Abstract. The velocity structure of the upper lithosphere in the Northern Baikal rift (NBR), Siberia, is analyzed in order to bring some constraints on the early stages of continental rifting. A simultaneous hypocenter-velocity inversion is performed using a set of local and regional seismological data recorded in the NBR. First-arrival times of 3.492 direct and refracted P and S waves allow to investigate the crustal velocity structure and the Moho geometry. Results of inversion reveal an unbalanced crust made of two adjacent parts of different thickness, delimited by a rather sharp discontinuity following approximately the northern edge of the rift zone. This structure is interpreted as the result of ante-rift deformation phases, creating a weakened zone where rifting deformation may easily occur. Oblique rifting is invoked to explain the formation of seismic swarms which occurred between 1976 and 1990. Uncertainties on P and S arrival times are 0.1 and 0.4 seconds, respectively. One-dimensional velocity models have already been used in this region for event location. An average 1-D structure is shown on Figure 2A. More detailed 1-D model and relocation concerning the North Muya region (Figure 2B) are exposed in Déverchère et al. [1993]. These models were determined with a method of RMS (Root Mean Square) residual minimization. For better homogeneity, we locate in this paper the whole set of events with the simplest 1-D model (Figure 2B), which is a good approximation of the average velocities in the NBR. We make a selection of the best-constrained events using the following criteria: number of P arrivals > 7 and at least 1 weighted S phase, RMS < 1 s, azimuthal gap < 20°, error on horizontal position < 7 km, and error on vertical position < 12 km. Because of the small magnitude of most earthquakes, the first criterion is often very selective, especially for P recordings. The resulting dataset consists of 204 events (1937 P and 1555 S arrivals, Figure 1). A Vp/Vs ratio of 1.72 +/- 0.01 was obtained from a modified Wadati diagram. It is used to specify S wave velocities in further inversions.

Method
We use a method that allows simultaneous inversion of hypocentral locations and 3-D velocity structure. Adjustments on velocities and hypocenter locations are made simultaneously to minimize (in damped least square sense) the travel time residuals [see e.g. Abers and Roecker, 1991]. The earth is modelled by a set of layers cut into rectangular blocks of variable dimension and constant velocity. A velocity damper may be used to avoid too important velocity adjustments in badly sampled blocks. Iterations go on until the change in the variance and velocity adjustments become small enough (2%). Travel times and derivatives are calculated with an approximate method of ray-tracing through 3-D structures [Thurber and Elsworth, 1980].

Because of the shallowness of the events and the distribution of stations (Figure 1), we focus on the continental crust and upper mantle. In our particular case, we had the opportunity to use two kinds of data, quite visible on selected seismo-
The aim of 1-D inversion is to determine the most suitable 1-D velocity structure in order to provide a good starting point for further 3-D inversions [e.g., Scott et al., 1994]. We test several priori models based upon slight changes applied to model B (Figure 2). The number of layers does not exceed 3, because more detailed structure is beyond the resolution we expect from our data. Different thicknesses and velocities are tested for each layer. Model A was used for initial event location. We choose to present here three models (C, D and E in Figure 2), which give a good overview of the obtained results. The routine needs 2 iterations to converge. Resolution matrix diagonals \( R_m \) [Tarantola and Valette, 1982] are greater than 0.90, which means that velocity parameters are well resolved. Standard errors are 0.01 and 0.04 km/s for layers 2 and 3, and layer 1, respectively. They reflect computed errors linked to the inversion, but not actual uncertainties on velocities. In order to estimate the latter, we calculate the differences resulting from small perturbations applied to each model (Figure 2). High values of data variance for models B and D seem to show that they do not fit well the data. Moreover, an unlikely low-velocity zone is appearing in model D. It appears that only models C and E give coherent results and small data variance. Velocities in layers 2 and 3 are constrained and stable enough around 6.3 and 7.9 km/s, respectively. Layer 1 is characterized by a P velocity between 5.9 and 6.0 km/s. Depth of the first boundary seems better constrained in model E (12 km) because it depicts the strongest velocity change between layers 1 and 2 [Roeker et al., 1987]. For the same reasons, the second boundary (Moho) can be located between 45 and 48 km depth. Based upon these results, we deduce a fourth average model taking into account these observations (model F in Figure 2). This model depicts the smallest data variance. Nevertheless, it is now obvious that 1-D inversion only gives a rough idea of the earth structure below this area: local structure may exhibit significant lateral variations. Three-dimensional inversion is clearly needed to image these differences.

Three-Dimensional Inversion: Results and Discussion

In a first step, let us infer that crustal layers are flat and horizontal in order to describe the \( V_p \) spatial variations. Blocks are delimited by N-S and E-W trending boundaries spaced every 31 km and 28 km, respectively. Adjacent blocks may have the same velocity parameter. Each layer has 10 metablocks. Model F is used as initial one to start 3-D inversion procedure. Result of inversion is shown in Figure 3. For layer 1, resolution is rather good (\( R_m = 0.6 \) to 0.9) for most of the blocks. Standard errors are ranging between 0.03 and 0.07 km/s. Using the same procedure as above, we estimate actual uncertainties to range between 0.03 and 0.30 km/s. Lowest velocities are located below the Upper Angara basin (5.9 and 5.6 km/s), in the Muyakan basin (5.8 km/s), and in the Kitchera basin (5.9 km/s). Highest velocities (6.1 and 6.4 km/s) are found in the eastern edge of the rift.

For layer 2, all the blocks are well resolved (0.79 < \( R_m < 0.97 \) ) and estimated uncertainties are lower than for layer 1 (about 0.1 km/s, see Figure 3). The main feature is a clear asymmetry of the rift, with a strong decrease of \( V_p \) from NW (6.5 km/s) to SE (6.0 km/s). High values are also found in the Barguzin area (6.4 km/s) and south of the Upper Angara basin (6.5 km/s).

For layer 3, \( V_p \) values are well constrained for most of the blocks, except below the Upper Angara and Kitchera basins (Figure 3). A clear minimum is located below the lake and Kitchera basin (about 7.6 km/s). Strong \( V_p \) variations in layer 1 are expected because of sedimentary infills in the numerous basins of the rift axis. Corresponding decrease of density from 2250 kg/m\(^3\) for cen-
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The initial velocity structure is: 5.95 km/s between 0 and 12 km depth, 6.30 km/s between 12 and 52 km (54 km for model E), and 7.90 km/s below 52 km (54 km for model E), considering the results of 1-D inversion (Figure 2). Each block has a different velocity parameter, in order to observe independently changes on travel time delays. Considering as invalid the assumption of flat layers, let us assume now a constant velocity for layer 2 and calculate what should be the difference of thickness needed to reproduce the same travel time delays. We first compute the travel times in both NW and SE metablocks for a refracted P wave coming from one virtual earthquake and using the P velocities resulting from 3-D inversion (6.5 km/s and 6.0 km/s, respectively). A simple calculation gives the corresponding thickness needed to reproduce the same travel times with a constant velocity layer of 6.3 km/s. Travel time for refracted rays is:

\[ T = \frac{2H - d}{V_{pg} \cos i + L/V_{pn} - (2H - d) \tan i/V_{pn}}, \]

where:
\[ T = \text{travel time}; \ d = \text{hypocentral depth}; \ H = \text{thickness of layer 2}; \ i = \text{angle of incidence of Pn wave}; \ L = \text{length of travel on the Moho}; \ V_{pg} = \text{P velocity in layer 2}; \ V_{pn} = \text{P velocity in layer 3}. \]

This hypothesis allows us to propose an inferred depth of the Moho below the NW and SE edges of the NBR within a 2 km range. This assumption is more realistic than involving flat layers, considering that previous observations resulting from different methods have led to similar conclusions [Zorin et al., 1989; Burov et al., 1994]. We now try to determine the Moho geometry by reproducing structural changes along the rift. This is done using several possibilities involving one, two or three vertical boundaries (VB) delimiting a northwestern block (NWB), the rift axis (RA) and a southeastern block (SEB). We test the following hypotheses (Figure 4): one VB in A separating NWB from RA and SEB (Figure 4A); one VB in B separating NWB and RA from SEB (Figure 4B); two VB in A and B separating NWB from RA and SEB, with a Moho below RA at intermediate depth (48 km) or uplifted (41 km, see Figure 4C and 4D, respectively); three VB in A, B and C simulating an upward bending of the Moho in NWB and downward bending in SEB (Figure 4E).

The initial velocity structure is: 5.95 km/s between 0 and 12 km depth, 6.30 km/s between 12 and 52 km (54 km for model E), and 7.90 km/s below 52 km (54 km for model E), considering the results of 1-D inversion (Figure 2). Each block has a different velocity parameter, in order to observe independently
its convergence towards crustal or mantelllic value. Results are shown in Figure 4. All the parameters are well resolved ($R_m > 0.9$). Data variance is not improved compared to flat-layer version (Figure 3), but we neglect here the effect of heterogeneities in layer 1. Models C and D depict too high discrepancies to be kept. Model A gives the best fit to data and depicts the strongest velocity change in layer 2 between NWB and SEB: Vp in NWB reaches 7.40 km/s below 44 km, which may reveal an anomalous mantle, meanwhile SEB has a Vp of 6.35 km/s, which is close to crustal velocity. For models B and D, this distinction is much less clear; moreover, an unlikely low velocity zone appears at depth greater than 40 km.

Conclusions

A first attempt of velocity inversion of nearby earthquake arrival times in the North Baikal rift has allowed us to assess a rough structure of the crust in this region. The upper part of the lithosphere in the NBR exhibits lateral variations which cannot be fully described by 1-D modelling. Three-dimensional modelling allows to preserve the atypical character of this region compared with the southern rift [Gao et al., 1994]: models involving major crustal thinning below the rift axis seem not suitable. Instead, a rather sharp discontinuity (step model of Figure 4A) separating two adjacent blocks of different thickness (44 km in the NW and 52 km in the SE) better explains the velocity anomalies found. This result agrees remarkably well with gravity modelling which predicts downward flexure of the southern plate (1-2 km) and upward flexure of the northern one (4-5 km) [Burov et al., 1994]. This discontinuity may result both from asymmetric load redistribution between the two plates and from anti-rift deformation phases which possibly led to thickened and heterogeneous crust in the south [Delvaux et al., in press]. Another point is to explain why superficial extensional structures are not linked with crustal thinning at depth in this part of the rift. Oblique rifting evidenced in the NBR [Houdry, 1994] may be responsible for this phenomenon: like beneath many strike-slip and transform faults involving major crustal thinning below the rift axis seem not suitable. Instead, a rather sharp discontinuity (step model of Figure 4A) separating two adjacent blocks of different thickness (44 km in the NW and 52 km in the SE) better explains the velocity anomalies found. This result agrees remarkably well with gravity modelling which predicts downward flexure of the southern plate (1-2 km) and upward flexure of the northern one (4-5 km) [Burov et al., 1994]. This discontinuity may result both from asymmetric load redistribution between the two plates and from anti-rift deformation phases which possibly led to thickened and heterogeneous crust in the south [Delvaux et al., in press]. Another point is to explain why superficial extensional structures are not linked with crustal thinning at depth in this part of the rift. Oblique rifting evidenced in the NBR [Houdry, 1994] may be responsible for this phenomenon: like beneath many strike-slip and transform faults, deep depressions may take place at the surface without significant crustal thinning at depth.

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