Depth distribution of earthquakes in the Baikal rift system and its implications for the rheology of the lithosphere

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SUMMARY

The correspondence between the predicted brittle–plastic transition within the crust and the maximum depth of earthquakes is examined in the case of the Baikal rift, Siberia. Although little accurate information on depths is available through large- and moderate-size earthquakes, there are frequent indications of foci at 20 km depth and more. We have relocated 632 events recorded at nearby stations that occurred between 1971 and 1997, with depth and epicentral uncertainties less than 5 km, over the eastern and southern parts of the Baikal rift. We have compared these results with other depth distributions obtained in previous studies from background seismicity in the NE rift (1365 events in the Kalar-Chara zone and 704 events in the Muya region). The relative abundance of earthquakes is generally low at depths between 0 and 10 km (7–15 per cent) and high between 15 and 25 km (≈ 50 per cent). Earthquake activity is still significant between 25 and 30 km (9–15 per cent) and persists between 30 and 40 km (7–13 per cent). Very few earthquakes are below the Moho. We use empirical constitutive laws to obtain the yield-stress limits of several layers made of dominant lithologies and to examine whether the observed distribution of earthquakes at depth (519 events controlled by a close station and located within the extensional domain of the Baikal rift system) can match the predicted crustal strength proportion with depth and the deeper brittle–ductile transition in the crust. A good fit is obtained by using a quartz rheology at 0–10 km depth and a diabase rheology at 10–45 km depth with a moderate temperature field which corresponds to a ~100 Myr thermal lithosphere. No dioritic composition of the crust is found necessary. In any case, earthquakes occur at deep crustal levels, where the crust is supposed to be ductile, in a way very similar to what is found in the East African Rift System. From these results we conclude that the seismogenic thickness is ~35–40 km in the Baikal rift system and that the depth distribution of earthquakes is at first order proportional to the strength profile found in a rheologically layered crust dominated by a mafic composition in the ~10–45 km depth range. An upper mantle core with high strength, however, generally prevents it from reaching stress failure at greater depth.

Key words: Asia, Baikal Rift, continental crust, extension, rheology, strength profile.

INTRODUCTION

About 25 years ago, simple models of the rheology of the continental lithosphere (Sibson 1977; Brace & Kohlstedt 1980; Kirby 1980) led to the conclusion that there was a rough correspondence between the predicted brittle–plastic (or brittle–ductile) transition within the crust and the maximum depth of earthquakes (Sibson 1982; Meissner & Strehlau 1982; Ranalli & Murphy 1987; Scholz 1988). Since that time, substantial progress has been made in understanding the importance of the regime of semi-brittle behaviour and the stability of sliding in the crust (Kohlstedt et al. 1995 and references therein). Although the onset of ductile deformation (which is at the transition to the stable frictional regime) is often considered as defining the base of the seismogenic zone (Scholz 1990), the transition from semi-brittle deformation to plastic flow is likely to mark the maximum rupture depth of large earthquakes, and coincides with the base of the schizosphere, where shear strain energy is the highest (Sibson 1982; Das & Scholz 1983; Scholz 1990).
On the other hand, several factors have been recognized that significantly modify the depth–frequency distribution of earthquakes inside continents. These include thermal state and strain rates (Chen & Molnar 1983; Smith & Bruhn 1984; Doser & Kanamori 1986), dry or wet rheologies for various rock compositions (Brace & Kohlstedt 1980; Kohlstedt et al. 1995), and development (‘maturity’) of fault zones and style of faulting (Marone & Scholz 1988). In nature, the relative importance of these parameters is highly dependent on the geological and tectonic context. Nevertheless, laboratory studies and analytical or numerical simulations have shown that the thermal state and mineral composition are a first-order control on the strength of the lithosphere, and hence on the depth–frequency distribution of earthquakes (e.g. Chen & Molnar 1983; Burov & Diament 1995; Cloetingh & Burov 1996).

A comparison of strength envelopes with observed depth distributions of earthquakes may be quite useful in a given geological context, since it may improve our understanding of the rheology of the lower crust and help to discriminate the dominant parameters determining the structure and dynamic behaviour of faults at depth. Of particular interest is this approach in the vicinity of rift zones, as several results point to the occurrence of events at depths well below 20 km (e.g. Shudofsky 1985; Shudofsky et al. 1987; Déverchère et al. 1991, 1993; Nyblade & Langston 1995), i.e. within the lower crust, whereas conventional curves of yield strength versus depth predict a much thinner and shallower seismogenic zone, typically restricted to the upper 15–20 km of the crust (Meissner & Strehlau 1982; Sibson 1982; Chen & Molnar 1983; Jackson & White 1989; Scholz 1990). This deep activity is even more surprising if we consider that the temperature in rift zones is expected to be higher than that found in intracontinental areas, since the lithosphere and/or the crust is thinner, and hence should lead to a shallower brittle–ductile transition (Chen & Molnar 1983; Fadaie & Ranalli 1990).

Figure 1. Seismicity of the Baikal rift system recorded by the regional network (about 35 stations) from 1960 January to 1997 June (110 728 events). Triangles denote seismic stations, permanent (large) or temporary (small).

Among present-day rifts, the East African Rift System (EARS) is probably the one that has recently received the most attention in this respect. Unfortunately, well-determined source depths are limited in the EARS, since accurate determinations from waveform analyses are restricted to a data set of ~50 events with magnitudes $M_s$ above 5.1 (Shudofsky et al. 1987; Nyblade & Langston 1995; Zhao et al. 1997; Foster & Jackson 1998). This is mostly due to the scarcity of permanent stations in Africa. Nevertheless, this data set, together with a few microseismicity surveys (e.g. Maasha 1975; Camelbeek & Iranga 1996), undoubtedly demonstrates a seismogenic thickness of up to ~35 km in the EARS, which appears particularly clear in the western branch and at the southern end of the EARS (Foster & Jackson 1998). According to Nyblade & Langston (1995) and Foster & Jackson (1998), this requires a dry, mafic composition of the lower crust, given the range of temperature of 325–475 °C predicted at this depth. Whether or not a similar behaviour occurs in other rifts, and specifically in one of the largest rifts on Earth, namely the Baikal rift system (BRS), is not as clear yet, perhaps because large- and moderate-magnitude earthquakes have been comparatively less numerous there since the 1960s (Doser 1991a,b). However, strong shocks with source depths of ~30 km reported in the BRS (Golenetsky 1964; Golenetsky 1990; Doser 1991a,b; Doser & Yarwood 1994) suggest an additional resemblance between the two rift systems, besides several geometrical and mechanical similarities (Petit & Ebinger 2000, and references therein). On the other hand, continuous recording of earthquakes has taken place in the BRS for the last 40 years, providing a significant database of more than 110 000 events (Fig. 1). A small part of this data set has already been relocated (Déverchère et al. 1991, 1993) and indicates, again, the occurrence of events at depths as great as 30 km, and sometimes deeper. Until now, no systematic and quantitative analysis of earthquake depth distribution from this large data set has been performed.
In this study we first aim at precisely estimating the seismogenic thickness in and around the Baikal rift, and at quantifying its possible spatial variations. To achieve this, we review previous available information on focal depths of large- and small-magnitude earthquakes in the BRS, and perform a systematic examination of the depth distribution of events by relocating a set of 632 earthquakes recorded by local and regional networks in the time span 1971–1997. We then attempt to deduce from depth distributions of events some constraints on the yield-stress envelopes (YSEs) of the lithosphere that we can reasonably model using the observed heat flow at the surface and the assumed bulk composition and rheology of the crust. In particular, we aim at examining the rheology of the lower crust and its ability to behave as a seismogenic layer. This should help us to infer a realistic thermomechanical model of the lithosphere under extensional stress and allow some comparisons with the rheology of the EARS and other continental interiors, taking into account their respective geological frameworks.

DEPTHs OF EARTHQuakes PREVIOUSLY REPORTED IN THE BAIKAL rift

Large- and moderate-size earthquakes and aftershocks

Little is known about the exact depth distribution of the large earthquakes that have occurred in the BRS during the 20th century. Golenetsky (1964) made a first attempt to relocate the earthquakes that have occurred in the BRS during the 20th century. Golenetsky (1964) made a first attempt to relocate the earthquakes that have occurred in the BRS during the 20th century. Golenetsky (1964) made a first attempt to relocate the earthquakes that have occurred in the BRS during the 20th century.

Table 1. Epicentral parameters of events determined using body-waveform inversions and Harvard CMT best double-couple mechanisms selected in the Baikal rift. $h$ is either the centroid depth (CMT solutions) or the depth obtained (with computed uncertainties) in the multistep inversion technique of Baker & Doser (1988). D91a,b is for Doser (1991a,b), respectively, M for Misharina (1972), and CMT for the Harvard CMT database 1977–2000. For Event 2 we chose the fault-plane solution from Misharina (1972) instead of the body-waveform solution, and for Event 10 the solution modelled by Doser (1991b) instead of the CMT solution. See Fig. 2 for the locations and plot of fault-plane solutions.
Earthquakes in the Baikal rift system

Small earthquakes (background seismicity and swarms)
Several attempts have been made to locate the depths of small earthquakes accurately in the 
BRS. These efforts mostly focused on the youngest part of the Baikal rift (i.e. probably
younger than 10 Ma), which extends for more than 800 km
from northern Lake Baikal to the easternmost basin (Déverchère
et al. 1993; San’kov et al. 2000). Indeed, the regional permanent
network here is denser (Fig. 1), and several local seismic
stations related to the construction of the Baikal–Amur railway
line have been added and maintained there for several years
(Golenetsky 1985). Most studies use the well-known rule that
the distance to the nearest station, together with the $V_p/V_S$
ratio, are critical to constrain depth (e.g. Toth & Kisslinger
(1981) has checked this point from experimental data in the
BRS and has consequently inferred that the accuracy expected
on the depth $H$ of an earthquake is

$$\delta H = \delta E(\Delta_{\text{east}}/H),$$

where $\delta E$ is the accuracy of the location epicentre, and $\Delta_{\text{east}}$
the distance to the closest station. She has extracted several
hundreds of earthquakes from background seismicity and from
several dense and long-lasting swarms that occurred in this area
(Déverchère et al. 1991), considering only events with $\Delta_{\text{east}}$
less than 40 km and computed $\delta E$ less than 3.5 km. She has
detected systematic differences of arrival times at several close
stations and has obtained depth-change estimates by grouping
events with the same epicentral positions within error bars.
The error $\delta H$ for most selected events is estimated to be less
than $\sim 5$ km according to eq. (1). Swarm activity provides

contrasting results, with depths in the range 12–30 km or some-
times 2–22 km, whereas background seismicity is distributed
over 0–40 km. The region of Kalar-Chara, eastern Baikal rift
(Lat. 56°–57° N, Long. 116°–118° E) is especially attractive,
since the seismicity has been recorded continuously from 1962
to 1981 by at least three close stations (Fig. 1), and has shown
periods of intense activity separated by seismic quiescences
of 1.5–3 years (Vertlib 1997). The depth distribution from the
background seismicity (1365 events recorded from 1963 to
1975, Fig. 3a) shows a relative activity decrease at $\sim 4$–12 km
(by 10 per cent of events) and an increase at 15–25 km depth
($\sim 45$ per cent). Events are common at 25–30 km during the
whole time period ($\sim 15$ per cent) and often occur between 30
and 40 km ($\sim 13$ per cent). Deeper events are very few but do
nevertheless occur: 23 and 11 earthquakes are found at depths
of 40–50 km and 50–60 km (i.e. undoubtedly in the upper
mantle), respectively, during the 20 years of recording.

Other microseismicity studies in the Baikal rift are focused
in the region of the Muya basin, eastern Baikal rift, where a local
network of six stations was operated continuously between
1977 and 1988 (Solonenko et al. 1985; San’kov et al. 1991; Déverchère et al. 1993). More than 12 000 events of all magni-
tudes have been located there, but there are a large number of
clusters at various depths (5, 10, 24 km). Golenetsky (1985) and
Anikanova & Borovik (1981, 1986) have used part of this data
to determine accurate velocity models and $V_p/V_S$ ratios.

From various subsets of selected events, they have found depth
distributions where $\sim 80$ per cent of events lie between 10
and 20 km. A significant number of events ($\sim 15$ per cent) is
found between 20 and 30 km. Anikanova & Borovik (1986)
excluded from this data set all events related to aftershocks and
swarms. Using the 650 remaining events, they found a Gaussian

![Figure 2](image-url)
distribution for events between 4 and 24 km, with 10.2, 83.5, and 6.3 per cent of events with depths $H < 10$ km, $10 \leq H < 22$ km, and $H \geq 23$ km, respectively. Finally, part of the whole data set of the Muya region (704 best-recorded events) has been further relocated using station residuals and a master event technique by Déverchère et al. (1993) to obtain a constraint on focal depths of $\pm 2$ km within the network. In the time window selected, several dense swarms occurred (in particular the Angarakan one, which contained $\sim 60$ per cent of the events occurring near 10 km), and therefore the histogram of computed depths (Fig. 3b) is strongly biased towards shallow values in this case. Nevertheless, this study has demonstrated that steady-state seismic activity occurred along large, planar faults which are seismogenic at depths down to 28–30 km.

Figure 3. Histograms showing the best-determined focal depth distributions of small earthquakes in the Baikal rift from previous studies (Vertlib 1981, 1997; Déverchère et al. 1993). (a) Plot of 1365 events relocated in the region west of the Chara basin (Kalar-Chara), eastern Baikal rift (Latitudes 56°—57° N, Longitudes 116°—118° E) in the time span 1963–1975. The relocation method is from Vertlib (1981). No dense swarms (clusters of several hundred events of low or moderate magnitudes and lasting several months or years—see Déverchère et al. 1991) occurred during this period, and therefore this seismicity data set is considered as representative of the background activity in the area. Three stations are located at less than 40 km from most relocated events, and form a nearly equilateral triangle, constraining the depth at $\pm 5$ km. Additional stations were added during some periods of time. (b) Plot of 704 relocated events from the north Muya region (Latitudes 55.5°—56.5° N, Longitudes 112.5°—114.5° E) in the time span 1977–1980. Events have been relocated by a master event technique (Déverchère et al. 1993) in the region where six local stations constrained depths at $\pm 2$ km. About 60 per cent of this activity comes from the dense swarm ($10 \times 10$ km) of Angarakan, with most depths spread between 5 and 13 km.
RELOCATION OF EARTHQUAKES CLOSE TO SEISMIC STATIONS

In order to obtain a more homogeneous and complete view of the depth distribution of earthquakes in the BRS, we performed a systematic relocation of events recorded by the standard short-period, three-component permanent network, beginning with the best time period relative to the number of stations (from 1971 until 1997; see Table 2). First we searched for $M_l > 2$ events that are at a distance of less than 40 km from any given station of the permanent seismic network, following the general relationship between the earthquake depth and the distance to the closest station (eq. 1). Only nine stations (Fig. 4, Table 2) appear to have recorded enough close events with a reasonable azimuthal coverage (better than 160°), and allows us to select 5274 events. From this original data set, we keep events with at least two additional stations at less than 150 km, and with clear arrivals of both direct $Pg$ and $Sg$ waves at five stations or more, in order to obtain at least three $Pg$ wave times as first arrivals, thus avoiding instabilities linked with the less accurate $Pn$ arrival times (see, for example, Deichmann 1987). The magnitude range of events selected is ~2.2–4.5 (Russian energy-class scale $K$ between 8.0 and 12.0, converted using the formula from Rautian 1964). We base our location procedure on the computer program HYPVERSE (Klein 1978) adapted for station elevation corrections. We use a distance weighting scheme which is maximum from 0 to 50 km, and a cosine taper down to zero at 800 km. A similar weighting scheme is applied for station elevation corrections. We use a distance weighting formula from Rautian 1964). We base our location procedure entirely on the computer program HYPVERSE (Klein 1978) adapted for station elevation corrections.

Velocity model

Seismic refraction and deep-seismic sounding, combined with seismological studies carried out in the Baikal rift, have provided a good idea of the main structures of the crust (Putyrev et al. 1978; Krylov et al., 1981, 1991). The western tip of the Baikal rift (stations MON, ARS, Fig. 2) has a Moho at depths of 48–49 km, whereas the Moho ranges from 40 to 45 km below the six other stations used in this study. Similar Moho depths have been estimated from independent gravity modeling (Zorin et al. 1989; Burov et al., 1994; Petit et al., 1997). These Moho depths are in agreement with the critical distance of $Pn$ arrivals, which is found to be about ~180 km. Velocities in the crust have been determined using $P$-wave hodographs at various depths (Anikanova & Borovik 1986) or refraction data. The $V_P$ values continuously increase from 5.8 to 6.3 km s\(^{-1}\) down to 20 km, with a velocity reduction sometimes seen at intermediate levels (a decrease of ~0.2 km s\(^{-1}\) at 12–17 km below the basins, and ~5 km deeper below the ranges). At greater depths the mean $V_P$ value is 6.4 km s\(^{-1}\) down to 40 km. In the mantle, $V_P$ is generally ~7.8 km s\(^{-1}\) below the basins and ~8.2 km s\(^{-1}\) away from the rift axis. Considering that: (1) clear arrivals on seismograms are found for $Pg$, even beyond the critical distance of ~180–200 km; (2) no impedance contrast inside the crust is strong enough to allow us to detect reflected or refracted waves, or even enough clear reflections from the Moho (Déverchère et al., 1993); and (3) $Pn$ and $Sn$ arrivals are heavily down-weighted compared with the $Pg$ and $Sg$ arrivals in the distance and residual weighting schemes chosen, we have relied only on the latter phases for relocation.

Table 2. Results of the step-by-step relocation of $M_l > 2$ events in the BRS. The first column gives the total number of events kept at each step of the relocation procedure and the mean values of some parameters after steps 4, 5 and 6. The nine following columns provide the results for the stations selected, shown by the three-letter station code (see Fig. 4 for locations). See text for a description of the steps. $N$ is the initial number of $M_l > 2$ events found to be less than 40 km from the selected stations; $h$ is the computed focal depth in kilometres, $\epsilon r_c$ the vertical error computed in kilometres, $H$ the average arithmetic depth in kilometres; $\sigma$ is the standard deviation on the parameters analysed, and $V_P$ and $V_S$ are $P$- and $S$-wave velocities, respectively, in km s\(^{-1}\). Mean heat flow in mW m\(^{-2}\) (after Lysak 1984, 1992) and time windows for the selection of earthquakes are shown in the two last lines.

<table>
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<th></th>
<th>MOY</th>
<th>ARS</th>
<th>TAL</th>
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Therefore we use a simple half-space model with an initial $V_P$ value of 6.15 km s$^{-1}$, in agreement with most mean values proposed in similar studies (Golenetsky 1985, 1990; Anikanova & Borovik 1981, 1986; Déverchère et al. 1991, 1993), provided that $P_g$ waves are sampling only the upper $\sim 30$ km of the crust. We then changed this value in further tests in steps of 0.1 km s$^{-1}$ in the range 5.7–6.5 km s$^{-1}$.

Poisson's ratio

It is important to examine values of Poisson's ratio $\sigma$ in our study since they may help to elucidate the petrologic nature of the crust (Christensen 1996) and may modify the depth determination in the earthquake location procedure (e.g. Toth & Kisslinger 1984). In the BRS, these values were obtained using traveltimes of $P_g$ and $S_g$ waves at distances of 70–220 km (Krylov et al. 1981, 1991). Mean $\sigma$ values are found to vary from rift zone to the Siberian craton (0.262 and 0.246, respectively). In the northern Baikal rift zone, a better sampling (600 determinations) provides $\sigma$ values of 0.22–0.25 in the upper crust ($\sim 0$–15 km) and of 0.26–0.28 in the lower crust ($\sim 20$–40 km). This corresponds approximately to $V_p/V_S$ ratios of $\sim 1.68$ and 1.78, respectively. Krylov et al. (1991) explain this change with depth by the intrusion of hot, mafic ('basic') and ultra-mafic bodies in the lower crust. Mean $V_p/V_S$ values of $1.75 \pm 0.1$ are found in the Muya region which is equipped with dense networks (Golenetsky 1985; Anikanova & Borovik 1981, 1986; Déverchère et al. 1993).

From these results we infer that there are indeed significant changes in Poisson's ratio values at mid-crustal levels in the BRS, which probably correlate with mineralogical changes associated with progressive metamorphism (Christensen 1996). A similar increase in Poisson's ratio with depth was found in southwestern Germany and is thought to indicate a high plagioclase feldspar content of the lower crust (Holbrook et al. 1987). Although the spatial resolution on Poisson's ratio values is not high enough and our geological knowledge is too poor to make a simple correlation of Poisson's ratios with rock compositions, we recall that the BRS is situated at the contact of the Archean lithosphere with the Palaeozoic Sayan-Baikal 'mobile belt' made of numerous calc-alkaline to alkaline intrusions which crop out widely (Logatchev & Zorin 1992; Zorin 1999; Jahn et al. 2000). Therefore, if we consider results from Christensen (1996), the depth change in Poisson's ratio found here is likely to be governed by changes in lithology [from granite to diorite or gabbrroic (diabase) compositions, characterized by increasing temperatures of creep activation] and in metamorphic facies (from amphibolite to mafic granulite). For the earthquake location procedure, we set the initial value of $V_p/V_S$ at 1.72, and consider the influence of this parameter in the range 1.66–1.80.

Tests on initial parameters: depth, $V_p/V_S$, $V_P$

Considering the depths found in previous studies (see above), we made a first run using a trial depth ($z_{tr}$) of 15 km and the initial values of $V_p/V_S$ and $V_P$ selected above. First we impose
the following criteria in order to discard unreliable solutions: standard horizontal ($\varepsilon_{rh}$) and vertical ($\varepsilon_{rz}$) errors less than 5 km, and $rms$ less than 0.6 s. Using these additional criteria we obtain the initial data set (shown in Fig. 4) to be processed (i.e. 671 events, Table 2, Step 1).

Step 2 (Table 2) tests the effect of the initial depth $ztr$. Using $V_P=6.15 \text{ km s}^{-1}$ and $V_P/V_S=1.72$, we repeatedly compute the nine solutions per event with $ztr$ increasing from 0 to 40 km in depth intervals of 5 km. Events that depict depth changes of larger than 5 km between solutions are rejected. At this step, about two-thirds of the 671 selected events show depth changes of less than 5 km, although initial depths too far from the 'true' ones probably induce numerical instabilities in the inversion process.

Step 3 (Table 2) estimates the effects of changes in $V_P/V_S$. We fix $ztr$ at 20 km, since it appears that, among all the $ztr$ values tested, this value gives the best convergence results for the 428 remaining events. Leaving $V_P$ unchanged, we compute $rms$ for the nine subsets of events with $V_P/V_S$ values ranging from 1.66 up to 1.80 (Fig. 5a). Events with multiple minima are rejected. For each station we therefore obtain an average value of $V_P/V_S$ with its standard deviation (Table 2, Fig. 5b). About half of the selected events are still rather sensitive to the choice of the $V_P/V_S$ ratio, which is not surprising (see, for example, Hatzfeld et al. 1987). Nevertheless, the fairly constant minimum in the $V_P/V_S$ values, consistent with other studies, suggests that the $V_P/V_S$ value chosen should not induce a systematic bias on locations.

Figure 5. Plot of root-mean-square residuals ($rms$) versus $V_P/V_S$ ratio or $P$-wave velocity ($V_P$) for the 148 events located near station YLY (Fig. 4) and selected after Step 2 (see text for details). Plots (a) and (b): variation of $rms$ for each event versus $V_P/V_S$ ratio and $V_P$, respectively. Plots (c) and (d): mean variation of $rms$ for the 148 events versus $V_P/V_S$ ratio and $V_P$, respectively. Figs 5 and 6 may be viewed in colour in the online version of the journal (http://www.blackwell-synergy.com).
Finally, we perform a test on the remaining 229 events by varying the $V_p$ values from 5.7 to 6.5 km s$^{-1}$ (Table 2, Fig. 5c). Events without clear minima are again rejected (Table 2, Step 4). We compute nine mean $V_p$ values and associated errors by using the remaining 203 events (Table 2, Fig. 5d). These results show that changes in $V_p$ have a limited effect on locations. Note from Table 2 that standard deviations on depths are always less than 5 km for the selected events.

**Final relocation**

From Table 2 it appears that mean values of $V_p/V_S$ and $V_p$ do not change significantly from one site to another and that standard deviations remain in the range of mean values found. Therefore we again locate the 671 events selected at Step 1 using $V_p$, $V_p/V_S$, and $ztr$ values of 6.15 km s$^{-1}$, 1.73, and 20 km, respectively. It appears that 632 solutions depict vertical errors $\varepsilon zr$ less than 5 km (Table 2, Step 5) and rms values less than 0.4 s. Among these 632 events, 83 per cent, 99 per cent, 80 per cent, and 50 per cent depict values of $rms < 0.2$ s, $erh < 2$ km, $erz < 3$ km, and $erz < 2$ km, respectively. Most events selected therefore have standard errors $erz$ significantly less than 5 km. From these results we therefore estimate that the ‘true’ uncertainties on hypocentres are probably less than this value of 5 km for most of the 632 events, considering the control offered by the closest seismic station, the azimuthal coverage, the stability of the velocity structure found, and the upper bound imposed on $rms$ values.

Although the relative number of earthquakes depicts important changes from one station area to another (Fig. 6a), probably related to the limited number of events (especially for TIG and TAL stations), a general trend appears which is characterized (Fig. 6c) by: (1) very few earthquakes between depths of 0 and 10 km (~7 per cent); (2) a maximum activity between depths of 15 and 25 km (~50 per cent); and (3) significant activity in the lower crust, at depths from 25 to 45 km (~17 per cent).

From the depth variations depicted here, it is difficult to establish a clear link with mean observed heat flow variations (Lysak 1984, 1992) or steady-state geotherms (Dorofeeva & Lysak 1989) calculated in the vicinity of the nine stations (Table 2). We may only note that the two stations with the largest relative increase of shallow events surrounding them are the ones with the largest mean heat flow (YLJ and KMO, Fig. 6a and Table 2). Local heat flow values, geological changes, rock fracturing, and the earthquake depth distribution at each site seem to us too poorly known to enable a local interpretation of each histogram in terms of temperatures and rock composition. Therefore we try in the following section to select a mean depth–frequency distribution that depicts the overall seismogenic behaviour in the BRS.

**Interpretation of the Depth Distribution of Earthquakes**

**Background for interpretation and data selection**

Inferring the rheological state of the crust and/or lithosphere from the distribution of earthquakes at depth (e.g. Ranalli 1993, 1995; Seno & Saito 1994; Maggi et al. 2000a,b) rests on the basic assumption of consistency between the short-term deformation imaged by seismicity and the long-term mechanical properties of the lithosphere. One must keep in mind that this assumption is not necessarily true for many reasons, among which are the following: (1) the presence of seismicity indicates that the material is weak enough with respect to the ambient stress and deforms in a brittle way, whereas no seismicity means that the material is either too strong (elastic) or too weak (ductile) to deform seismically; (2) the time interval of seismic recording is short with respect to the time recurrence of earthquakes (i.e. brittle regions can appear aseismic for long periods of time). For these reasons, the presence of seismicity is meaningful (i.e. the material is brittle and the ambient stress is close to the yield limit), whereas its absence is much more difficult to interpret.

In the following section, we compare the abundance of earthquakes calculated at 5 km depth intervals with the strength predicted by various yield-stress envelopes (YSEs). The presence of earthquakes down to ~35–40 km indicates that the crust is close to the yield limit in stress for almost three-quarters of its depth extent. Hence, the earthquake distribution at a given depth may be closely related to crustal strength. The aim of this section is to determine which crustal rheology (defined by its thermal age and composition) best accounts for the observed depth distribution of earthquakes.

Taking into account the small variations of earthquake depth distributions observed from one site to another and the accuracy in depth determinations, still limited, we choose to plot the 632 events finally selected (Step 5, $erz < 5$ km, Table 2) and assume that this curve is representative of the mean depth distribution of earthquakes in the whole BRS (Fig. 6b). Indeed, the general pattern found here is roughly similar to the ones observed in regional curves previously determined in the BRS (Figs 3a and b), and, due to the larger spatial and temporal extents of events, is less likely to be dependent on regional variations of the physical properties of the crust (Fig. 3a) or on transient swam activities (Fig. 3b). As the absolute value of the yield stress depends on the mode of deformation (Byerlee 1978), we decided to rely on events that lie in areas undergoing extensional stress (Fig. 6c). Events located around stations ARS and MOY (Fig. 4) are discarded since the present-day stress field there is of strike-slip type (Petit et al. 1996). Both histograms (Fig. 6c), however, depict a similar pattern. In the following analysis, we use the 519 earthquakes relocated within the extensional zone of the BRS (Fig. 6c) as the basis for an interpretation of the yield-stress envelopes.

We now refer to the abundance of earthquakes (as a percentage of the total number) rather than to the absolute number. From Fig. 6, we find that about 99 per cent of earthquakes occur between 0 and 35 km depth. Following Lamontagne & Ranalli (1996), we use this depth of 35 km as a representative value for the lower limit of the seismogenic crust, and will hereafter call it the seismicity cut-off depth. In order to better compare the depth distribution of events and crustal rheology, we also represent the crustal yield strength at 5 km depth intervals as a percentage of the integrated yield strength of the crust. This allows a direct comparison between yield-stress envelopes (YSEs) and earthquakes/depth histograms. One must note, however, that these curves must not be interpreted in terms of absolute strength: a ‘weak’ rheology will tend to concentrate the crustal strength in a thin upper layer, thus showing a high strength proportion in that layer, while a ‘strong’ rheology will...
Figure 6. Plots of depth–frequency distribution of earthquakes determined from this study in the BRS. (a) Individual plot for each subset of events relocated around the nine stations (three-letter codes refer to stations in Fig. 4), classified according to the number of events at each station. The total number of events is 632 (Step 5, $erz < 5$ km, Table 2). (b) Histogram for the 632 events finally selected in the whole BRS (Step 5, $erz < 5$ km, Table 2), with a depth interval of 2 km. (c) As (b), with a depth interval of 5 km. The superimposed plot (solid grey bars) shows the 519 remaining events after discarding events located near stations ARS and MOY (Fig. 4), where the present-day stress field is not extensional (Petit et al. 1996).
distribute the strength over the whole crustal thickness, thus showing a relatively lower strength proportion in the same layer.

Assumptions for analysing the relative depth distribution of earthquakes

As explained above, comparing the earthquake distribution at depth with crustal yield-stress envelopes involves strong assumptions on the state of stress, stress drop during earthquakes, and stress regime, and requires a sufficient and relevant number of reliable focal depths. The abundance of seismicity (Nseis) in a given depth interval z depends on several parameters which describe the state of stress of the medium and the capacity of earthquakes to release this stress: the applied geological deviatoric stress σs(z), the stress drop for one earthquake Δσ, and the ratio of stress released by seismicity versus geological deviatoric stress [σs(z)/σs(z)]. To compare Nseis with the crustal yield-stress envelopes, we briefly justify here the following necessary assumptions. 

(1) Whatever the depth, the actual (geological) deviatoric stress σs is close to the deviatoric strength τ for the whole crust. This has recently been shown for depths of up to ~10 km by in situ measurements (Tournend & Zoback 2000). In the BRS, the presence of seismicity throughout the first 30–40 km of the crust, and sometimes deeper, indicates that, within this region, the deviatoric stress reaches the yield-stress limit down to the Moho (i.e. σs ~ τ).

(2) The static stress drop Δσ for one earthquake is constant, whatever the magnitude and depth of the earthquake. Indeed, for simple circular cracks there is a linear relationship between the seismic moment M0 and the source radius a which is expressed by (Keilis-Borok 1959; Lay & Wallace 1995, p. 388)

\[ M_0 = (16/7)Δσa^3. \]  

(2)

The observed dependence of M0 and a^3 suggests that the stress drop is approximately constant and independent of the earthquake size (Scholz 1990). This scaling relation is not easy to check and is only approximate, but since our earthquake database samples a small magnitude range (2–4.5), we consider here that it is true, so that the fault slip is roughly proportional to the fault surface.

(3) The effective stress released by seismicity (σs) is proportional to the deviatoric stress σs, whatever the depth. Common values of stress drop (a few megapascals) suggest that the stress released by seismicity in a given time period is much lower that the stored elastic stress (Scholz 1990). However, given the magnitude range examined here and the time window of several decades, there is no reason to consider that one particular level in the crust is more efficient in releasing stress through earthquakes than another.

Furthermore, we assume that our data set is representative of the long-term vertical seismicity distribution and that depth sources do not vary significantly with magnitude (see Discussion in Scholz 1990, and observations from Foster & Jackson 1998). Given these assumptions, we propose that the abundance of earthquakes at a given depth depends mainly on the mechanical strength at this depth. This allows us to make a straightforward comparison between the yield-stress envelopes and the observed depth–frequency distribution of earthquakes.

RHEOLOGICAL YIELD-STRESS ENVELOPES

Yield-stress modelling

We use empirical constitutive laws to obtain the yield-stress envelopes of continental crust with various lithologies and thermal ages. In this modelling approach, brittle failure controls the upper parts of the crust and mantle and is related to seismic activity, whereas non-Newtonian viscous creep occurs in the deeper parts of the lithosphere which do not deform seismically. For brittle failure we use the Von Mises criterion, so that (Byerlee 1978)

\[ σ_2 = (σ_1 − σ_3)/3.9, \quad σ_3 < 120 \text{ Mpa}, \]  

\[ σ_2 = (σ_1 − σ_3)/2.1 − 100, \quad σ_3 ≥ 120 \text{ Mpa}. \]  

Ductile behaviour is represented by power-law creep:

\[ \dot{ε} = A^* \exp(H^*/RT)(σ_1 − σ_3)^n, \]  

with A* = 5 × 10^-12, 5 × 10^-15, and 6.31 × 10^-20 Pa^-n s^-1 for quartz, diorite, and diabase, respectively; H* is the activation enthalpy with H* = 190, 212, and 276 kJ mol^-1 for quartz, diorite, and diabase, respectively; R is the gas constant; T is the absolute temperature at the given depth; and n is the power-law exponent (n = 3, 2.4, and 3.05 for quartz, diorite, and diabase, respectively).

Note that deeper, stronger, brittle levels—e.g. olivine-dominated upper mantle—can deform quasi-elastically if deviatoric stresses do not exceed the yield strength: In this case, no seismic activity should occur there (Scholz 1988; Burov & Diamant 1995). Determination of the yield-stress envelopes is carried out by assuming that the depth of the brittle–ductile transition depends primarily on thermal gradient and crustal composition (Goetze & Evans 1979; Chen & Molnar 1983; Smith & Bruhn 1984). The strain rate is fixed at 5 × 10^-10 s^-1, which roughly corresponds to ~3.5 mm yr^-1 of velocity across the 200 km wide Baikal rift (Calais et al. 1998; Calais & Amarjargal 2000; Zorin & Cordell 1992; San’kov et al. 2000). Strain-rate variations of one order of magnitude produce depth variations of the brittle–ductile transition of less than 5 km (see, for example, Cloetingh & Banda 1992), so that we consider this uncertainty to have a negligible effect on the model given the resolution of our data.

As explained in more detail in the discussion, we do not take into account possible effects of pore pressure and bending forces (flexure) on the YSEs; we acknowledge that these factors may indeed influence the depth of the brittle–ductile transition but they are rather difficult to constrain. Moreover, they can be highly variable along the rift system (and even along a single fault system). Since our study aims at analysing the average crustal strength in the BRS, we believe that we cannot precisely address these points owing to the limits imposed by our data set and to the assumptions described above. We present the total (brittle + ductile) crustal YSE and compare it with the depth distribution of earthquakes, in order to discuss the possible discrepancies between the predicted brittle–ductile transition and the maximum depth of earthquakes (e.g. Lamontagne & Ranalli 1996).
Heat flow measurements indicate a moderate surface heat flow accreted onto the Archean lithosphere to the north and east. According to 100 Ma (Jahn et al. #)

The surface temperature is constant whatever the age (0°C), as well as the temperature of the bottom of the thermal lithosphere \( T_m = 1350 \)°C at 250 km. At initial time \( T_0 \), the temperature distribution is homogeneous \( (= T_0) \). We define the thermal age as the age of the last large-scale thermo-tectonic event. Although the Baikal rift is located close to the Siberian craton (Fig. 2), for which the thermal age coincides with the geological one, all the seismicity used here is located inside the ‘mobile belt’ (generally called ‘Sayan-Baikal’), made of calc-alkaline to alkaline series intruded in several episodes from \( \sim 450 \) to \( \sim 200 \) Ma (Jahn et al. 2000, and references therein), and accreted onto the Archean lithosphere to the north and east. Heat flow measurements indicate a moderate surface heat flow of about 50–60 mW m\(^{-2}\) in the mobile belt and of less than 40 mW m\(^{-2}\) in the craton (Lysak 1984, 1992). For all our models, we use an average crustal thickness of 45 km, which is in agreement with previous studies of the crustal structure in the BRS (see above, and Zorin et al. 1989; Krylov et al. 1991; Petit et al. 1997). Note that changing the Moho depth by \( \pm 5 \) km, for instance, would not change the depth of the brittle–ductile transition within the crust, which occurs above the Moho; it would just increase or decrease the thickness of the ductile crust and hence reduce or enhance the effective lithospheric strength, by allowing or impeding mechanical decoupling between the strong upper crust and upper mantle layers (e.g. Burov & Diament 1995, 1996). In our case there is a 5–10 km layer between the seismicity cut-off depth \( (\sim 35 \) km) and the Moho \( (45 \pm 5 \) km). Finally we modify the thermal ages and crustal composition (lithologies) of rheological models in order to fit the predicted curves with the depth distribution observed in the previous section. Major constraints on these parameters are obtained from the present-day knowledge of the structure.

Temperature structure and crust thickness (Moho depth)

We first test various thermal ages for the continental lithosphere and compute corresponding yield-stress limits. We consider the temperature structure of the continental lithosphere to be governed, at first order, by a cooling plate model (for details see Burov & Diament 1995) incorporating radiogenic heat production in the crust and frictional heating at the crust–mantle boundary. The temperature is constant whatever the age (0°C), as well as the temperature of the bottom of the thermal lithosphere \( T_m = 1350 \)°C at 250 km. At initial time \( T_0 \), the temperature distribution is homogeneous \( (= T_0) \). We define the thermal age as the age of the last large-scale thermo-tectonic event. Although the Baikal rift is located close to the Siberian craton (Fig. 2), for which the thermal age coincides with the geological one, all the seismicity used here is located inside the ‘mobile belt’ (generally called ‘Sayan-Baikal’), made of calc-alkaline to alkaline series intruded in several episodes from \( \sim 450 \) to \( \sim 200 \) Ma (Jahn et al. 2000, and references therein), and accreted onto the Archean lithosphere to the north and east. Heat flow measurements indicate a moderate surface heat flow of about 50–60 mW m\(^{-2}\) in the mobile belt and of less than 40 mW m\(^{-2}\) in the craton (Lysak 1984, 1992). For all our models, we use an average crustal thickness of 45 km, which is in agreement with previous studies of the crustal structure in the BRS (see above, and Zorin et al. 1989; Krylov et al. 1991; Petit et al. 1997). Note that changing the Moho depth by \( \pm 5 \) km, for instance, would not change the depth of the brittle–ductile transition within the crust, which occurs above the Moho; it would just increase or decrease the thickness of the ductile crust and hence reduce or enhance the effective lithospheric strength, by allowing or impeding mechanical decoupling between the strong upper crust and upper mantle layers (e.g. Burov & Diament 1995, 1996). In our case there is a 5–10 km layer between the seismicity cut-off depth \( (\sim 35 \) km) and the Moho \( (45 \pm 5 \) km). Finally we modify the thermal ages and crustal composition (lithologies) of rheological models in order to fit the predicted curves with the depth distribution observed in the previous section. Major constraints on these parameters are obtained from the present-day knowledge of the structure.

Results of modelling

We compute thermal profiles and yield-stress envelopes (YSEs) for thermal ages of 20, 50, 80, 100 and 150 Ma (Figs 7 and 8). Surface heat flow \( Q \) is computed from thermal gradients in the upper 20 km by

\[
Q = -k \Delta T / \Delta z ,
\]

where \( k \) is the thermal conductivity in the crust \( (k \sim 3 \) W m\(^{-1}\) K\(^{-1}\)).

The thermal gradients for ages of 20, 50, 80, 100 and 150 Ma correspond to surface heat flows of about 99, 69, 58, 53 and 46 mW m\(^{-2}\), respectively. Compared to measured surface heat flows (see above and Lysak 1984, 1992), thermal profiles thus give us a first indication of a thermal age of about 80–100 Ma for the BRS lithosphere. Note that, for a 100 Ma lithosphere, the temperature at the seismicity cut-off depth of 35 km is

![Figure 7](image-url)

Figure 7. Computed temperature/depth profiles for a cooling plate with a thermal age of 20, 50, 80, 100 and 150 Ma. The critical depth of 35 km corresponding to the seismicity cut-off is marked.

![Figure 8](image-url)

Figure 8. Yield strength (as a percentage of the total crustal strength) for various thermal ages and abundance of earthquakes (as a percentage of the total number of events, in steps of 5 km intervals) versus depth. Yield strength is computed each 5 km for thermal ages of 20, 50, 80, 100 and 150 Ma. The crust is made of a single diabase layer extending from 0 to 45 km. Black dots are the observed depth distribution of earthquakes from this study. Horizontal lines delineate the theoretical depth of brittle–ductile transitions for each thermal age.
around 500 °C (Fig. 7), which is higher than the onset of feldspar plasticity (450 °C; see Scholz 1988). We compute corresponding yield-stress envelopes with a relatively mafic crustal composition (diabase; see, for example, Mackwell et al. 1998) for the crust, because: (1) we know that deep (30 km) seismicity cannot be found given a quartz-rich composition throughout the crust; (2) previous studies (see above) have shown that the change in Poisson’s ratio with depth is likely to correlate with changes in lithology (from granite to diorite or gabbro) and in metamorphic facies (from amphibolite to mafic granulite); and (3) we have no constraints, at this point of the study, on a possible quartz (or quartz-feldspar)-diabase transition within the crust. By choosing a ‘strong’ crust instead of a ‘weak’ one, we avoid overestimation of the thermal age (since a ‘weak’ rheology would require a much colder geotherm to fit the observed seismicity at depth).

The strength distribution at depth for a single diabase layer extending from 0 to 45 km (Fig. 8) shows that for young thermal ages the strength is concentrated at shallow (10–15 km) depths and is much too low at depths greater than 20 km compared to the observed earthquake distribution. For an old (150 Ma) crust, the strength proportion is relatively too low between 10 and 20 km compared to the proportion of earthquakes at these depths (Fig. 8). This first test shows that ages of 80–100 Ma (with a better fit for 100 Ma) provide a good fit between yield strength and earthquake distribution at depth. However, some discrepancies remain near the upper (10 km) and lower (35 km) bounds of the depth profile. The YSE indicates too high a strength at shallow (10 km) depths and too low a strength at greater (30–35 km) depths compared with the observed distribution. Furthermore, for all thermal ages, the predicted brittle–ductile transition is shallower than the seismicity cut-off depth (i.e. 35 km).

In a second step, we want to test whether variations of crustal composition at depth could be responsible for the observed discrepancies. Indeed, a quartz-rich composition may explain the relative abundance of earthquakes in the uppermost crust. This may appear paradoxical since a quartz rheology is weaker than a diabase one. In fact, the choice of a diabase rheology for the whole crust (Fig. 8) induces a single, thick brittle layer where the seismicity is linearly distributed, while a quartz–diabase rheology introduces an intermediate ductile aseismic level, thus increasing the relative proportion of earthquakes in other (upper and lower) elastic layers. To test this, we compute yield-stress profiles for a two-layer crust, with the first layer composed of quartz and of varying thickness from 0 to 20 km, and the second layer made of diabase extending to the Moho (Fig. 9). The thermal age is fixed at 100 Ma. Brittle–ductile transitions for various rheologies and abundance of earthquakes versus depth. The crust is made of a shallow quartz-rich layer and a deeper diabase-rich layer of varying thicknesses [for example, Q(10) Db(35) means the quartz and diabase layers are 10 and 35 km thick, respectively]. The thermal age is fixed and is 100 Ma. Brittle–ductile transitions in the quartzic and diabase layers are indicated. See Fig. 8 for other notations.

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**DISCUSSION**

Whether the maximum depths found in seismicity (particularly after large shocks) determine the transition to stable friction or to plastic flow remains questionable (Boatwright 1985; Scholz 1990; Kohlstedt et al. 1995). This is explicitly expressed by Lamontagne & Ranalli (1996), who discuss the ‘brittle–ductile’ model against the ‘velocity weakening–strengthening’ model. This
leads us to question the overall significance of the seismogenic layer: does it represent the long-term strength or the frictional instability? In this study, rheological predictions made for a 10 km quartz and 35 km diabase rheology and for a range of acceptable parameters concerning thermal state, strain rate, composition and pore fluid pressure of the crust allow us to build a YSE able to explain fairly well the depth distribution of earthquakes found in the BRS. Although the second model seems more suitable to explain the depth distribution of earthquakes in some other continental areas in the world (e.g. Lamontagne & Ranalli 1996; Kaikkonen et al. 2000), we find, as, for instance, Cloetingh & Banda (1992), Seno & Saito (1994), Nyblade & Langston (1995) or Maggi et al. (2000a,b) that a ‘brittle–ductile’ model implying increased strength at mid-crustal levels satisfactorily predicts the lower crustal seismicity found here (i.e. at ~35–40 km depth) and in other continental areas such as the East African rift system. This behaviour could be linked to residues of anhydrous mafic rocks left from the loss of granitic melts which are known to strongly modify crustal strength (e.g. Mackwell et al. 1998). Below we develop the reasons justifying this opinion in our case study (the BRS), and discuss the ability of YSEs to describe correctly the thermomechanical behaviour of the continental lithosphere under extension.

Pore pressure effects
The yield-stress envelopes shown here do not take into account possible effects of high pore pressure, as do Seno & Saito (1994), for example. High pore pressure reduces frictional resistance and thus allows brittle failure to occur at low deviatoric stress. Given the relatively strong rheology used here, pore pressure effects are probably necessary to reduce the crustal strength with respect to the ambient stress, which is known to be of the order of ~100 MPa (e.g. Zoback & Magee 1991). However, adding a high pore pressure throughout the crust would not change the general shape of strength percentage curves in the crust: it would just allow the brittle–ductile transition to occur at slightly greater depths (<5 km), because the slope of the elastic–brittle domain limit increases with pore pressure. The question is to know whether we could predict the same fit to the depth–frequency distribution of earthquakes with a different rheology using a high pore pressure. If we choose a younger thermal age (50 Ma), and play with the pore pressure parameter, we can reduce the slope of the elastic–brittle limit until it fits the depth distribution between 0 and 25 km; at greater depths, however, we would still be in the ductile domain and would not be able to predict the seismicity encountered there. Choosing a weaker (quartz–diorite) rheology would give the same misfit, namely a too shallow brittle–ductile transition.

Significance of earthquakes in the lower crust
Our results show unequivocal evidence for earthquakes in the lower crust at depths where it is supposed to be ductile according to YSEs. Indeed, the brittle–ductile transition is at ~25 km depth according to our best-fitting model (Fig. 9). Still, the earthquake distribution at depth is consistent with the strength proportion predicted there, even within ductile levels. The preferred YSE (quartz-dominated between 0 and 10 km and diabase-dominated between 10 and 45 km, for a ~100 Ma thermal age) is consistent with previous studies of Poisson’s ratio (where changes correlate with a diabase composition at depth; see, for example, Christensen 1996), heat flow measurements and the geological history of the rift zone. Let us recall that the crust below the Baikal rift may be rather dry and mafic owing to the presence of the underlying Archean crust of the Siberian craton and to strong differentiations of the Caledonian and Hercynian crust within the Sayan-Baikal mobile belt (Logatchev & Zorin 1992). Therefore, we believe that this best-fitting strength profile is representative of the long-term behaviour of the crust, and that the presence of earthquakes in ductile levels is not due to a poor evaluation of the YSE, but is likely to represent brittle failure in the semi-brittle field within a mafic lower crust, above the level of a fully plastic mode of deformation (Scholz 1990). Another possibility is that flexure and extensional forces linked to rift dynamics modify the YSE and drag the brittle–ductile transition downwards (e.g. Kuszniir & Karner 1985; Burov & Diament 1995). One must remember, however, that our study encompasses the whole rift zone, for which the flexural state of the lithosphere is known to vary a lot along a single fault system (e.g. Petit et al. 1997); therefore the effects of flexure should be unimportant when averaged on the scale of the rift system.

We are aware that the deep crustal seismicity found in several other intracontinental settings, such as Tien Shan or the Indian shield (see the review in Maggi et al. 2000b), is not necessarily controlled by the brittle–ductile transition, as suggested above. Heterogeneities of composition, structure, fluid content and metamorphic grade, which are known to be large within or around cratons at mid-crustal levels (e.g. Burov et al. 1998; Karlstrom & Williams 1998), have important consequences for strength, but have not been considered here, since they are difficult to model. Deichmann & Rybach (1989) have also reported lower crustal earthquakes in Alpine Switzerland, at depths where the crust is supposed to be ductile, but where brittle failure seems unlikely given the plausible temperature field and petrological models. They point out the need for more accurate experiments on mineral composition, pore pressure and temperature variations in the lower crust. An important point to note is that brittle behaviour is not time-dependent, so the elastic–brittle limit does not change with time; conversely, non-linear ductile behaviour is time-dependent, so that short-term deformation is not necessarily representative of the long-term (steady-state) behaviour described by the YSE. In other places, the brittle–ductile transition can be deeper than the maximum depth of earthquakes, as shown by Lamontagne & Ranalli (1996). These authors found that, in the Charlevoix, Canada, intra-plate seismic zone, most of the earthquake foci are located not deeper than 25 km, while the crust is supposed to be brittle down to 30 km, according to their thermal models. They infer that the velocity weakening/velocity strengthening transition within the crust can be a factor controlling the seismicity cut-off depth, as also found by Kaikkonen et al. (2000) in the Fennoscandian shield.

Mantle earthquakes and upper mantle rheology
This study focused mainly on rheological properties of the crust, since most earthquakes recorded in the BRS are located within the crust. Our results agree with those of Maggi et al. (2000a,b) regarding the scarcity of events in the continental upper mantle. These authors conclude from this observation.
that the upper mantle has no strength, and therefore that the effective thickness of the lithosphere \((T_e)\) is the crustal seismogenic thickness \((T_s)\). However, as explained above, the lack of seismicity in the upper mantle has no real significance to us, since we do not know how it is stressed and how reliable the seismicity database is if earthquakes have long recurrence intervals (which is indeed to be expected if the upper mantle is particularly strong). As underlined by Seno & Saito (1994), appropriate conditions for pore pressure and strain rate are required in order to generate earthquakes within an upper mantle with high strength; indeed, in the case of the BRS, even if the strength is lowered by high pore pressure within intermediate heat flow areas, it could be difficult for the ambient stress to reach the peak strength necessary for brittle failure for a relatively low strain rate. Furthermore, earthquakes within the upper mantle, although very rare, are found by Vertlib (1981, 1997), for instance, and are suspected by Déverchère et al. (1991) in the BRS. This should not occur if the hypothesis of a weak mantle is true. Moreover, forward modelling of lithospheric flexure (which is not sensitive to the use of Bouguer or free-air gravity, as the coherence method is) provides \(T_e\) estimates in the Baikal and East African rifts that are significantly greater than the total crustal thickness, and, therefore, \(T_e\). This is in disagreement with what is postulated \((T_e < T_s)\) by McKenzie & Fairhead (1997) and Maggi et al. (2000a,b). In the BRS, many \(T_e\) estimates from forward modelling reach \(\sim 50\) km (Ruppel et al. 1993, Burov et al. 1994; van der Beek 1997; Petit & Ebinger 2000). For a two-layer lithosphere of elastic thicknesses \(h_1\) and \(h_2\), the integrated strength \(T_e\) is given by (Burov & Diamant 1995)

\[
T_e = \left( h_1^2 + h_2^2 \right)^{1/3},
\]

(7)

where \(h_1\) and \(h_2\) are the thicknesses of the crust and mantle elastic layers, respectively. Considering a crustal elastic thickness \(h_1\) of about \(30 \pm 5\) km and an integrated strength \(T_e\) of \(\sim 50\) km, we obtain an elastic thickness \(h_2\) for the upper mantle of about \(45 \pm 3\) km. Therefore, in our view, an alternative reason for the upper mantle to be aseismic is that it is too strong; that is, the applied deviatoric stress is lower than the yield-stress limit. Indeed, the YSE for an olivine-dominated rheology shows that the yield-stress limit can easily reach more than 1 GPa at depths of 50–60 km (see the review in Vauchez et al. 1998). From another point of view, the assumption that the mantle has no elastic strength would also imply that surface loads such as topography are isostatically compensated within the crust (McKenzie & Fairhead 1997), and that there is almost no Moho topography due to isostatic compensation of relief, which goes against direct evidence provided by Bouguer gravity data (see the discussion in Petit & Ebinger 2000). Cloetingh & Banda (1992) have similarly explained the absence of earthquakes in the mantle part of continental lithosphere by a strong mantle decoupled from the overlying crust, in the light of the high values of \(T_e\) found for most continental interiors.

**Conclusions**

From a revised data set of 632 focal depths in the Baikal rift, we have built bulk and regional depth–frequency distributions of earthquakes that have been compared with estimated YSEs of the lithosphere, taking into account the temperatures, stress regimes and strain rates known in the area. We have shown that the crust is seismogenic at depths as great as \(35–40\) km in all areas studied in the BRS, and that very few earthquakes are below the Moho (one additional event has been found in this study). The depth at which a decrease in the distribution of earthquakes occurs is generally found to be \(20–25\) km. Brittle–ductile transitions and YSEs predicted for a 10 km quartz and 35 km diabase rheology agree fairly well with this depth distribution of earthquakes, using a steady-state temperature structure and a strain rate fitting the BRS case. From this analysis we suggest that a brittle–ductile model with a relatively high strength at mids-crustal levels is quite able to explain the lower crustal seismicity found (i.e. at \(\sim 35\) km depth), as proposed in other continental areas such as the East African rift system. Although many parameters such as instability factors and lithospheric heterogeneities may explain the occurrence of deep crustal earthquakes and the nearly aseismic character of the upper mantle, we infer from our modelling that the depth distribution of seismicity in the BRS is governed by a proportionality to the yield-stress limit in a rheologically layered lithosphere with a dominant mafic composition of the crust in the \(10–45\) km depth range and with an upper mantle high-strength layer \(\sim 45\) km thick deforming mostly elastically.

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