The 1999 $M_w$ 6.0 earthquake sequence in the southern Baikal rift, Asia, and its seismotectonic implications

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SUMMARY
The 1999 February 25 earthquake ($M_w = 6.0$) is the largest event to have occurred in the Baikal rift system since the 1959 $M = 6.8$ earthquake. This earthquake sequence took place in the southern basin of Lake Baikal, a region characterized by a sharp transition from strike-slip faulting on the Main Sayan Fault to normal faulting along the border faults of the lake. Thanks to the development of the regional network and a temporal station installed after the main shock, we performed a relocation of the aftershocks to describe the rupture area and to determine fault-plane solutions. We show that the main shock was caused by normal faulting on an intrabasin fault striking northeast and steeply dipping to the northwest. Aftershock hypocentres did not cluster at the same depth but were spread from 5 to 25 km depth. Stress inversion carried out using focal mechanisms of the sequence reveals a radial extensive stress regime with a minimum principal stress axis $\sigma_3$ (N155$^\circ$E) slightly oblique to the regional stress field (N139$^\circ$E), which depicts a stronger intermediate $\sigma_2$-axis. This behaviour suggests transient stress perturbation induced by the main shock in the neighbouring zones of the rupture. Strain release deduced from GPS data and historical seismicity shows that within the southern tip of the Baikal rift basin, a significant stress release has occurred for at least two centuries, probably through active normal faulting shared among several faults, while the Main Sayan strike-slip fault further west was undergoing interseismic strain accumulation.

Key words: earthquake location, fault-plane solutions, Lake Baikal, seismic stress and strain.

1 INTRODUCTION
On 1999 February 25 at 18:58 UT, an earthquake of magnitude $M_w = 6.0$ occurred in the centre of the southern Lake Baikal basin, one of the deepest basin of the Baikal rift system, Siberia (Fig. 1). The event was strongly felt in a large area: the intensity level was estimated to be $I_0 = $ VIII (in the MSK-64 scale) and VI–V in the largest city of Irkutsk with a population of 600 000. The main shock was followed by a long series of aftershocks and was also preceded by numerous foreshocks. This sequence started in early February and culminated on February 25 with the magnitude $M_w = 6.0$ main shock.

This earthquake sequence is of particular interest, since the main shock has been the largest earthquake to shake the southern Baikal region over the last 40 yr, i.e. since the 1959 $M_s = 6.8$ Central Baikal earthquake. The 1999 earthquakes were the first events inside Lake Baikal which could be located rather accurately, owing to the development of the seismological regional network and to the addition of a temporal station close to the epicentral area after the main shock. Furthermore, the epicentral area is located at the junction of the left-lateral strike-slip Sayan Fault to the west, and the main normal faults of Lake Baikal to the east (the Obruchevsky and Chersky faults) (Fig. 2), thus providing important clues to the mechanics of opening of the southern termination of the Baikal rift: indeed, there is no clear understanding of the way the strike-slip motion turns into dip-slip motions through time, i.e. on the mechanics of opening of the southern termination of the Baikal rift (see e.g. Balla et al. 1991; Agar & Klitgord 1995; Delvaux et al. 2000). More specifically, we would like to address the question of whether this earthquake sequence should be related to one of the normal border faults or to another fault located inside the basin, which should allow us to elucidate the present-day style of opening and to better analyse the space and time evolution of faulting and its seismic potential. For this purpose, we will first examine the inherited structures of this area and then the seismicity before, during and after the earthquake. By comparing the spatial extent of this activity with the main structural features and the time and space evolution of the stress field deduced from focal mechanisms of earthquakes, we will finally have the opportunity to go from a large-scale to a detailed investigation of tectonic stress distribution in this modern...
rift system and to discuss the tectonic meaning of this sequence and its relevance in terms of seismic potential in this region.

2 Geological Structures of the Southern Baikal Rift Region

2.1 Pre-rift structures

The Baikal rift was formed at the junction of the ancient Siberian Platform and the Sayan-Baikal fold belt. The Siberian platform is separated from the fold belt by a suture zone (Zamaraev 1967). The southern edge of the platform is bounded by two segments of the suture, which are the Main Sayan and Primorsky shear zones (Fig. 1).

The basement of the south Baikal basin is made with several geological complexes. Towards the north, four of these complexes—Early Precambrian (Archean), Upper Proterozoic, Riphean to Lower Palaeozoic and Mesozoic—make up the platform basement and its uplifted basement of the Siberian platform. It is mainly composed of basic crystalline schists, amphibolites, gneisses, granitoids, ultrabasic rocks and other rocks under the granulite and amphibolite metamorphic facies (Melnikov et al. 1994). The Late Proterozoic–Riphean rocks of the Baikalian complex were formed in the marginal narrow depressions of the platform, which extend along the northeastern Baikal shoreline. This complex consists of terrigenous-carbonate rocks deformed as linear northeast-striking folds and often affected by numerous reverse and thrust faults (Alexandrov 1990). Finally, the Riphean to Lower Palaeozoic terrigenous and carbonate sediments make up most of the platform sedimentary cover which is only slightly deformed. Along the River Angara, a Jurassic molasse fills the southeastern centricline of the piedmont foredeep (Fig. 1). A gently dipping thrust zone striking...
The 1999 southern Baikal rift earthquake sequence

2.2 Synrift and neotectonic structures

Lake Baikal is the central part of the Baikal rift zone and is subdivided into three basins from south to north. They are typical half-grabens with steeply dipping master faults on their northwestern side (Logatchev & Zorin 1992; Logatchev 1993). The south and middle Baikal basins are separated by the Selenga–Buguldeika sedimentary–tectonic ridge (Fig. 2). According to Logatchev (1993), the onset of southern Baikal basin formation is Palaeogene–Eocene, which makes it the oldest part of the rift.

The main structures of the basin have been reviewed in Levi et al. (1995, 1997). From the bathymetry, the southern Baikal basin can be subdivided into two depressions (Kultuk and Mishikhinsk) located west and east of the Angara Fault, respectively (Fig. 2). The lower part of the Cenozoic sediments are folded,
except within the Mishikhinsk depression, probably owing to left-lateral fault motions in the basement of the western part of the Kultuk basin. From multichannel seismic reflection data (Hutchinson et al. 1992), three first-order stratigraphic units have been identified: (1) proto-rift deposits of 4–5 km thickness with faults breaking the basement; (2) 1–2 km of middle-rift deposits which are characterized by numerous internal faults, generally steeply dipping; and (3) a modern-rift unit less than 500 m thick. The upper part of Unit 3 is unfaulted except near active faults at the steep edges of the lake. In all basins, depocentres are shifted to the northwestern side, i.e. towards the north basin-bounding faults.

Fault tectonics of the region has been studied mainly near the shorelines (Danilovich 1941; Sherman et al. 1973; San’kov et al. 1997; Chipizubov & Smekalin 1999; and others). The major faults are the Main Sayan Fault (MSF) striking ∼310° and the border faults of the lake. Left-lateral motions were deduced from structural and geomorphic field studies, aerial photographs, space imagery and seismicity (Sherman et al. 1973; San’kov et al. 1997; Chipizubov & Smekalin 1999). The onshore parts of the MSF are believed to control the tectonics of the southern Baikal basin and to accommodate a large part of recent motions. For instance, left lateral displacements on palaeoescarp have been evidenced at the junction of the MSF and Obruchovsky faults (Fig. 2). Late Pleistocene–Holocene cumulative offset is about 50 m. According to 14C dating, the last largest event produced about 4.0 m of horizontal displacement 400–1000 yr ago (Chipizubov & Smekalin 1999).

The northern side of the basin is bordered by the Obruchovsky Fault, which is assumed to control the formation and development of the whole Baikal rift basin (Logatchev & Zorin 1992). The slope towards the basin is very steep (50–70°), and the cumulative amplitude of dip-slip displacement is at least 5 km (Zorin 1971) and could be as great as 10 km (Ten Brink & Taylor 2002). At the southwestern tip of the basin, the fault splits into several normal en echelon fault steps (Levi et al. 1997), while on the opposite side, in the vicinity of the Selenga–Buguldeika ridge, its general strike gradually changes from 80° to 60°, forming the Primorsky Fault (Fig. 2). Although normal faulting seems to be dominant, this southern branch of the Obruchovsky Fault is characterized by a clear left-lateral strike-slip component. Faults of the Chersky system are limiting the southeastern side of the lake (Fig. 2). They form long, widely overlapping faults with dominant dip-slip faulting. They appear to accommodate much less motion than the northern faults, since they underline much smoother relief onshore and offshore. Although seismic exploration offshore is sparse, several other long segments (50–100 km) have been hypothesized within the basin and in the basement (Hutchinson et al. 1992; Colman et al. 1996; Levi et al. 1997, Fig. 2). Finally, the Angara Fault is a tectonic feature striking along the upper part of the Angara valley and crossing the southern Baikal craton. The Siberian craton, this fault depicts a wide zone of steeply dipping segments (Seminsky et al. 2001a) and is reactivated during the Cenozoic as a right-lateral strike-slip fault with a significant reverse component. Lacustrine ruptures are also widespread in the floor and the slopes of the basin (e.g. Colman et al. 1996; Levi et al. 1997; Scholz et al. 1998): for instance, the Buguldeika graben and associated saddle show much evidence of recent activity. In the samples obtained during the Baikal Drilling Project (Kuzmin et al. 2000, and references therein), ruptures at different scales have been also documented (Seminsky et al. 2001b).

### 3 Background Seismicity of the Southern Baikal Basin

In the central and southern basins of the Baikal rift, about 7000 events of magnitude larger than 2.7 have been recorded by the regional network between January 1950 and January 1999. Compared with the neighbouring regions of the Selenga Delta and central Lake Baikal, the south Baikal basin appears to be moderately active (Fig. 2). However, numerous active scarps reported near the coastlines and several historical, destructive earthquakes located here argue for a significant ongoing deformation (Solonenko et al. 1977; San’kov et al. 1999).

Epicentres within the Baikal depression depict several belts and clusters. Earthquakes located at the southern tip of the lake might be connected with the MSF since the epicentres appear to trend in the prolongation of this fault (Fig. 2). Further east, seismic activity becomes rather low and then increases to form two distinct northeast-striking bands of epicentres. These bands underline the most active fault segments inside Lake Baikal at present:

1. The western band (from 104.3°E to 106°E) is relatively narrow and might be under the control of the Obruchevsky Fault, at least between 105.2°E and 106°E, since it follows closely the western bank; note that at 105.2°E, the epicentres seems to change trend, from ENE at the west to northeast at the east, where it vanishes in the Buguldeika graben;

2. The eastern band (from 106.2°E to 107.5°E) is tentatively linked to the Chersky fault system owing to its closer position to the eastern bank of the lake. The most recent active segment known is the Delovy Fault, which crosses the Selenga Delta (Fig. 2). The 1862 Tsagan earthquake (M 7.1) could have occurred on this segment. It was followed by numerous aftershocks and was felt on a large area. The earthquake shaking caused the collapse of a large area, which has formed the Proval Bay (Solonenko et al. 1977). Two other large earthquakes have shaken the area, one in 1903 (M 6.7) and the other one on 1959 August 29 (M 6.8), which has been located 15–20 km north of Proval Bay. Solonenko (1964) reported that northeast-striking surface ruptures of the 1862 earthquake were further activated northeastwards during the 1959 M = 6.8 event. Its focal mechanism shows a pure normal motion on northeast planes.

Apart from these two bands, significant activity occurs inside the central part of the basin (Fig. 2), showing complication of the fault system, though these faults seem to be parallel to the rift sides. Note also that in this instrumental period, seismicity is weakly expressed along the Primorsky Fault, in spite of the outstanding morphological expression of the fault scarp. Seismic activity north of 53°N enlarges and seems to follow the Olkhon branch running through the Baikal basin (Fig. 2).

Few constraints are available concerning the depths of events in the Baikal rift. The 1959 earthquake offered the first opportunity to determine them: most aftershocks recorded by temporary stations clustered at 15–22 km depth. A study of the difference between arrival times of waves reflected from the Moho and direct waves has provided further evidence for depths ranging between 17 and 22 km, in good agreement with the depth estimation of the main shock found to be 18 ± 3 km (Golenetskii 1964). Moderate-size earthquakes which have occurred in this area during the past 50 yr were generally not followed by aftershocks (Golenetskii 1990; Déverchère et al. 1991). An earthquake swarm of roughly 100 events (Mmax 3.8) took place in 1973 southwest of the 1999 earthquake sequence and depicted typical strike-slip events.

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The 1999 sequence occurred at approximately the same distance from the lake sides and appears to trend westwards of the assumed prolongation of the Angara Fault inside the basin. Although the location of the epicentre of the main shock approximately coincides with an intrabasin fault (Levi et al. 1995, 1997) (Fig. 2, inset) deduced from multichannel seismic reflection data, its activity level and seismic potential are not yet obvious and the border faults have been supposed to play key role in basin development. Therefore, it is not clear whether the 1999 event is related to one of the main fault systems or to another fault located inside the south Lake Baikal basin. Ruzhich et al. (2002) have proposed to link the sequence to the intersection of the Obruchevsky Fault and Angara Fault. In their view, activation of this fault node resulted in motions on numerous fractures striking in different directions. In order to clarify this setting, it is necessary to carry out a detailed seismological study and to gather the available geophysical and geological data in this area.

4 THE 1999 EARTHQUAKE SEQUENCE DISTRIBUTION

4.1 Foreshocks, main shock and aftershocks: overview

Earthquakes were located by the regional seismic station network operated under the Baikal Experimental Methodical Seismic Expedition (BEMSE). By the beginning of 1999 the network was made of 23 analogue stations and five digital stations, plus one IRIS station (TLY). All regional stations are equipped with three short-period component seismometers. Two types of digital station (‘Baikal’ and ‘Delta-Geon’) record the ground velocity. For the former, the band-pass is 0.5–20 Hz and the sampling rate is 100 samples per second. ‘Delta-Geon’ stations belong to the Buriatian part of the Altay-Sayan Seismological Expedition and are maintained by the Buriatian Geological Institute (BGI). The time checking is performed by satellite control. We have estimated the accuracy in the readings of P arrival times at 0.2 s on paper records and at 0.1 s on digital records.

In the following, magnitude is estimated by using a linear relationship between the energetic class \( K \) used in Russia and the local magnitude \( M \). The conversion used is the one introduced by Rautian (1964):

\[
K = 4 + 1.8M. \tag{1}
\]

Focal mechanisms of the sequence were determined using first motion polarities of \( Pg \) and \( Pn \) phases. Fault plane solutions were constrained for 11 foreshocks and for the main shock and for four aftershocks. Focal mechanisms of aftershocks were difficult to determine because numerous events have occurred in a short time period, therefore first arrivals on the seismograms were often overlapping. In order to judge the quality of the solutions, first arrivals on the seismograms at regional stations and nodal plane scattering are shown in the Appendix.

From July 1998 until February 1999, the region under study did not suffer any significant earthquakes. Activity began on 1999 February 9 with a \( M 3.8 \) earthquake (Fig. 3) and was followed by six \( M \geq 1.1 \) events on that day and more than 50 on February 10, including two \( M 3.8 \) and three \( M 4.5 \) earthquakes. From February 9 until the main shock on February 25, 680 events with \( M \geq 1.6 \) were recorded by the permanent network. The epicentres of these first events stretched out in the NW direction.

On February 10, the focal mechanisms determined depict normal dip-slip motions on mainly northeast planes, except for one \( M 4.7 \) event for which the strike-slip component is significant. These ‘rift-type’ focal mechanisms were also occurring on February 11. From February 12 to February 24, seismic activity significantly decreased and became insignificant in the last 2 days. The main stage of activity started on February 25 with a \( M 4.5 \) earthquake at 05:37 UT, depicting pure normal faulting on approximately east–west planes, followed by another normal faulting event but along a northwest plane. At 18:56 UT, i.e. 2 min before the main shock, a \( M 3.3 \) event occurred. The 05:37 event may represent the initiation of the main rupture because its focal solution was nearly the same as the one from the main shock.

The source parameters of the main shock on February 25 at 18:58 UT are given in Table 1. In the prompt output from EMSC there is an error or misprint in latitude. As seen from Table 1, there is a rather good agreement between the focal solutions obtained by different sources. All solutions (either CMT or first motion solutions) show normal faulting on northeast-striking planes. One nodal plane is steeply dipping northwest and the second one is gently dipping southeast. The depth of the main shock has not been estimated very accurately owing to the lack of close permanent stations and the moderate magnitude. The centroid depth is 21 km, and the depth reported by EMSC based on two depth phases observed is 10 km. Anyway, there is no doubt that the nucleation of the rupture was in the crystalline basement.

Finally, the four focal solutions determined for the aftershocks depict a wide range of strikes of nodal planes and kinematic types (Figs 3 and 6). Nevertheless, normal faulting solutions with either northeast-southwest (‘rift-type’) or east–west directions of nodal planes prevail. Through time, the epicentral field has enlarged (Fig. 3). During March the activity decreased and after April it was becoming more and more scattered. In that period, only occasional earthquakes in the central part of the south Baikal basin occurred, striking northeast. In 2000 a \( M 5.2 \) earthquake took place there with pure normal faulting again along northeast nodal planes.

4.2 Aftershock relocation

Thanks to a dense regional network operated during March–May 1999, we present here a relocation procedure which has allowed for reliable depth estimations of the aftershock sequence. Location of the epicentres of all the earthquakes of the sequence was first performed by BEMSE. The epicentres were surrounded by stations at rather close distances in a large azimuthal range, therefore the accuracy of epicentral determination is generally better than 5 km. A few days after the main shock occurred (on March 1), an additional station (LST) was installed on the western side of Lake Baikal, at about 28 km from the main shock, which has brought an important additional constraint on locations. Besides that, the records of several stations of the BGI network were involved in the procedure of relocating the aftershocks (Table 2).

4.2.1 Data selection

Our aim was to select events which depict a reasonable and reliable depth location. For that purpose we have selected the aftershocks satisfying particular criteria: first of all, they should be recorded by the nearest stations (LST and/or ZBT), so that the smallest epicentral distance to them is less than 30 km. Unfortunately, these stations sometimes ceased operation due to difficult maintenance conditions in winter time. The other requirements were good time control at the stations and the existence of at least three stations with epicentral
distances less than 150 km as well as clear first arrivals on seismograms. Table 2 and Fig. 2 show the stations which were used for relocation.

Sixty-five events which occurred from March 1 to the end of May have been finally selected using these criteria and have been used as the initial data set for applying the relocation procedure and testing velocity models. The data used were the arrival times of $P_g$ and $S_g$ waves only, owing to the proximity of stations and to the complex geological structures (basin and rift shoulders) surrounding the sequence. It should be noted that the $S_g$ arrivals on the seismograms, especially at close stations, are quite distinct (see an example in Fig. 4). The importance of using the reliable $S$-wave arrival times for earthquakes location has been shown in many publications (Chatelain et al. 1980; Ellsworth & Roecker 1981; Urhammer 1982; Gomberg et al. 1990, and others). One of the main conclusions made in above-mentioned works is that including even a single $S$ phase
Table 2. Stations used in the relocation procedure.

<table>
<thead>
<tr>
<th>Station code</th>
<th>Latitude ('N)</th>
<th>Longitude ('E)</th>
<th>Elevation above sea level (m)</th>
<th>Type of station</th>
<th>Distance (km)</th>
<th>Azimuth (deg.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ZBT</td>
<td>51.55</td>
<td>105.12</td>
<td>450</td>
<td>Digital 'Delta-Geon'</td>
<td>25</td>
<td>100–125</td>
</tr>
<tr>
<td>LST</td>
<td>51.868</td>
<td>104.832</td>
<td>450</td>
<td>Digital 'Baikal-11'</td>
<td>25</td>
<td>350–10</td>
</tr>
<tr>
<td>IRK</td>
<td>52.243</td>
<td>104.271</td>
<td>467</td>
<td>Digital 'Baikal-10'</td>
<td>70</td>
<td>330</td>
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<tr>
<td>TLY</td>
<td>51.681</td>
<td>103.644</td>
<td>579</td>
<td>Digital 'IRIS'</td>
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<td>270</td>
</tr>
<tr>
<td>HRM</td>
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<td>106.955</td>
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<td>Digital 'Delta-Geon'</td>
<td>150</td>
<td>90</td>
</tr>
<tr>
<td>TIG</td>
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<td>106.348</td>
<td>718</td>
<td>Digital 'Baikal-11'</td>
<td>160</td>
<td>40</td>
</tr>
<tr>
<td>ARS</td>
<td>51.920</td>
<td>102.423</td>
<td>970</td>
<td>Analogue SKM-3</td>
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<td>280</td>
</tr>
<tr>
<td>ZAK</td>
<td>50.382</td>
<td>103.281</td>
<td>1200</td>
<td>Analogue SKM-3</td>
<td>180</td>
<td>215</td>
</tr>
<tr>
<td>UUD</td>
<td>51.867</td>
<td>107.663</td>
<td>600</td>
<td>Digital 'Delta-Geon'</td>
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<td>82</td>
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<tr>
<td>MOY</td>
<td>51.667</td>
<td>100.993</td>
<td>1303</td>
<td>Analogue SKM-3</td>
<td>260</td>
<td>270</td>
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</tr>
<tr>
<td>MXM</td>
<td>53.263</td>
<td>108.745</td>
<td>510</td>
<td>Digital station 'Delta-Geon'</td>
<td>320</td>
<td>55</td>
</tr>
</tbody>
</table>

‘Distance’ and ‘azimuth’ are the approximate epicentral distance and azimuth, respectively, from the epicentral field to a station.

Figure 4. An example of typical three-component digital seismograms recorded by ‘Baikal’ stations from the earthquake of the 1999 sequence occurring on March 6 at 00:54 UT, $H = 23$ km, $M = 4.4$. Two stations (HRM and LST) with different epicentral distances (D) and azimuths (Az) from the epicentre to the station are shown. $T_{P_g} - T_{S_g}$ is the time difference of the arrivals of $P$ and $S$ phases, accordingly.

recorded within a distance of 1–1.4 focal depths from the earthquake can significantly constrain focal depth and also reduce the semi-major axes of the error ellipsoid.

4.2.2 Location algorithms

For relocation we use the standard code HYPOINVERSE (Klein 1978) widely used for location of local or/and regional earthquakes. It is based on the principle of minimization of the root mean square (RMS) of the traveltimes residuals and uses the singular value decomposition method.

The input data are the time arrivals, station information, velocity model and parameters controlling the iteration process. Along with station and phase weighting, the program allows one to apply distance and residual weighting which follows a cosine taper between assigned values. $S$ phases were half-weighted relative to $P$ phases. Distance weighting in our computations began at 50 km. The purpose of the residual weighting is to reduce the weight of arrivals with large residuals. If the residual at one station exceeds a threshold value (we used 0.16 s), then the data are downweighted accordingly (for more details see Klein 2002). The output data are iteration output, the final location results and station list data. Here we note that location errors $e_{rh}$ and $e_{rz}$ are defined as the largest projection of the three principal errors on a horizontal plane and vertical line, respectively. These principal errors form the major axes of the error ellipse (standard deviation errors in km).

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4.2.3 Velocity models

Important information on crustal velocities of the region comes from seismic reflection/refraction experiments and also from the seismological observations over a long period of time. From the latter the average \(P\)- and \(S\)-wave velocity for the whole BRZ is 6.1 and 3.5 km s\(^{-1}\), respectively (Golenetsky 1964). Such a simple model of one layer over a half-space is used by BEMSE for routine proceedings. More detailed models for the Baikal basin are given in Krylov et al. (1981, 1991, 1993) in which the sedimentary layer in the southern basin is about 5–7 km thick (see Section 2.2). It is divided into three layers with sharp velocity discontinuities. Below them, a 3 km thick layer of pre-Cenozoic metamorphic sediments is assumed. Thus, the depth to the crystalline basement amounts to 9 km. \(P\)-wave velocity increases from 5.5 km s\(^{-1}\) at the basement roof up to 6.4 km s\(^{-1}\) at 20 km depth and deeper. At a depth of 12–17 km there is a low-velocity layer where \(P\)-wave and \(S\)-wave velocities decrease by 0.3–0.5 and 0.2–0.3 km s\(^{-1}\), respectively. The values of depth to the Moho boundary range from 35 to 40 km (Krylov et al. 1981; Suvorov et al. 2002; Ten Brink & Taylor 2002, and others) in the south Baikal basin. However, all the authors agree that this is the place where it reaches its minimum value in the rift system. We use here the value of 35 km following Zorin (1971), Zorin et al. (2002), Krylov et al. (1981) and Suvorov et al. (2002). According to these studies, \(V_p\) in the mantle is generally about 7.8 km s\(^{-1}\) below the basins and 8.2 km s\(^{-1}\) away from the rift axis.

In a preliminary work we found that the use of the sedimentary thickness divided into several layers gave large RMS values and increased the mean weighted station residuals for \(P\)- and \(S\)-wave arrivals times. Moreover, the depths of events tended to shift upwards. We believe that this is partly due to an inappropriate \(V_s\) value. Indeed, as shown for instance in Chiu et al. (1997), underestimating the shear wave velocity results in underestimating depth values, so that an independent \(V_s\) determination is required. The use of a low-velocity layer does not actually improve the results.

Therefore, we have chosen to apply a simpler velocity model to the 65 events selected, where the velocity of the upper layer is reduced at 5.9 km s\(^{-1}\). In this study, we merely try to test the influence of reasonable velocity changes on the location of hypocentres, and not to determine accurate velocities (which could be done only through a velocity inversion using a much higher ray density). For that purpose, three modifications of the simplest model were used (Table 3). The first model is a one-layer crustal model with the parameters \((V_p/\sqrt{V_s})\) found through some tests which had been carried out following the procedure described in Déverchère et al. (2001). The second model depicts a lower velocity in the upper part of the crust, whereas the third model is the more complex, with three crustal layers over a half-space.

Table 3. Crustal velocity models used in relocation.

<table>
<thead>
<tr>
<th>Model</th>
<th>Layer</th>
<th>Depth to top of layer (km)</th>
<th>(V_p) (km s(^{-1}))</th>
<th>(V_p/\sqrt{V_s})</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>0</td>
<td>6.1</td>
<td>1.71</td>
</tr>
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<td></td>
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<td>35</td>
<td>7.9</td>
<td>1.80</td>
</tr>
<tr>
<td>2</td>
<td>1</td>
<td>0</td>
<td>5.9</td>
<td>1.66</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>9</td>
<td>6.1</td>
<td>1.71</td>
</tr>
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4.2.4 Results of the relocation

As can be seen from Table 4, the mean values of depth, RMS and \(erh\) are nearly identical whatever the models used. Only the mean vertical errors \(erz\) slightly differ from one model to another. This behaviour is related to the shallowest earthquakes. Indeed, as expected, such events (0–5 km depth) depict the largest vertical errors. It is particularly clear when using the one-layer model 1. A comparison of our results and the studies performed elsewhere points out that these events may actually lie at rather shallow levels. For instance, in the Garm region, Central Asia, it was found that subsurface heterogeneities may result in extremely large uncertainties in the depth of shallow events (Rulev 1999).

The mean RMS value is less than 0.3 s whatever the model used. We have checked that the shifts between epicentres and depth values calculated using these models remain within the error ellipse computed for most events. These results show that applied velocity models do not significantly affect the location results.

Hypocentres are not clustering at the same depth but are spread from the surface down to 25 km depth (Fig. 5). However, they tend to get denser at the 15–20 km depth range. In order to make the link between faults and the relocated aftershocks we present two cross-sections (Fig. 5). The first one strikes around N70° E, i.e. parallel to the nodal plane of the main shock, and the other one follows a northwesterly direction. The hypocentre projection on the latter cross-section shows a northwest-dipping plane. In the northeast direction, the hypocentres are rather uniformly spread at various depths.

We have checked the consistency of the hypocentre depths obtained by BEMSE and our results. Although the fit is not perfect, more than half of the data set coincides within the confidence limits. Of course, the errors of BEMSE determinations are assumed to be larger, since our relocation was performed only for aftershocks occurring since March 1 and in the case of the best recording conditions. This relatively good agreement has led us to use hypocentre depths from the BEMSE bulletins for further events of the sequence. Here we are not considering the rheological meaning of the hypocentre depth values determined in this study. Nevertheless, we may note that depths found match those computed in different areas of the Baikal rift zone, which are dominant in the depth range 15–25 km (Déverchère et al. 1991, 1993, 2001).

5 STRESS FIELD EVOLUTION

The south Baikal basin is located at the transition from an area of pure extension typical of the central part of the Baikal rift system, to an area of transpressional deformation, similar to the one found in northern Mongolia (Misharina 1972; Petit et al. 1996; Solonenko et al. 1997; Delouis et al. 2002). Several numerical modelling attempts (e.g. Lesne et al. 1998; Polyansky 2002) show that the transition of strain and stress regimes is very fast in this part of the rift, implying a coeval northeast-southwest compression and

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Figure 5. Depth histogram and cross-sections of the relocated aftershocks obtained using model 1.

northwest–southeast extension. Therefore, ‘the rift’ character of the tectonic stress field found in the central and south Baikal basins seems to vanish close to the place where the 1999 event occurred. Previous inversions of stress directions deduced from focal solutions in the southern Baikal basin (e.g. Petit et al. 1996; Delouis et al. 2002) suggest a westward increase of the strike-slip component along approximately east–west to east–southeast directions, which appears to be progressive at a regional scale. A similar evolution is also suggested by Quaternary fault-slip data gathered near the coastlines of the southern basin (San’kov et al. 1997; Delvaux et al. 1997) and by fold axes striking northwest–southeast imaged in the youngest sedimentary units of the southern basin of the lake (Levi et al. 1997). In this study, we gather all individual and composite focal solutions since 1960 (Fig. 6) and confirm this general spatial evolution. In the western part of this basin, only one earthquake (1993, $M_{3.4}$), located importantly close to the epicentral area of the 1999 sequence, depicts normal faulting. All other solutions show dominant strike-slip motions along east–west and north–south nodal planes. Additionally, a composite focal solution derived from 21 small earthquakes (Fig. 6) also shows the same behaviour. Actual planes are most likely the planes parallel to the strike of the Obruchevsky Fault (Fig. 2), which controls basin development and acts as an oblique normal fault. Towards the northeast, normal faulting along northeast planes predominates, in agreement with the strike of border faults of Lake Baikal.

Based on this distribution of focal mechanisms, we divide the south Baikal basin into two subregions (Fig. 6) and carry out stress inversions following the procedure by Delvaux et al. (1995), using both individual and composite solutions. The main assumptions for carrying the inversion are classically that (1) stress is uniform in the volume of rocks considered and (2) slip occurs in the direction of the resolved shear stress on the fault plane. The stress regime computed for the part of the basin located west of the Angara Fault without the 1999 events (box I, 26 solutions, Fig. 6) is transtensive (using the terms from Delvaux et al. 1997), since shape ratio $\phi = 0.82$, where $\phi$ is $(\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)$. The $\sigma_1$-axis is vertical, and the minimum principal stress axis $\sigma_3$ is horizontal and strikes N139°E. The eastern part of the basin (box II, 81 solutions, Fig. 6) is characterized by a pure extensive stress regime with the same strike for horizontal $\sigma_3$ (N139°E) and a $\phi$ value of 0.56. Therefore, in spite of the slight changes in the stress regime, the direction of the extensional axis is stable throughout the whole southern Baikal basin, if we consider the last $\sim$40 yr of recorded events. This result reveals what can be called a ‘regional’ stress direction but questions the existence of a ‘regional’ stress field. Finally, a stress inversion based on the 16 focal mechanisms from the 1999 sequence (box III, Fig. 6) shows a nearly radial extensive stress regime with $\phi = 0.19$ and a $\sigma_3$-axis striking about N155°. No obvious change in stress direction seems to occur during the 1999 sequence according to the subset of fault plane solutions determined.

6 DISCUSSION

6.1 Faulting pattern

Although the aftershocks are spread over a large area (Fig. 3), cross-sections using the best relocated aftershocks (Fig. 5) and the focal solutions suggest that the main rupture of this sequence occurred on a steeply northwest-dipping plane of northeastern strike inside the basin. We may assume that this sequence is not directly related to the main border Obruchevsky Fault for the following reasons. First, the steeply dipping nodal plane of the main shock, which is likely to be the fault plane according to relocated aftershocks, dips towards the northwest and not to the southeast as the Obruchevsky Fault does. Second, the kinematics of both active faults appear to be different: whereas the Obruchevsky Fault depicts a significant strike-slip component, and although this part of the basin is under an extensive regime of strike-slip type (Fig. 6), the 1999 earthquake depicts mostly pure normal faulting. Finally, the epicentral field of the 1999 events is located at rather a remote distance from the

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Figure 6. (a) Fault-plane solutions (lower hemisphere projection, Schmidt net) for earthquakes of the south Baikal depression determined in the 1960–1999 time span. Dashed lines outline the epicentral areas of composite focal solutions (grey beach-balls). (b) Focal solutions of the 1999 sequence. (c) Orientation of principal stress axes deduced from inversion of fault plane solutions in the southern tip of the basin (box I), further east (box II) and by using the 1999 sequence (box III). Arrows are labelled, following the convention of Delvaux et al. (1997), considering the shape ratio of the stress tensor as $\phi = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)$. See text for details.

master Obruchevsky Fault which does not show any flattening at depth (Hutchinson et al. 1992; Levi et al. 1995, 1997): indeed, the Obruchevsky Fault separating the Siberian craton from the mobile fold belt is a deeply penetrating fault inherited from Palaeozoic times (Zorin 1999). Furthermore, the background seismicity is relatively sparse here (Fig. 2), suggesting the existence of other active faults. It should be noted that several focal solutions of earthquakes prior to 1999 also show similar steeply northwest-dipping nodal planes (Fig. 6).

Because the 1999 event and aftershocks occurred in an area of relatively sparse pre-main shock seismicity, also characterized by apparent earthquake clustering (Fig. 2), and because of the remaining relocation errors, it is difficult to assess the exact trends and extents of the faults responsible for the pre-shock and aftershock activity. If we use the scaling relations for aftershocks of dip-slip earthquakes (e.g. Henry & Das 2001), the approximate 1-day length of the February 25 earthquake and aftershocks should be about 15 km. The relatively large size of the aftershock area compared with this assumed main rupture length (Fig. 3), together with the numerous fault plane orientations observed in the 1999 focal solutions of aftershocks (Fig. 6) and with the various trends apparently revealed by the clusters of aftershocks at different time periods (northwest, and later northeast and north–south; see Fig. 3), favour a possible activation of several faults surrounding the main rupture zone by stress triggering. This could also explain the change in local stress state discussed above. Aftershock complexity is often documented after earthquakes of magnitude 5.5 and more and illuminates various fault directions that are understood as main shock-induced stress failures, possibly related to fault intersections (e.g. Ichinose et al. 1998) or to the growth of young faults (see e.g. Kilb & Rubin 2002). It is beyond the scope of this paper to determine the exact causes of this aftershock behaviour, because the resolution on relative location that we may achieve is not good enough. We may note that in this peculiar case the kinematics of the rupture sequence remains roughly consistent with (i.e. linked to) the long-term stress field. Indeed, on northeast-striking planes, i.e. the main direction of rupture, normal faulting prevails, whereas northwest planes are characterized by strike-slip and sometimes reverse faulting.

An important conclusion is that the development of the south Baikal basin occurs not only through strain localization on the MSF and on the main border faults but also on intrabasin ones, within the basement. These faults could represent a splay network forming a horsetail-type structure and resulting from the progressive development of the southern basin. This mode of opening implies a strain transfer from the localized shear zone of the MSF to a distributed normal fault system extending from the Obruchevsky Fault to the deep basin up to the Chersky Fault (Fig. 2). Additionally, our results suggest that neither the Obruchevsky Fault to the north nor the Chersky Fault to the south are likely to accommodate the whole relative motion between the Siberian platform and the Khamar Daban terrane: instead, inherited structures from the basement of the south Baikal basin suitably oriented in the present-day stress field appear to play a significant role in present-day basin opening.
6.2 Stress field

From the above analysis we note that the stress release associated to the 1999 sequence is slightly different from the stress regime determined previously in the area concerning stress type and stress direction. Indeed, the stress ratio reveals a radial type of extensive stress, whereas the ‘background’ stress regime usually depicts a much stronger intermediate (σ2)-axis, and the direction of extension (σ1)-axis is ~16° closer to the north (Fig. 6). Nevertheless, the stress tensor type remains extensive, and most solutions are compatible with a nearly uniform (northwest–southeast) minimum stress direction. Some speculation may be proposed based on these results of the stress field evolution. Following the two end-member models discussed by Rivera & Kanamori (2002), the spatial and/or temporal heterogeneity revealed by the 1999 sequence may reflect either (1) a variability of the friction on the various fault planes (which is implicitly the explanation to consider when applying a stress inversion procedure) or (2) a local, possibly transient, stress change, which requires us to rule out the assumption of uniform stress field. Stress perturbations following a large event are common and may change with time within the aftershock sequence (e.g. Slejko et al. 1999). They are often understood as a stress heterogeneity that arises from local responses to the seismic activity (large rupture) superimposed on the regional stress (Slejko et al. 1999), or from fault interactions (Petersen et al. 1991; Gupta & Scholz 2000). We may therefore assume that the 1999 main shock has induced a local, probably transient, stress perturbation in the nearby faulted rock volume.

6.3 Strain rate

The strain axes calculated from GPS measurements in the south Baikal basin (Loukhnev et al. 2003) show a strike-slip regime with similar lengths of maximum and minimum principal strain rate axes. The current rate of opening in this area is about 4 ± 1 mm yr\(^{-1}\) in a N140°E direction (Calais et al. 1998, 2002, 2003), a value in a good agreement with stress directions deduced from seismological data. Such a movement applied at southern Lake Baikal results in an extension obliquity of about 50–60° relative to the border faults (e.g. Petit et al. 1996). The same pattern is obtained in the best-fit model proposed by Polyanovsky (2002) in an attempt to reproduce GPS observations, focal mechanisms, fault-slip data and Holocene fault-slip rates. Therefore, a significant strike-slip component is expected on these border fault segments and on subparallel faults inside the basin, which should increase from east to west according to the shape of the southern tip of Lake Baikal (Fig. 2) and to modelling results of strain (Lesne et al. 1998; Polyanovsky 2002). In order to estimate the strain caused by the main shock of February 25, we have used its seismic moment tensor, given by (Gilbert 1970):

\[ M_{0ij} = M_0(n_i b_j + n_j b_i) \]

where \( n \) and \( b \) are unit vectors normal to the fault plane and parallel to the slip vector, respectively, and \( M_0 \) is the scalar moment. The seismic moment tensor was divided by twice the product of the shear modulus and the deforming volume, following Kostrov (1974). Note that we use here only the energy released during the main shock, since no aftershocks were strong enough to significantly increase the total seismic moment. The scalar moment \( M_0 \) from the Harvard CMT solution is 8.91 × 10\(^{17}\) Nm. Given a radius of the source of 15 km and a seismogenic thickness of 20 km, strain rate is estimated to be 2.2 × 10\(^{-8}\) yr\(^{-1}\), whereas the average strain rate for the Baikal rift zone from GPS measurements is of the order of 10\(^{-8}\) yr\(^{-1}\) (Fig. 6; Loukhnev et al. 2003; Calais et al. 2003). Furthermore, a map of the maximum horizontal shear strains based on GPS data (Fig. 7; Loukhnev et al. 2003) shows that the area of the 1999 sequence falls within the high-value domain. Therefore, an event like the 1999 main shock significantly contributes to strain, and the maximum shear strain estimated for this area is mainly linked to this event.

6.4 Seismic hazards

The 1999 \( M_w = 6.0 \) event shows tectonic activity inside the basin which is unexpectedly high. This raises the question of whether the intrabasin faults can produce large earthquakes and what are their seismic potential and recurrence intervals. Mean slip rates of 5 ± 2 and 2 ± 1 mm yr\(^{-1}\) have been proposed for the MSF and Tunka–Obruchevsky faults, respectively, from palaeoseismological (San’kov et al. 1997; Chipizubov & Smekalin 1999) and from GPS or numerical modelling (Polyanovsky 2002; Calais et al. 2003).

The occurrence of the 1999 earthquake does not only indicate that border faults of the rift basin may give rise to large earthquakes but also has implications for the recurrence intervals on each fault that can be assessed considering the rate of opening of the Baikal rift. For instance, the 350 yr interval assessed for magnitude 7.5 events on the Obruchevsky Fault (Calais et al. 1998) could be larger, its mean slip rate smaller, and the mean slip rates on other main border faults onshore assumed by Polyanovsky (2002) should be considered with caution. Secondly, the high level of historical seismicity of the southern basin is likely to be linked to relatively high strain rates at the place where the large strike-slip zone of the MSF turns progressively to a series of dip-slip normal faults within the southern Baikal basin (Figs 2 and 7). Therefore, the 1999 February 25 event provides further evidence for an ongoing stress drop through active normal faulting at the southern tip of the main rift basin, while the MSF further west is probably undergoing interseismic shear strain accumulation, since no significant relative motion (at the 1 mm yr\(^{-1}\) level) has been revealed between GPS sites located on each side of the southern tip of the MSF (Calais et al. 1998, 2003).
Considering the relative high strain localization occurring on the MSF, its estimated slip rate of about 5 mm yr\(^{-1}\), and the time (about 2000 yr) elapsed since the last large event on this fault (Chipizubov & Smekalin 1999), particular care must be taken for this fault system in terms of seismic hazards.

7 Conclusions

The 1999 Baikal earthquake sequence appears to be related to active faults located in the basement of the southern basin of Lake Baikal. Indeed, the \(M_w = 6.0\) main shock of February 25 depicts normal faulting along a northeast-striking plane steeply dipping towards the northwest, which is not related to the border faults of the rift. Furthermore, relocation of the aftershocks shows that the hypocentres are not clustered but spread from 5 to 25 km depth, with a maximum in the depth range 15–20 km. The distribution of foreshocks, aftershocks and focal solutions favours a zone of stress perturbation encompassing some volume surrounding the main rupture. The stress regime during this earthquake sequence is radially extensive, i.e. significantly more extensive than the background stress assumed from previous seismological and geodetic data. This earthquake shows that the south Baikal basin develops not only owing to movements on the main border faults but also on intrabasin ones, a fact that modifies previous estimates of seismic potential and slip rates on main faults known onshore. The high level of seismicity of the southern basin for two centuries is likely to reveal relatively high strain rates at the place where strike-slip motion turns progressively to dip-slip motions in the deep basin, and the present-day high strain deduced from GPS measurements in the epicentral area of the 1999 event provides further support for a significant interseismic strain accumulation on the strike-slip MSF further west. In the future, a more detailed monitoring of the seismic pattern of the MSF and the southern lake are necessary to provide clearer kinematic and dynamic patterns of this area. In particular, denser seismological networks on land and in the deep lake will certainly be a key point in order to improve our knowledge of background and aftershock distribution.

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References


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Figure A1. View of focal mechanisms determined with first motions on the focal sphere (Schmidt net, lower hemisphere projections). Solid lines represent the nodal plane projections considered as fault plane solutions; dashed lines are possible uncertainties. Closed circles represent compression at the stations and open ones represent dilation. Small circles in the focal solution of the main event (1999 February 25, 15:58, $M = 5.9$) are the data from distant stations.