km); finally the blocks surrounding the rift area have a 50 km deep Moho.
The variance reduction associated to this model is more than 25% (0.990 to 0.740 s\(^2\)).

As already observed in the residuals, the resulting model shows a strong velocity reduction (−3%) over the rift basins in the upper layer, which is probably connected with a thick, low-density, sedimentary infill (Fig. 7a). The lowest crustal layer (Fig. 7b) also shows a negative anomaly covering the southern and northern thirds of Lake Baikal, while positive anomalies are found in its central third. Finally, the mantle layer depicts a strong velocity reduction at the northern and southern extremities of the lake. A small (1−2%) velocity reduction is found over the southern basin of the lake. Velocity increases at the central part of the rift. These results fairly well agree with those of residual analysis for layers 1, 2 and 3, which means that the approximation made on the kinematics of refracted waves is acceptable (see Figs. 6 and 7).

5. Discussion and conclusion

5.1. Depth extent of velocity anomalies

The first aim of this study was to combine independent images of the P-wave velocity structure obtained by traveltime tomography at teleseismic and local distances. First of all, one may note a good correlation in map view between velocity patterns obtained at ‘deep levels’ (50–70 km) with a local analysis (Fig. 6C,D, and Fig. 7c) and at ‘shallow levels’ (100 km) by the teleseismic model (Fig. 3a,b). This correlation may be considered as an evidence for the reliability of both models, which have been obtained using absolutely independent approaches and datasets.

Some anomalies observed in the upper mantle pass through the Moho interface up to the lower crust. The most important one is a negative anomaly of about 150–250 km width and 2–4% amplitude observed beneath Lake Hubsugul and the southeastern margin of the Siberian craton (Irkutsk area), throughout the upper mantle, i.e. from 500 to 100 km. At 48–70 km depth, this anomaly seems to be shifted beneath the southern and central lake’s basin, whereas it is remarkably continuous throughout the crust in the region of Hangai–Lake Hubsugul, an area which is characterized by late Cenozoic basaltic volcanism and a wide and high topographic dome (Windley and Allen, 1993), contrasting with the Ubsu-Nur depression where no deeply rooted negative anomaly and very little volcanism are observed.

Conversely, a velocity reduction is observed at the top of the upper mantle in the northeastern part of the lake (Fig. 7c), which is not seen at deeper locations. It was already detected on the accurate global tomography model of Levshin et al. (1996) and on deep seismic soundings (Puzyrev, 1981; Krylov and Mishenkin, 1984). This anomaly might be linked to the presence of underplated anomalous mantle at the bottom of the crust. This hypothesis would also explain the existence of a high topography underlined by low Bouguer anomaly in this area.

5.2. Consistency with previous studies

Our main result is that the mantle structure beneath the BRZ is characterized by a negative anomaly which seems deeply rooted beneath the Mongolian (Hubsugul) region; its narrow lateral extent suggests that it is not linked to a major asthenospheric upwelling, but rather to a narrow mantle plume impinging on the lithosphere. This is in agreement with recent high-resolution global models (Zhang and Tanimoto, 1993; Grand et al., 1997; van der Hilst et al., 1997) and with studies of mantle xenoliths (Ionov et al., 1995) which do not evidence any large mantle diapir reaching the Moho beneath the Baikal rift. Our image of the deep mantle structure beneath the BRZ therefore quite differs from the one proposed by Gao et al. (1994a). This discrepancy might be related to the use of significantly different a-priori models. Indeed, the negative delays of teleseismic rays observed over the Baikal rift can be interpreted either by an existence of a 3D positive velocity anomaly in the mantle — as we obtain in our model — or by a significant reduction of the
lithosphere thickness — as inferred by Gao et al. (1994a). However, our velocity model appears to us more objective because the a-priori assumption that all the traveltime delays are related to the geometry of the lithosphere–asthenosphere boundary — as postulated by Gao et al. (1994a) — can lead to an important overestimation of the asthenosphere upwarp. Moreover, our model seems more in agreement with the average low heat flow (Lysak, 1992), the lack of recent volcanism, and the relatively high lithospheric strength (Petit et al., 1997) observed in this area.

5.3. Deep mantle structure and present-day stress field

Two kinds of velocity anomalies are evidenced: shallow-rooted anomalies possibly related to crust and upper mantle density variations, and deep-rooted anomalies which are probably related to mantle dynamics, the latter being obviously the most important to understand the possible influence of mantle convection over the rift opening. For this purpose, we compare the geometry of the main negative anomalies with the stress field patterns observed in the Baikal rift (Petit et al., 1996): we can see that the negative seismic anomalies are not systematically connected with the presence of an extensional stress field (Fig. 8). Indeed, the N–S anomaly which runs from the southern Mongolian border up to Lake Hubsugul, crosses successive areas of compressive, wrench and transtensive stress regimes (Baljinnyam et al., 1993; Petit et al., 1996). Moreover, it covers more or less the western half of the rift, where very sharp stress field changes are observed. In addition,
extensional processes are quite well expressed in the northeastern part of the rift (Deverchère et al., 1993), which is located above a large positive anomaly. We can thus infer that the observed mantle structure does not strongly influence the present-day stress field of the BRZ. It is hence difficult to conclude from these results that the Baikal rift opening is driven by a mantle flow occurring beneath the rift axis, as proposed by Gao et al. (1994b). Their assumption is based on SKS anisotropy measurements which show a fast polarization in a direction orthogonal to the rift. We feel that this anisotropy has little effect on our tomography results, mainly because (1) we use various ray azimuths, therefore the effect of anisotropy is averaged with respect to that of velocity anomalies, and (2) we observe consistent velocity patterns at depths up to 600 km, whereas the source of shear-wave anisotropy is supposed to lay at 250 km or above (Gao et al., 1994b). We may note that the interpretation of SKS splitting is highly ambiguous in the Baikal region: indeed, directions of active strain, fast polarization, and modelled finite strain are roughly the same (Davis et al., 1997), which may be attributed to asthenospheric flow, inherited ‘frozen’ lithospheric fabric or shear at the base of the lithosphere. Conversely, the fast polarization directions differ from the maximum horizontal elongations in the calculated finite strain of the north Mongolian region (Davis et al., 1997), the former striking roughly orthogonal to our N–S, deeply rooted anomaly: this may reveal a more pronounced active strain of the asthenosphere below this part of the Asian lithosphere, as proposed by Windley and Allen (1993).

Fig. 9. Simplified interpretative cross-section of the tomographic images in terms of mantle dynamics (see Fig. 1 for approximate location). Dark and white areas represent the regions of upwelling ‘hot’ mantle and downgoing ‘cold’ mantle, respectively. Open arrows represent the extensional stress field.
5.4. Interpretation and conclusion

Fig. 9 presents an interpretation of the teleseismic and local velocity models on a cross-section located around the central and northern parts of the lake (Fig. 1). The negative seismic anomaly beneath the Siberian craton is interpreted as a narrow, hot mantle plume which reaches the bottom of the resistant Siberian plate and follows its border in the Baikal area. The geometry of the anomaly suggests that the rising hot material should be guided into the lithosphere by the craton rim, as already proposed by Kiselev and Popov (1992). This plume injected within the lithosphere along the suture bounding the Siberian craton may play an important role in the extensional process by adding a thermal weakening to the mechanical discontinuity separating the Siberian plate from the Baikal folded system. Therefore, it seems that localization of rifting is highly dependent upon the existence of a mechanical discontinuity bounding the craton which guides the impingement of mantle material on the asthenosphere, but also by the presence of inherited rheological contrasts within the continental lithosphere, the Palaeo–Mesozoic Baikal folded zone being less resistant than the Archaean Siberian craton (e.g., Petit et al., 1997).

As a conclusion, it is possible to propose that both active and passive mechanisms take place in the Baikal rift zone. Our results suggest that a narrow mantle diapir following the southern border of the Siberian craton can be responsible for the weakening of the lithosphere along the suture zone, which favours the development of extensional processes in response to far-field tectonics caused by the India–Asia collision. This interpretation mainly differs from the one proposed by Windley and Allen (1993) and Gao et al. (1994a,b) in the size and position of the involved mantle structure. Detailed studies of Cenozoic volcanism and numerical modelling of the rift opening should help to understand the varying relative importance of deep mantle and plate tectonics effects during the rift history.

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