Lithosphere rheology and active tectonics in Mongolia: relations between earthquake source parameters, gravity and GPS measurements

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SUMMARY
In this paper, we use observations of earthquake source parameters and gravity to investigate the mechanical properties and the active faulting of the lithosphere in Mongolia. Well-determined earthquake centroid depths, including 10 from inversions of P and SH waveforms that are presented here for the first time, show that the seismogenic thickness (Ts) within Mongolia itself is less than 20 km. However, to both the east, in parts of the Lake Baikal rift system, and the west, adjacent to the Junggar basin and Kazakhstan platform, the seismogenic thickness is considerably greater, and includes essentially the whole crust. From the admittance between the free-air gravity and the topography, and also from profiles across a flexural foreland basin, we determine the effective elastic thickness (Te) in central Mongolia to be <10 km, though it may be a little greater (<20 km) adjacent to the Gobi-Altay range in the south. Further west, adjacent to the Kazakhstan platform, the same techniques show that Te > 30 km. In both Mongolia and its surroundings, Te is comparable with Ts and, where it is well determined, Te < Ts. Nowhere do the data require that Te > Ts. These data are consistent with the view that the strength of the continental lithosphere resides in its seismogenic part, which in Mongolia is the upper crust, but to both the east and west appears to be the whole crust. The earthquake source parameters also allow us to ask how the active faulting in Mongolia accommodates the velocity field revealed by GPS measurements. It is likely that the entire Mongolian Altay range in the west rotates counter-clockwise relative to stable Asia, and is responsible for the distributed E–W left-lateral shear seen further east in central Mongolia. The admittance observations show no evidence at the present day for convective mantle support, or a ‘hotspot’, responsible for the elevated region of the Hangay dome in central Mongolia, even though the geochemical data from nodules in late Cenozoic basalts and seismic tomography studies suggest elevated temperatures at shallow depths (<125 km) and probably thinned lithosphere.

Key words: continental tectonics, earthquakes, GPS, gravity, lithosphere, rheology, Mongolia.

1 INTRODUCTION
In recent years there has been much renewed interest in the rheology of the continental lithosphere and, in particular, in the relations between manifestations of mechanical behaviour, such as the depth extent of earthquakes and the effective elastic thickness determined from gravity anomalies. In the case of earthquakes, a temperature-dependent change from shallow, friction-dominated slip on faults to deeper, aseismic creep processes is expected (Brace & Kohlstedt 1980), and in most continental regions earthquakes are restricted to the upper half of the crust (Chen & Molnar 1983). But variations are seen, with some areas, particularly those associated with Archean or Proterozoic shields such as in East Africa and India, being associated with earthquakes throughout the thickness of the crust (Maggi et al. 2000a). In a reassessment of continental seismicity, Maggi et al. (2000a) and Jackson (2002) found little evidence for significant earthquake activity in the upper mantle beneath continents. On continents, the thickness of the seismogenic layer (Ts) in which earthquakes occur appears to involve either the upper crust, or the whole crust, but not, to any significant extent, the mantle.

Gravity anomalies associated with topographic or internal loads on the continents indicate the ability of the lithosphere to support elastic stresses over geological timescales. The gravity anomalies and loads are usually modelled as if they were supported by a uniform elastic beam of thickness Te over an inviscid half-space. Then Te is known as the effective elastic thickness. In a reassessment of
gravity anomalies on the continents McKenzie & Fairhead (1997), Maggi et al. (2000a) and McKenzie (2003) found that the effective elastic thickness ($T_e$) tracked the seismogenic thickness ($T_s$), such that larger values of $T_e$ were found where $T_s$ involved the whole crust. Although $T_e$ was not always well resolved, they found, in general, that $T_e < T_s$, and they found nowhere where the data required that $T_e > T_s$. As Maggi et al. (2000a) and Jackson (2002) pointed out, the simplest interpretation of these results is that the long-term strength of the continental lithosphere resides in its seismogenic layer, which is either the upper crust or the whole crust, but does not include the mantle. It is possible to devise other interpretations of either the gravity data or the seismological data on their own, but the interpretation that the long-term strength resides in the seismogenic layer is the simplest that reconciles both gravity and earthquake data together, and is also consistent with the known temperature dependence of rock properties. That interpretation works satisfactorily in both oceans and continents (McKenzie et al. 2005).

These arguments require high-quality estimates of $T_e$ and $T_s$ and also data from more regions, if we are to observe patterns in lithosphere strength variations that can be related to geology (e.g. Jackson et al. 2004). Routinely determined earthquake focal depths based on teleseismic traveltime data are not reliable, and good quality depths are those confirmed by waveform analysis or by dense local seismograph networks (e.g. Chen & Molnar 1983). McKenzie & Fairhead (1997) and McKenzie (2003) critically examined the various methods for estimating $T_e$ from the relationship between gravity and topography. They found that $T_e$ estimates based on the coherence method are generally upper bounds, which are sometimes as much as an order of magnitude greater than estimates based on the transfer function itself. One purpose of this paper is to present seismogenic and elastic thickness estimates for Mongolia. The seismic data are in the form of source parameters for moderate-sized ($M_w 5.2–6.5$) earthquakes determined by teleseismic waveform analysis, and the $T_e$ estimates are based on gravity and topographic analyses in both the spectral and space domains.

The currently deforming area of Mongolia is a sizeable part ($\sim 1000\times 1000$ km$^2$) of the India–Asia collision zone, and of particular interest. It occupies a space (Fig. 1) between the NE Tien Shan, where $T_e$ and $T_s$ are unusually large ($\sim 50$ and 30 km, respectively; Maggi et al. 2000a) and the Baikal rift zone, which in its NE part also has evidence for lower crustal earthquakes (Doser 1991a; Déverchère et al. 1991, 1993, 2001) and possibly a larger elastic thickness than normal (Diamant & Kogan 1990; who used the coherence method, and whose $T_e$ estimate of 30 km is therefore an upper bound). Mongolia has experienced four earthquakes of $M_w > 8$ in the last 100 yr (Baljinnyam et al. 1993), and the overall deformation rates have now been estimated by GPS measurements (Calais et al. 2003). We will use the earthquake source parameters determined here to ask how the pattern of faulting in Mongolia achieves the motions observed in GPS. This is another reason to publish the earthquake source parameters presented here, which come from the PhD thesis of Bayasgalan (1999); some have already been quoted in the literature for seismotectonic purposes (e.g. Calais et al. 2003; Delouis et al. 2002) but have not been generally available before.

Thus we begin by discussing the earthquake source parameters, and their significance for, firstly, the mechanical properties of the lithosphere (using their depth distribution) and, secondly, the fault
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2 EARTHQUAKE SOURCE PARAMETERS

2.1 Data sources

An aim of this paper is to present what is known of the source parameters (i.e. fault plane solutions and depths) of earthquakes in, and adjacent to, Mongolia in as complete and homogeneous way as is possible, with due regard for variations in data quality. The data we present and discuss are necessarily determined from teleseismic waveforms, as until very recently, there was no effective local seismograph network in Mongolia. Local networks do exist in the Baikal region, and their data have been discussed by Déverchère et al. (1991, 1993, 2001), but these are mostly outside the region of focus here, which is shown in Fig. 2. Within Mongolia itself, the network used by Déverchère et al. (2001), with a typical station spacing of >100 km, is too sparse for well-determined focal depths (though it is much better in the Baikal region itself). The highest quality source parameters are those for 17 earthquakes determined from analysis of $P$ and $SH$ body waveforms (Table 1), which we discuss in the next section. The Harvard CMT catalogue provides mechanisms for 34 other earthquakes that were too small for us to study by our body wave inversion method (Table 2). We present their ‘best-double-couple’ solutions, in which the eigenvalue with the smallest absolute value is set to zero, while maintaining the direction of its associated eigenvector (the null axis). The extent to which the ‘best-double-couple’ solution is representative of the whole moment tensor is given by the parameter $\gamma$, such that if $\gamma = 100$ the whole moment tensor was a pure double-couple, and if $\gamma = 0$ the whole moment tensor is a compensated linear vector dipole (see the caption to Table 2). The data for the routine CMT solutions are low-pass filtered, and consequently lose the ability to determine centroid depth, for the shallow depths of interest here. In a few cases where broad-band data were used (Ekström & England 1989), the CMT depths are well determined, and have been included. Finally we include 22 earthquakes larger than $M_w$ 4.7 for which fault plane solutions are available from first motion polarities or, in some cases, from surface faulting. These provide estimates of the fault

Figure 2. Fault plane solutions of Mongolian earthquakes. Solutions constrained by body wave modelling (Table 1) are in black; Harvard CMT ‘best-double-couple’ solutions (Table 2) are in dark grey; first-motion solutions (Table 3) are in light grey. Small grey dots are earthquake epicentres from the catalogue of Engdahl et al. (1998) and later updates, from 1964 through 2002. Black numbers are centroid depths in km, for earthquakes whose depths are constrained by body waveforms. The two numbers in white boxes in the SW are centroid depths for two earthquakes without waveform constraints (Table 2), which are discussed in the text. Thin black lines are faults, taken from a variety of sources including Baljinnyam et al. (1993) and our own observations based on satellite images and topography.
Table 1. Earthquake source parameters determined by body wave modelling. Epicentres after 1963 are from Engdahl et al. (1998) and their updated catalogue. $M_w$ is the moment-magnitude, calculated from the formula: $M_w = (\log_{10} M_0 - 9.11)/1.5$, where $M_0$ is the moment in N m. The strikes, dips and rakes of the two nodal planes are $s_1$, $d_1$, $r_1$ and $s_2$, $d_2$, $r_2$. The slip vector azimuth used in Fig. 7 is $sv$. The centroid depth in km is $z$. The final column refers to the work in which the inversion is published (the last reference is the one used here): D is Doser (1991b); d is Delouis et al. 1995, 2003; C is Chen & Kao (1996); B is Bayasgalan & Jackson (1999).

| Date | Time | Lat. | Long. | $M_w$ | $s_1$ | $d_1$ | $r_1$ | $s_2$ | $d_2$ | $r_2$ | $sv$ | $z$ | R |
|------|------|------|-------|-------|-------|-------|-------|-------|-------|-------|------|-----|--|
| 1950 | 4    | 184410 | 52.00 | 101.40 | 6.9 | 100 | 75 | 0 | 10 | 90 | 165 | 100 | 14 | d,D |
| 1959 | 8    | 1470314 | 52.64 | 106.90 | 6.2 | 100 | 30 | 35 | 88 | 20 | 71 | 118 | 14 | D |
| 1965 | 2    | 043354 | 43.84 | 87.76 | 6.3 | 264 | 48 | 98 | 72 | 43 | 81 | 354 | 44 | N |
| 1967 | 1    | 001440 | 48.25 | 102.90 | 7.0 | 28 | 83 | 179 | 272 | 89 | 7 | 5 | H,B |
| 1967 | 1    | 015720 | 48.08 | 103.06 | 6.4 | 318 | 42 | 103 | 121 | 49 | 79 | 13 | 8 | H,B |
| 1970 | 5    | 171314 | 50.20 | 91.26 | 6.3 | 272 | 50 | 87 | 97 | 40 | 94 | 2 | 7 | A |
| 1974 | 7    | 193042 | 45.19 | 93.94 | 6.3 | 264 | 48 | 103 | 87 | 97 | 40 | 2 | 7 | A |
| 1987 | 9    | 215839 | 47.23 | 89.53 | 5.2 | 147 | 83 | 173 | 238 | 83 | 7 | 328 | 14 | A |
| 1988 | 7    | 001640 | 41.32 | 89.65 | 6.3 | 195 | 150 | 115 | 339 | 46 | 63 | 30 | A |
| 1988 | 7    | 037811 | 48.74 | 90.53 | 5.7 | 328 | 75 | 163 | 63 | 74 | 16 | 333 | 16 | A |
| 1991 | 5    | 033501 | 50.16 | 105.41 | 5.5 | 210 | 87 | 164 | 301 | 74 | 3 | 8 | A |
| 1990 | 6    | 124728 | 47.87 | 85.08 | 6.4 | 116 | 89 | 157 | 25 | 67 | −1 | 34 | C |
| 1990 | 9    | 091506 | 47.96 | 84.96 | 5.9 | 119 | 36 | 178 | 210 | 88 | 54 | 33 | C |
| 1991 | 12   | 090939 | 51.08 | 98.17 | 6.3 | 244 | 72 | 15 | 339 | 76 | 161 | 13 | A |
| 1995 | 6    | 010120 | 50.35 | 89.96 | 5.4 | 89 | 31 | 82 | 278 | 59 | 95 | 8 | 10 | A |
| 1995 | 6    | 230230 | 51.91 | 103.19 | 5.7 | 67 | 37 | 18 | 172 | 79 | 126 | 14 | A |
| 2003 | 2    | 173420 | 43.90 | 85.92 | 5.2 | 94 | 26 | 84 | 281 | 64 | 93 | 11 | 19 | A |

orientation, but not, of course, depth (Table 3). The first motion solutions we include have all been published before, and are the ones that we judge to be the most reliable: all of them reported the individual first-moment polarity readings, and in most cases those polarities were read on long-period WWSSN instruments by the authors reporting them. In all the maps that follow, only one fault plane solution is given for each earthquake, chosen in the hierarchy: waveform modelling, CMT ‘best-double-couple’, then first-motion solution.

2.2 Teleseismic bodywave modelling

Of the 17 earthquakes whose source parameters are listed in Table 1, ten are new body-wave inversion results carried out by us. Others, particularly those by Nelson et al. (1987), Huang & Chen (1986), Chen & Kao (1996), Bayasgalan & Jackson (1999) were determined in the same way used by us here, using essentially the same inversion algorithm. The method we used involved either digitizing the analogue long-period WWSSN records, or taking the digital broad-band records from stations of the GDSN and convolving them with a filter that reproduces the bandwidth of the old WWSSN 15–100 long-period instruments. We analysed $P$ and $SH$ waveforms at stations in the epicentral range 30–90°. We then used the MT5 version (Zwick et al. 1994) of McCaffrey & Abers’ (1988) and McCaffrey et al.’s (1991) algorithm, which inverts the $P$ and $SH$ waveform data to obtain the strike, dip, rake, centroid depth, seismic moment and the source time function, which is parameterized by a series of isosceles triangle elements of half-duration $\tau$. We always constrained the source to be a double-couple. The method and approach we used are described in detail elsewhere (e.g. Nábělek 1984; McCaffrey & Nábělek 1987; Molnar & Lyon-Caen 1989; Taymaz et al. 1991) and are too routine to justify repetition here. The 10 solutions for new earthquakes are presented in the on-line Appendix, together with a paragraph on each one concerning previous analyses, comparisons with the Harvard CMT solution (where available) and any information on the faulting in the epicentral region. In Table 1 we list the solutions we use from other authors. Of particular note are the two solutions for earthquakes in the 1950s by Doser (1991b) and Delouis et al. (2002); not many seismograms were available for these events, but there were enough to be confident that the centroid depths are approximately correct (both are $\sim$14 km) and not unusually deep.

2.3 Focal depths and seismogenic thickness

The compilation of focal mechanisms, and the centroid depths we consider reliable, are shown in Fig. 2. On this, and later, maps, the solutions constrained by body-wave modelling (black) can be distinguished from the Harvard CMT ‘best-double-couple’ solutions (dark grey) and the first motion solutions (light grey). The numbers adjacent to the black (waveform-modelled) solutions are centroid depths in km. All are uncertain by about ±4 km.

Within the central part of Fig. 2, and everywhere within the political boundaries of Mongolia, the available centroid depths are less than 20 km. Available estimates of crustal thickness in central and western Mongolia (Ionov et al. 1998; Zorin et al. 2002) are in the range 40–50 km. Thus most of Mongolia is, in this sense, typical of most active regions on the continents, where seismicity is usually confined to the upper crust.

However, the centroids in the west, on the northern and southern sides of the Junggar Basin, are deeper. In the north are two events at 33 and 34 km, reported by Chen & Kao (1996) near Lake Zaysan. In the south are two events at 44 km (Nelson et al. 1987) and 30 km (Appendix, Fig. A8). In addition, there are two events in the south whose waveforms were too small and noisy for us to check with our inversion method, but whose reported Harvard CMT depths are 65 and 56 km. These both have similar depths reported in the relocations of Engdahl et al. (1998) and their subsequently updated catalogues (Table 2). Our experience elsewhere (e.g. Jackson et al. 2002) is that, once real depths exceed about 40 km, the depths determined by Engdahl et al. (1998) and the CMT catalogue are often approximately correct, probably because the reflected $pP$ and $sP$ phases are clearer at these depths and properly identified. In this case, these three events occur where depths of $\sim$40 km are confirmed by waveform modelling for other earthquakes, so we include...
Table 2. Best-double-couple Harvard CMT solutions. This table shows the Harvard CMT solutions used in Figs 2–7, but omits those earthquakes whose source parameters are already listed in Table 1. Epicentres through 2002 are from the catalogue of Engdahl et al. (1998) and its successors. $M_w$ is the moment-magnitude, calculated as in Table 1. The strikes, dips and rakes of the two nodal planes are $s_1$, $d_1$ and $r_2$, $d_2$, $r_2$. The slip vector azimuth used in Fig. 7 is sv. $\gamma$ is the percentage double-couple component, defined from the absolute value of the intermediate eigenvalue relative to the average of the other two, normalized so that a pure double-couple source (with eigenvalues $-1, 0, +1$) is 100 per cent, while a linear vector dipole (e.g. $-0.5, -0.5, 1.0$) is 0 per cent.

| Date       | Time | Lat. | Long. | $M_w$ | $s_1$ | $d_1$ | $r_1$ | $s_2$ | $d_2$ | $r_2$ | sv | $\gamma$ | $z$ | Ref. |
|------------|------|------|-------|-------|-------|-------|-------|-------|-------|-------|----|----------|-----|------|
| 1978 10 16 | 163027 | 45.23 | 93.80 | 5.0 | 168 | 75 | 179 | 258 | 89 | 15 | 79 | 78 | H |
| 1979 8 24  | 165931 | 41.16 | 108.16 | 5.7 | 111 | 44 | 65 | 259 | 51 | -112 | 69 | H |
| 1980 11 6  | 013428 | 43.77 | 86.19 | 5.1 | 82 | 18 | 66 | 287 | 74 | 97 | 83 | H |
| 1980 12 15 | 221148 | 46.03 | 90.44 | 5.5 | 273 | 44 | 113 | 63 | 50 | 70 | 85 | 5 | E |
| 1981 5 22  | 095122 | 52.00 | 105.77 | 5.4 | 18 | 18 | -118 | 227 | 74 | -81 | 137 | 85 | H |
| 1981 8 16  | 175413 | 50.60 | 96.83 | 5.2 | 166 | 53 | 139 | 283 | 58 | 45 | 58 | H |
| 1982 8 3  | 045023 | 48.90 | 89.71 | 5.1 | 314 | 34 | 130 | 89 | 64 | 67 | 86 | 11 | E |
| 1983 3 3  | 033218 | 43.86 | 86.65 | 5.3 | 148 | 38 | 128 | 283 | 61 | 64 | 98 | 65/66 | H |
| 1983 12 15 | 105254 | 42.91 | 87.35 | 5.1 | 165 | 44 | 132 | 294 | 59 | 57 | 53 | 56/56 | H |
| 1984 4 24  | 002216 | 47.43 | 89.61 | 5.0 | 345 | 70 | 161 | 82 | 72 | 21 | 352 | 68 | H |
| 1984 11 4  | 161915 | 50.86 | 95.25 | 5.3 | 159 | 46 | 129 | 284 | 61 | 64 | 98 | 65/66 | H |
| 1986 3 30  | 152253 | 42.91 | 87.35 | 5.1 | 165 | 44 | 132 | 294 | 59 | 57 | 53 | 56/56 | H |
| 1986 11 15 | 064054 | 48.53 | 89.57 | 5.1 | 314 | 34 | 130 | 89 | 64 | 67 | 86 | 11 | E |
| 1987 8 31  | 072554 | 44.00 | 107.02 | 5.3 | 148 | 38 | 128 | 283 | 61 | 64 | 98 | 65/66 | H |
| 1988 6 30  | 152253 | 42.91 | 87.35 | 5.1 | 165 | 44 | 132 | 294 | 59 | 57 | 53 | 56/56 | H |
| 1988 11 6  | 013428 | 43.77 | 86.19 | 5.7 | 82 | 18 | 66 | 287 | 74 | 97 | 83 | H |
| 1989 1 30  | 035105 | 41.65 | 88.50 | 5.4 | 116 | 36 | 99 | 286 | 54 | 84 | 26 | 95 | H |
| 1990 2 25  | 185829 | 44.02 | 104.94 | 5.6 | 66 | 66 | 286 | 54 | -93 | 152 | 100 | H |
| 1990 5 31  | 162807 | 51.68 | 105.02 | 5.0 | 247 | 40 | -75 | 46 | 52 | -102 | 44 | H |
| 1991 3 1  | 031218 | 46.10 | 94.06 | 5.1 | 121 | 23 | 124 | 265 | 71 | 76 | 76 | H |
| 1991 7 5  | 025800 | 48.53 | 89.57 | 5.1 | 317 | 51 | 134 | 80 | 56 | 49 | 350 | 66 | H |
| 1993 9 27  | 113325 | 50.04 | 87.81 | 5.1 | 117 | 67 | 156 | 217 | 68 | 25 | 94 | H |
| 1993 10 11 | 010325 | 50.21 | 87.72 | 6.2 | 221 | 67 | 5 | 129 | 85 | 157 | 72 | H |
| 1993 10 9  | 160602 | 50.09 | 87.86 | 5.0 | 294 | 41 | 129 | 68 | 59 | 61 | 91 | H |
| 1993 10 13 | 052637 | 50.26 | 87.70 | 5.1 | 58 | 69 | 10 | 324 | 80 | 159 | 77 | H |
| 1993 10 17 | 053020 | 50.17 | 87.75 | 5.1 | 223 | 80 | -6 | 314 | 84 | -170 | 87 | H |
| 1993 10 23 | 002546 | 49.93 | 88.34 | 5.1 | 170 | 73 | -169 | 77 | 79 | -17 | 91 | H |
| 1993 11 11 | 224230 | 50.15 | 87.89 | 5.1 | 129 | 48 | 43 | 6 | 60 | 129 | 26 | H |
| 1993 11 17 | 013547 | 50.19 | 87.65 | 5.2 | 67 | 47 | 52 | 296 | 55 | 124 | 94 | H |

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whatever those basements may be. But in the rest of Mongolia, east of the Junggar Basin, there is no evidence for seismicity other than in the upper half of the crust.

3 EARTHQUAKE FAULTING PATTERNS IN MONGOLIA

In this section we give a brief overview of the styles of faulting in the regions surrounding the relatively inactive, elevated dome of Hangay in central Mongolia (Fig. 2), before discussing how this faulting achieves the motions observed in the GPS velocity field in Section 4.

3.1 Mongolian Altay

Most earthquakes within the Mongolian Altay range are consistent with right-lateral strike-slip on NW–SE faults or thrusting on roughly E–W faults (Figs 3 and 4). The area is susceptible to great earthquakes such as that in 1957 (Ms 8.3), which ruptured during the 1957 earthquake, is the most prominent active fault system within the range and is discussed in greater detail in Bayasgalan et al. (1999a) and in Section 4.

In the northern part of the mountains, near the border between Mongolia and Russia, the mountain ranges of the Mongolian Altay turn gradually into a WNW trend. In this region, many of the earthquakes involve thrust faulting (events 700515 and 950622 in Fig. 3). One event (880630) has a Harvard CMT solution showing E–W normal faulting, and is located very close to the 1970 May 5 Üüreg Nuur earthquake, which occurred on an E–W thrust (Appendix, Fig. A1). It is rare for parallel, active, normal and reverse faults to co-exist close together. However, the earthquake of 880630 was small (Mw 5.0, Mw 5.3), and the eigenvalues of the CMT solution were 11.8, −3.5 and −8.3 × 1016 N m, indicating that it is a poor double-couple solution (γ = 48 per cent). Its seismograms were too noisy to allow even one dilatational first motion to be confirmed, and too small for the inversion technique used here to be applied, so we cannot confirm the CMT solution for this apparently normal-faulting earthquake.

At the southern end of the range (Fig. 4) there are also E–W striking events (e.g. 801215), although this trend in the topography is not as prominent as in the north. The configuration of faulting, and especially the thrust-fault terminations of the strike-slip faults at both ends of the Mongolian Altay, might be related to rotations about a vertical axis; this topic is discussed in more detail in Bayasgalan et al. (1999a) and in Section 4.

3.2 Gobi Altay

The Gobi Altay mountains consist of E–W elongated ridges about 600 km long (Fig. 2). These mountains also have the potential to generate great earthquakes such as that in 1957 (Mw 8.3), which produced about 260 km of surface ruptures involving left-lateral strike-slip motion with a reverse component (Kurushin et al. 1997). The Valley of Lakes fault system (Fig. 5), the eastern part of which ruptured during the 1957 earthquake, is the most prominent active fault system within the range and is discussed in greater detail in Kurushin et al. (1997), Florensov & Solonenko (1965) and

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Table 3. First motion fault plane solutions. First motion fault plane solutions and mechanisms based on surface faulting. Epicentres after 1964 are from Engdahl et al. (1998). M is magnitude (Ms in italics, ‘magnitude’ reported by Irkutsk in bold, and Ms from Engdahl et al. (1998) in normal type). The strikes, dips and rakes of the two nodal planes are s1, d1, r1 and s2, d2, r2. The slip vector azimuth used in Fig. 7 is sv. The final column refers to the work in which the first motion polarities are published, or the surface faulting is described (indicated by an asterisk). In each case, the last reference is the one used here. O6 is Okal (1976); O7 is Okal (1977); S is Schupp (1996); M is Molnar & Deng (1984); F is Flerov et al. (1965); B is Baljinnyma et al. (1993); T is Petit et al. (1996); D is Doser (1991b).

| Date   | Time  | Lat. | Long. | M   | Ms  | s1  | d1  | r1  | s2  | d2  | r2  | sv  | R     |
|--------|-------|------|-------|-----|-----|-----|-----|-----|-----|-----|-----|-----|-------|
| 1905   | 7 9   | 094024 | 49.50 | 90.30 | 8.2 | 60  | 63  | 30  | 135 | 64  | 150 | 45   | O7,M,S |
| 1905   | 7 23  | 024612 | 49.20 | 90.60 | 8.3 | 90  | 90  | 200 | 290 | 90  | 0   | 90   | O7,M,S |
| 1931   | 8 10  | 211848 | 46.57 | 89.97 | 8.0 | 160 | 80  | 180 | 70  | 90  | −10 | 340  | M*    |
| 1957   | 2 6   | 203456 | 50.12 | 105.32 | 6.0 | 154 | 80  | −10 | 345 | 80  | 170 | 12   | D     |
| 1958   | 6 23  | 051008 | 48.70 | 103.20 | 6.2 | 1   | 70  | 151 | 102 | 63  | 23  | 12   | B,P   |
| 1962   | 1 22  | 072641 | 52.37 | 100.13 | 5.2 | 174 | 33  | 154 | 286 | 76  | 60  | 12   | D     |
| 1966   | 2 10  | 064829 | 52.56 | 106.85 | 5.0 | 245 | 45  | −68 | 35  | 49  | −110 | D    |
| 1966   | 5 10  | 210408 | 51.83 | 99.01  | 5.0 | 265 | 70  | 15   | 170 | 76  | 159 | P    |
| 1967   | 8 30  | 061034 | 51.74 | 104.64 | 4.8 | 167 | 68  | 168 | 262 | 79  | 22  | D    |
| 1967   | 2 11  | 092734 | 52.12 | 106.42 | 5.1 | 218 | 67  | 168 | 313 | 76  | 23  | D    |
| 1970   | 3 28  | 094459 | 52.24 | 105.88 | 4.9 | 351 | 86  | 146 | 84  | 56  | 25  | T    |
| 1970   | 8 13  | 192665 | 51.91 | 105.58 | 4.7 | 244 | 53  | −74 | 39  | 40  | −110 | D    |
| 1972   | 2 26  | 233107 | 50.53 | 97.14  | 5.3 | 130 | 75  | −26 | 227 | 65  | −163 | D    |
| 1972   | 8 9   | 194216 | 52.87 | 107.72 | 5.0 | 324 | 71  | −164 | 229 | 75  | −19  | D    |
| 1972   | 8 31  | 140315 | 52.32 | 95.37  | 5.5 | 295 | 75  | 90   | 115 | 15  | 90   | 25   | T    |
| 1974   | 12 18 | 075437 | 48.37 | 103.25 | 5.2 | 255 | 86  | −38 | 348 | 52  | −174 | P    |
| 1975   | 3 31  | 100525 | 46.66 | 91.25  | 5.2 | 180 | 70  | 135 | 288 | 48  | 27   | 19   | T    |
| 1976   | 4 1   | 043114 | 51.16 | 98.03  | 5.2 | 148 | 80  | 166 | 240 | 76  | 10   | P    |
| 1980   | 2 6   | 163550 | 51.75 | 105.36 | 4.8 | 217 | 80  | −96 | 68  | 12  | −59 | 127  | D    |
| 1980   | 10 2  | 011243 | 51.24 | 107.83 | 5.0 | 250 | 66  | 14   | 154 | 77  | 155  | D    |
| 1985   | 9 3   | 033309 | 52.87 | 107.08 | 4.8 | 343 | 82  | −167 | 251 | 77  | −8   | D    |

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Figure 3. Topography and fault plane solutions of the northern Mongolian Altay. Topography is taken from the GTOPO30 dataset, and illuminated from the SW. Solutions constrained by body wave modelling (Table 1) are in black; Harvard CMT ‘best-double-couple’ solutions (Table 2) are in dark grey; first-motion solutions (Table 3) are in light grey, as in Fig. 2. The date of each mechanism is given above the focal sphere (year, month, day). For clarity, some of the smaller aftershocks of the 2003 sequence near 50°N 88°E (Table 2) have been reduced in size. White circles are other earthquakes from the catalogue of Engdahl et al. (1998) and later updates from 1964 through 2002. Faults that are prominent in the topography and well expressed on satellite images are shown in black. Senses of lateral movements are shown in black arrows and are from Baljinnyam et al. (1993), Schlupp (1996) and our own field observations. Black triangles show inferred thrust faults.

Figure 4. Topography and fault plane solutions of the southern Mongolian Altay. Notations and shading conventions are the same as in Fig. 3.
Bayasgalan et al. (1999b). There are many other late Quaternary fault scarps further west in the Gobi Altay, probably produced by earthquakes in the Holocene, but there have not been many significant earthquakes within the range since 1957. Those of 881215 and 960801 both occurred in the western part and are consistent with E–W left-lateral strike-slip faulting (Fig. 5). In the south, the range is bounded by a very prominent fault, running NW–SE from 45.25°N 96.5°E to 44.2°N 100.2°E (Fig. 2), well expressed in the topography. However, there has been no recent seismic activity on this fault and no late Quaternary ruptures have been reported there either.

The area at the southern end of the NW-trending Mongolian Altay ranges has a different topographic character, with much smaller isolated mountains. Not many earthquakes have occurred in this part of Mongolia. The Tahiin Shar earthquake of 740704 (Fig. 5) is the biggest (Mw 6.4) in the last 100 yr, and produced short surface ruptures on a WSW trending fault with left-lateral slip, approximately at same place as the instrumental epicentre (Appendix and Fig. A3). This event occurred south of the point where the mountain ranges of the Mongolian Altay and Gobi Altay join. It is somewhat anomalous in that it involved left-lateral slip, consistent with the earthquakes in the Gobi Altay, but its position is closer to the Mongolian Altay mountains where right-lateral strike-slip faults are dominant (Fig. 5). It is not clear what happens in detail where the Mongolian Altay and Gobi Altay conjugate strike-slip fault systems meet. Although this is an interesting and general problem in structural geology and active tectonics, and occurs, for instance, also in eastern Iran (Walker & Jackson 2004), at the moment there is insufficient information to resolve kinematic details in this region.

3.3 Hövsgöl and the grabens of northern Mongolia

A number of moderate-sized earthquakes have been recorded in the Hövsgöl area of northern Mongolia, where three north–south grabens, known as the Hövsgöl graben system, are the most prominent expression of recent tectonic activity (Fig. 6). Surprisingly, most earthquakes west of lake Hövsgöl have strike-slip fault plane solutions, such as that of 911227. A similar first motion solution was obtained for an earlier earthquake in 760401 by Petit et al. (1996). It is not clear whether the slip in these events was left-lateral on a northeast fault or right-lateral on a northwest fault. Both events occurred near the sharp eastern edge of the Busiin Gol graben, a prominent NNE rift along the Mongolian border. Southwest of Busiin Gol is the triangular-shaped Tere-Kol basin with a sharp faulted boundary on the NW side. The earthquake of 720226 may have occurred on a SW continuation of the Tere-Kol fault, involving predominantly NE–SW right-lateral strike-slip. Further west, thrusting becomes dominant (780803 and 810816).

Other earthquakes in the region outside the grabens themselves mainly show strike-slip faulting or strike-slip with a significant thrust component. The events NE of lake Hövsgöl, towards Baikal, have strike-slip faulting mechanisms (see 500404 and 950629 in Fig. 6), associated with E–W left-lateral faulting on the Tunka fault system (Delouis et al. 2002).

3.4 Hangay

The Hangay region of central-western Mongolia is a Proterozoic continental block, reactivated in the Paleozoic, and then eroded into a peneplain surface probably in the early Tertiary (e.g. Ionov et al. 1998). It is often referred to as a ‘dome’, and some of its elevation, which reaches 4000 m, is thought to postdate the formation of the old erosion surface, which is warped (Baljinnyam et al. 1993; Cunningham 2001). But the original elevation of that surface is unknown, and it is also possible that some of its elevation is inherited from pre-Cenozoic events (Petit et al. 2002). Hangay is a site of sparse basaltic volcanism dating from about
3.5 Central Mongolia

Seismic activity is much less in the area east of Hangay (Fig. 2). The 1967 January 5 is the largest recent event in this region and was located near the NE edge of the Hangay. It produced right-lateral strike-slip ruptures on north–south faults that end in NW–SE striking thrust faulting in the south (Huang & Chen 1986; Baljinnyam et al. 1993; Bayasgalan & Jackson 1999). The 890513 earthquake at ~50°N 105°E (Fig. 2) in north-central Mongolia is located near structures that appear to be eastern continuations of the large E–W left-lateral faults which bound the northern edge of the Hangay dome. However, its source mechanism implies right-lateral, not left-lateral slip on those structures, and so is enigmatic. Another event in essentially the same place, on 570206, apparently had a similar mechanism (Doser 1991b).

4 Earthquake Slip Vectors and the GPS Velocity Field

As Huang & Chen (1986) pointed out, the active faulting in western Mongolia forms a pattern resembling a parallel megathrust system. As a consequence, the observed NNW–SSW shortening in this part of Asia (Holt & Haines 1993; England & Molnar 1997). Although the Hövsgöl graben system appears to be an extension of the Baikal rift system, the few earthquake mechanisms available in that region show strike-slip faulting. The pattern of late Quaternary faulting seen in earthquakes, especially west and south of Hangay, is also seen in the Late Quaternary geological structures (Cunningham et al. 1996a,b, 1997). Enough is therefore known of the pattern of faulting surrounding Hangay for us to be able to choose, in many cases, which nodal plane is likely to have been the fault plane in many of the earthquake fault plane solutions. The identification of the fault plane, in turn, establishes the slip vector azimuth in each earthquake.

Fig. 7 shows the slip vector azimuths in our earthquake compilation, with directions showing the motion of the west or south sides relative to the east or north. We have only included slip vectors for those mechanisms where the identification of the fault plane is either obvious (from coseismic or Late Quaternary surface faulting) or unimportant (in pure dip-slip mechanisms). For those earthquakes where the correct choice is ambiguous, we have simply omitted the slip vector, though shown the epicentre, so that Figs 7 and 2 may be compared. The slip vectors themselves are given in Tables 1–3.

It is instructive to compare these earthquake slip vectors with the GPS-derived velocities of points in the region (from Calais et al. 2003), shown in a reference frame relative to stable Eurasia in Fig. 8. In the west, the GPS velocities show the ~10 mm yr–1 NNE motion on the N side of Tien Shan rapidly decreasing in amplitude northwards as shortening occurs in the Mongolian Altay. In the east, GPS velocities are directed eastwards, increasing in amplitude to the south. These motions are achieved by earthquake slip vectors showing a marked divergence; directed NW in the Mongolian Altay and E in the Gobi Altay (Fig. 7).

A prominent feature of the comparison between Figs 7 and 8 is that the slip vectors in the Mongolian Altay are mostly in the wrong direction (NNW) to take up the GPS motions (NNE). A comparison of Figs 7 with Fig. 2 shows that the NW-directed slip vectors are from strike-slip faulting events, whose nodal planes are well constrained because they pass through the centres of the focal spheres, where teleseismic data are concentrated. The discrepancy between slip vectors and GPS directions may not, therefore, be attributed to errors in the slip vectors. This discrepancy is not an unusual situation and occurs also in other places, such as central Greece (Goldsworthy et al. 2002) and the western Transverse Ranges of California (Jackson & Molnar 1990).

The distributed deformation in Figs 7 and 8 is taken up on a number of fault-bounded blocks, and the earthquake slip vectors measure the relative motion between the blocks on either side. In contrast, the GPS vectors in Fig. 8 determine the motion of the reference points...
with respect to Eurasia. The simplest form of distributed deformation involves all the blocks moving in the same direction with respect to Eurasia at different velocities. The slip vectors for all earthquakes between the blocks must then be parallel to the GPS vectors. This behaviour is clearly too simple to account for the observations. A more complicated type of deformation occurs when the blocks have an arbitrary velocity component that is parallel to the boundaries of the deforming zone. If the resulting velocity field does not involve rotations, which is equivalent to requiring that the curl of the velocity, \( \nabla \times \mathbf{v} \), is zero, the sum of the slip vectors across the zone must give the GPS vector. Therefore, if the angles between the slip vectors and GPS vectors are positive in one part of the deforming zone, they must be negative elsewhere. In contrast, Figs 7 and 8 show that almost all the slip vectors from earthquakes in the Mongolian Altay are rotated anticlockwise with respect to the GPS vectors. The only velocity field that can rotate all relative velocity vectors in the same sense with respect to the GPS vectors is one that involves rotation about a vertical axis. The sense of the required rotation is most easily determined by using a sketch like that in Fig. 9.

The discrepancy between the slip vectors and the GPS vectors in Figs 7 and 8 requires the Mongolian Altay to be rotating anticlockwise with respect to Eurasia, as suggested by Bayasgalan et al. (1999a). A sketch of the likely geometry is shown in Fig. 9. Block rotations commonly occur in regions of distributed continental deformation, such as central Greece and the western Transverse Ranges of California. In both of those places this interpretation is supported by the rotation of palaeomagnetic declinations. In the Mongolian Altay, the only relevant palaeomagnetic data is from Thomas et al. (2002), who found no significant palaeomagnetic rotations in the Oligocene and Miocene sediments of the Zaysan basin (\( \sim48^\circ N 84^\circ E \)) on the southern edge of the Altai (Fig. 2), but a counter-clockwise rotation of \( 39 \pm 8^\circ \) in middle Miocene to lower Pliocene sediments of the Chuya basin (\( \sim50^\circ N 88^\circ E \)) in the NW Mongolian Altay. This is the region where the NW-trending right-lateral strike-slip faulting changes to E–W thrusting (Figs 2 and 3), which is the mechanism by which the strike-slip faults seem to terminate (Bayasgalan et al. 1999a). On balance, therefore, it seems to us that a counter-clockwise rotation of the Mongolian Altay relative to Eurasia is the most likely way to reconcile the earthquake slip vectors with the GPS measurements of Calais et al. (2003). By contrast, palaeomagnetic data show that the Turfan depression (\( \sim42^\circ N 88^\circ E \)) in the eastern Tien Shan has not rotated significantly relative to Eurasia since the Miocene (Huang et al. 2004). This is not surprising, since the earthquake slip vectors match the GPS velocities in this region (Figs 7 and 8).

A consequence of the counter-clockwise rotation of the Mongolian Altay is the eastward motion of Hangay region relative to Eurasia, with velocities that should increase to the south as, indeed, they are observed to do in Fig. 8. As Calais et al. (2003) point out, this motion can be taken up by a distributed E–W left-lateral shear without rotation, predominantly on the faults of Tunka, Bulnay and the Gobi Altay, with slip rates on these faults compatible with what is

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**Figure 7.** Slip vectors of earthquakes in Mongolia. The arrows show the movement of the south or west side relative to the north or east, and are given in the Tables. The shading convention is the same as for the focal mechanisms in Fig. 2. It is not always possible to resolve which nodal plane is the fault plane: in those cases, the slip vector is omitted, but the epicentre is plotted (in the same shading convention), so the reader can associate this figure with Fig. 2.
Finally, it is worth noting that the whole issue of how faulting accommodates the velocity field is better addressed through the use of earthquake slip vectors, rather than through the $P$ and $T$ axes of strain (or stress, if the principal stress axes are obtained by the inversion of many focal mechanisms; e.g. Delouis et al. 2002). The $P$ axes simply show N to NE shortening throughout the Mongolian and Gobi Altay ranges, whereas it is the divergence of earthquake slip vectors, and their discrepancy with the GPS velocities that require either rotations about a vertical axes or slip partitioning.

5 EFFECTIVE ELASTIC THICKNESS

We estimated effective elastic thickness in Mongolia using two methods; 2-D spectral analysis of free-air gravity admittance, and stacked free-air gravity profiles across a basin that has the characteristic asymmetric gravity signature of a flexed foreland basin. Both methods are described and discussed in detail in McKenzie & Fairhead (1997) and McKenzie (2003). The gravity data were kindly made available by Derek Fairhead of GETECH.

5.1 2-D admittance analysis

In this method, the comparison between topography and free-air gravity is carried out in the frequency domain in two dimensions. The region chosen is defined by the box in Fig. 1, which includes...

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Figure 10. (a) Free-air admittance versus wavelength for the box outlined in Fig. 1. The gravity and topography data are shown in the Appendix, Fig. A24. Error bars are one standard deviation. The solid line is the fit for an elastic sheet with $T_e = 6.2$ km. (b) Coherence of the free-air gravity and topography versus wavelength for the same region. Error bars are one standard deviation. (c) A plot to show how the misfit of the observed to calculated admittance varies as a function of the fraction of the total load produced by internal density variations ($F_2$). The misfit is for the best fit, allowing $T_e$ to vary; equivalent to a section along the minimum trough in Fig. 11. Although the misfit is not sensitive to $F_2$ between values of 0.15 and 0.45, showing that $F_2$ trades off against $T_e$, the actual changes in $T_e$ are small (see Fig. 11). (d) Misfit as a function of $T_e$ for a constant value of $F_2 = 0.33$; equivalent to a section through the minimum in the surface in Fig. 11 at constant $F_2 = 0.33$. Note the well-defined minimum in $T_e$. The misfit is given by $\sqrt{\frac{1}{N} \sum (\bar{Z}_o - \bar{Z}_c)^2}$, and is therefore dimensionless, as $\sigma$ has the units of mgal km$^{-1}$. If the errors are normally distributed, the misfit should have a value of unity. In this case (and in Fig. 12), the misfit is greater than one, which means the errors are not normally distributed.

The admittance, $Z(k)$ between gravity ($g$) and topography ($t$) is defined by

$$\bar{g} = Z(k)\bar{t},$$

where 2-D Fourier transforms are shown by barred variables and $k$ is the magnitude of the wavenumber. Fig. 10a shows the admittance as a function of wavelength $\lambda$ (where $\lambda = 2\pi/k$), and Fig. 10b shows the coherence $\gamma^2$, also as a function of $\lambda$, where

$$\gamma^2 = \frac{\langle \bar{g}^2 \rangle^2}{\langle \bar{g}^2 \rangle \langle \bar{t}^2 \rangle}.$$

(2)

The asterisks denote complex conjugates and the angle brackets denote the average value over a wavenumber band centred on wavenumber $k$.

At short wavelengths the admittance tends to a constant value and the coherence between free-air gravity is high, as expected when short-wavelength loads are supported elastically. As the wavelength increases to 50–200 km, the admittance and coherence both decrease as the loads are increasingly compensated by flexure, and it is the shape of this slope in admittance that is used to estimate the elastic thickness. At long wavelengths, the admittance again becomes constant, this time with a low value, as the long-wavelength loads are not supported flexurally at all. In this case the asymptote in the admittance at long wavelengths is effectively zero, rather than the value of 20–40 mgal km$^{-1}$ that is characteristic of support by mantle convection (McKenzie 1994). There is therefore no evidence, in the gravity at least, that the relatively high elevation of the Hangay dome is supported by mantle convection (or a ‘hotspot’) today.

The solid-line fit to the admittance curve in Fig. 10(a) is for an elastic sheet of thickness ($T_e$) 6.2 km. The fit is good in the region of interest, where the elastic properties of the lithosphere most affect admittance (between about 50–200 km), and Fig. 10(d) shows the misfit of the calculated admittance curve to the data as a function of $T_e$. The minimum is well defined, with the misfit increasing to twice the value at the minimum if $T_e$ is outside the range 5.0–7.8 km.

Some controversy surrounds the use of this technique, and its variants using Bouguer gravity, because of the effect of internal loads on estimates of $T_e$ (e.g. Forsyth 1985; McKenzie & Fairhead 1997). This issue is discussed exhaustively by McKenzie (2003), to which enthusiasts are referred, and does not warrant repetition here. It is
important to distinguish loads that have no topographic expression from those that produce a topographic signal. Loads with no topography are generated by erosion and sedimentation. If they are responsible for a substantial part of the gravity field, McKenzie (2003) showed that the value of $T_e$ will be overestimated if it is obtained by the coherence method of Forsyth (1985). In the absence of topographic loads, the only method of obtaining an estimate of $T_e$ is to use the shape of the gravity anomaly in the space domain. When both types of load are present, then of particular interest is the fraction ($F_2$) of the total load produced by internal density variations, as this affects the shape of $Z(k)$. Fig. 10(c) shows how the misfit of the calculated curve to the observed admittance is affected by variations in $F_2$, and that there is a broad minimum centred at $F_2 = 0.33$: that is, where 67 per cent of the load is at the surface ($F_1$). But Fig. 11 shows that, although this ill-defined contribution of $F_2$ trades off against $T_e$, there is nonetheless a well-defined minimum in $F_2 - T_e$ space, with the misfit doubling if $T_e$ is outside the range 3–10 km whatever the value of $F_2$.

The conclusion from this analysis is clearly that the average elastic thickness over the region of the box chosen in Figs 1 and A24 is small: less than 10 km and probably closer to 5 km. It also shows that there is little evidence in the long-wavelength admittance for large-scale mantle upwelling beneath Hangay today.

5.2 Flexural profiles

We also used gravity profiles across the Valley of Lakes south of the Hangay dome to estimate $T_e$, modelled, in this case, as the flexure of an infinite elastic beam, in the manner described in detail by McKenzie & Fairhead (1997) and McKenzie (2003). The Valley of Lakes (Figs 1 and 5) is a depression running for ~400 km along the north side of the Gobi Altay range (Bayasgalan et al. 1999b). Its asymmetric free-air gravity signature, with an elongated high over the mountains and a parallel low over the valley, is characteristic of flexed foreland basins, and can be modelled from the shape of the gravity profile alone, ignoring the topography. To increase the signal-to-noise ratio, we stacked profiles perpendicular to the valley in boxes identified in Figs 1 and A24, in the manner described in McKenzie (2003). The resultant stacked profile ($g_{s\alpha}$) is shown in Fig. 12(b), with the solid line indicating the average, and the grey band showing the ±1σ range. The modelled gravity ($g_c$), as a function of $T_e$, is then fit to the observed by minimizing the function $H^2$:

$$H^2 = \frac{1}{N} \sum_{i=1}^{N} \left[ \frac{(g_{s\alpha} - g_{c})}{\sigma} \right]^2,$$

where $N$ is the number of points at which $g_{s\alpha}$ and $g_{c}$ are available. The misfit is shown as a function of $T_e$ in Fig. 12(a). Although the formal minimum is 19 km, this is clearly not well resolved, but the lower limit is about 15 km. It is interesting that this minimum value of $T_e$ is more than the value of 6 km found in the admittance analysis (Fig. 10), though that was averaged over a much larger area. It appears that the bulk of the Hangay region may have a slightly lower value of $T_e$ than its flank adjacent to the Gobi Altay.

6 DISCUSSION

6.1 Seismogenic and elastic thicknesses

A principal motivation for this study was to compare elastic and seismogenic thicknesses in Mongolia, and their relevance for lithospheric strength. From this perspective, the deforming region of Mongolia is quite typical of most other young or active tectonic
though, from the perspective of faulting that stretches from the Tien Shan through to Siberia. Al-
comparable to $T_s$ a slightly larger value. The gravity analysis thus confirms that $T_e$ analysis does not require that
pattern of $T_s$ and $T_e$ is well resolved then $T_e > T_s$. This is in agreement with the
region on the continents. In all the deforming belts surrounding the relatively aseismic Hangay dome, there is no evidence for earth-
quake centroid depths greater than 20 km. The spectral analysis of admittance gives a well-resolved average effective elastic thickness of
$< 10$ km over the broad region. The analysis of the flexural profiles across the Valley of Lakes constrains $T_e$ less well, but indicates
a slightly larger value. The gravity analysis thus confirms that $T_e$ is comparable to $T_s$: where $T_e$ is well resolved then $T_e < T_s$, and the
analysis does not require that $T_e > T_s$. This is in agreement with the pattern of $T_e$ and $T_s$ values reported by Maggi et al. (2000a).

Mongolia is part of a continuous zone of active seismicity and faulting that stretches from the Tien Shan through to Siberia. Al-
though, from the perspective of $T_e$ and $T_s$, Mongolia itself appears to be typical of most continental areas, on both sides of it are regions where the seismogenic and elastic thicknesses are substan-
tially greater. To the west, the seismogenic thickness of at least 45 km on the margins of the Junggar basin and Kazakh platform is
matched by an elastic thickness of at least 30 km in the region of Lake Balkhash (Maggi et al. 2000a). To the NE, earthquakes occur to depths of 30–40 km NE of Lake Baikal in the Amut and Muya areas, and on the edge of the Siberian shield (Doser 1991a; Déverchère et al. 1991, 1993, 2001), where the elastic thickness may also be correspondingly larger (Diament & Kogan 1990). The geological characteristics responsible for this contrast in mechanical

properties between the ‘typical’ young orogenic belts like Mongo-
lia and the regions to its west and east are not clear. The greater elastic and seismogenic thickness and, by implication, the greater mechanical strength, are sometimes associated with ancient Pre-
cambrian shields, as in parts of East Africa (Foster & Jackson 1998) and northern India (Maggi et al. 2000a), perhaps because the lower crust is composed of anhydrous granulite in those regions (Jackson et al. 2004). In this respect, the deeper earthquakes in NE Baikal, adjacent to the Siberian shield, are no surprise. Priestley & Debye (2003) show that the Siberian shield has an unusually thick ($\sim 200$ km) high-velocity lid that is typical of Archean shields in other regions, such as the Kapvaal, where mantle nodule data also confirm the relatively large lithospheric thickness (Qiu et al. 1996). The Kazakh platform to the west, however, is not a shield, though it has been stable since the Late Palaeozoic (Şengör & Natal’in 1996). The exact nature of its basement is unknown, though it is interesting to note that the thick high-velocity lithospheric lid detected by Priestley & Debye (2003) under the Siberian shield continues west beneath the Kazakh platform.

6.2 Hangay

The Hangay region remains a puzzle. There is no evidence in the admittance of the gravity and topography for convective support of the Hangay ‘dome’, in the form of a ‘hot spot’. Yet there is Phocene-
recent basaltic volcanism at the surface, in crust that is not thought to be anomalously thin, and the most obvious way of generating melt in such circumstances is through the upwelling of hot mantle. A thin ($\sim 70$ km) lithosphere beneath Hangay is consistent with the geochemistry of mantle xenoliths in the basalts (Ionov et al. 1998). Both Villaseñor et al. (2001) and Friederich (2003), using fundamental-mode surface-wave tomography, find low velocities in the mantle at $\sim 100$ km depth. Priestley & Debye (2003), with the increased vertical resolution of higher-mode surface waves, confirm low $S_v$ velocities beneath Hangay at $< 125$ km depth, but find higher velocities at greater depths. It is possible to produce a thin hot litho-
sphere by stretching, but, unless the mantle below is hot, stretching thins the crust as well (for which there is no evidence at Hangay) and causes the surface to go below sea level, instead of becoming elevated. Alternatively, the mantle can be made hot if it is above a plume, but there is no evidence for this in the gravity. The current elevation of Hangay is more likely to be caused by underplating, and thicker crust, which may also be why it is barely deforming compared with its surroundings. Petit et al. (2002) discuss this problem further. A detailed analysis of how the geochemistry of the Hangay mantle nodules constrains the geotherm, using new techniques that include conductivity as a function of temperature (McKenzie et al. 2005), and of how that compares with higher-mode surface wave tomography, will be the subject of a future paper. It is beyond the scope of this one.

7 CONCLUSIONS

Throughout the deforming region of Mongolia, which surrounds the relatively aseismic and elevated Hangay dome, all the reli-
bly determined centroid depths for moderate-sized earthquakes are less than 20 km. Where it is well determined by spectral analy-
isis, the elastic thickness is less than 10 km, and thus $T_e$ is com-
parable to, and probably less than, $T_s$. This result conforms to the pattern seen in other deforming regions today. Both west and east of Mongolia there is evidence for significantly larger values
of both $T_1$ and $T_2$, with $T_1$ including the whole crustal thickness, but still with $T_2 < T_1$, where it has been well determined. In the NE, these larger values may be associated with proximity to the Siberian shield. In the west, they are associated with the relatively unknown basements of the Junggar Basin and Kazakh platform. There is no evidence in the present-day free-air gravity admittance for convective support of the elevated Hangay dome. Some convective feature was presumably active beneath Hangay when the volcanism was active, but there is no evidence of its presence today.

The earthquake source parameters compiled here show how the faulting in Mongolia accommodates the $\sim 10$ mm yr$^{-1}$ NNE convergence between the Tien Shan and stable Eurasia. The discrepancy between earthquake slip vectors and GPS velocities in the west of Mongolia, in the Mongolian Altay, suggests that the NW-trending right-lateral strike-slip faults, which dominate that range rotate counter-clockwise relative to Eurasia. This is not the only possible interpretation, but it is consistent with observed palaeomagnetic rotations of Upper Miocene to Pliocene rocks, and also explains the N–S gradient in left-lateral shear across central Mongolia. Once again, this demonstrates the important information that is revealed by a comparison of earthquake slip vectors and GPS velocities, which is not revealed by the strain ($P$ and $T$) or stress axes alone.

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APPENDIX A

In this Appendix we present the observed and synthetic P and SH waveforms from the inversions for the source parameters of the ten earthquakes marked by the letter A in the last column of Table 1.

![Figure A1. P and SH waveforms for 700515 Üüreg Nuur event.](image1)

This study

272/50/87/7/3.242E18

Tapponnier & Molnar (1979)

256/60/50/13/1.918E18

![Figure A2. The minimum misfit solution for the 700515 Üüreg Nuur event compared with Tapponnier & Molnar's (1979) first motion fault plane solution.](image2)

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Figure A3. $P$ and $SH$ waveforms for 740704 Tahiin Shar event. The display convention is the same as in Fig. A1. The gains of some stations were changed where regular anomalous gains were reported. For $P$ waves the gain of station DAV was changed from reported 1500 to real 3000, and for the $SH$ waves of station MAT from 3000 to 1500, POO from 1500 to 3000. These anomalous stations were reported by Baker (1993) and Foster (1997).

We also include comments on, and comparisons with, other inversions for these earthquakes, including those of the Harvard CMT catalogue.

In general the source velocity structures in the epicentral areas are not known. For shallow events with depth less than 10 km, a half-space with $V_p = 6.5$ km s$^{-1}$ was used. For events deeper than about 10 km, we used a representative velocity above the source of $V_p = 6.0$ km s$^{-1}$, with the source in a half space of $V_p = 6.8$ km s$^{-1}$.

A.1 1970 May 15 (700515), $M_w$ 6.3

This earthquake occurred in NW Mongolia and is known as Üüreg Nuur earthquake (Figs 2 and 3). Khilko et al. (1983) briefly reported
Figure A4. The top two lines show the contribution of each subevent to the seismograms of the 740704 Tahiin Shar event. The first subevent was smaller (see $M_c$) and probably deeper than the second one. Line 3 shows a solution with the second subevent offset the same distance (9 km) as in line 2, but in the opposite direction ($030^\circ$). The display convention is the same as in Fig. A2.

Figure A5. Comparison of the minimum misfit solution in Fig. A3 with Huang & Chen’s (1986) solution. The display convention is the same as in Fig. A2.

870918 western M. Altay

Figure A6. $P$ and $SH$ waveforms for the 870918 event in the western Mongolian Altay. The display convention is the same as in Fig. A1. A $\$ sign next to the station code letter indicates that the ‘long-period’ data displayed here have been created by convolving the GDSN broad-band records with a filter that reproduces the bandwidth of the old WWSSN 15-100 long-period instrument, as described in the main text.
This study
A. Bayasgalan, J. Jackson and D. McKenzie

Figure A7. Comparison of the minimum misfit solution for the event 870 918 with the Harvard CMT solution. The display convention is the same as in Fig. A2.

871222 Tien Shan
195/50/115/30/9.741E16

Figure A8. P and SH waveforms for the 871 222 event in the eastern Tien Shan. The display convention is the same as in Fig. A1.

This study
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Figure A9. Comparison of the minimum misfit solution of the event 871 222 with the Harvard CMT solution. The display convention is the same as in Fig. A2.

the short surface ruptures produced during this earthquake, later re-examined by Baljinnyam et al. (1993). The ruptures consist of a spectacular zone of cracks striking E–W for a distance of 6–8 km (Baljinnyam et al. 1993). They resemble the complex deformation sometimes seen in the hanging wall of a thrust faults, similar to those in the 1981 El Asnam earthquake in Algeria (Philip & Meghraoui 1983). Apparently the main thrust fault did not reach the surface, but it is quite possible that it has not yet been found, as the area is very remote.

We had access to only six seismograms (4 P waves and 2 SH waves) for this event. The minimum misfit solution for this event is shown in Fig. A1, and shows reverse faulting with an E–W strike. It is obviously not well constrained, but the P waveforms are sufficient to constrain the centroid depth to be shallow (7 ± 4 km).
871222 Centroid Depth Tests

Figure A10. Tests showing the effect of varying the centroid depth on the synthetic seismograms of event 871222. The minimum misfit solution is in line C. Each line shows the P waveforms at CTAO, NWAO and KEV, and the SH waveforms at BCAO and KEV. In each case the centroid depth is fixed at the value shown in bold. As the centroid depth decreases, the source time function broadens in order to fit the shape of the waveform. As the depth is increased the rupture duration decreases because reflected and converted phases then broaden the waveform.

880723 Tsambagarav
328/75/163/16/4.89E17

Figure A11. P and SH waveforms for the 880723 Tsambagarav event in the northern Mongolian Altay. The display convention is the same as in Fig. A1. A $ sign next to the station code letter indicates that the ‘long-period’ data displayed here have been created by convolving the GDSN broad-band records with a filter that reproduces the bandwidth of the old WWSSN 15-100 long-period instrument, as described in the main text.

Tapponnier & Molnar’s (1979) first-motion solution had some right-lateral strike-slip component, but this is not required by the first motions themselves. Line 1 in Fig. A2 shows the waveforms for the minimum misfit solution. In line 2 the strike, dip and rake were held fixed at Tapponnier & Molnar’s (1979) first-motion solution, with other parameters left free in the inversion. The resultant fit of the waveforms is significantly poorer than in line 1. Thus, in spite of the few seismograms available for this earthquake, its focal mechanism appears to be dominantly reverse faulting with an E–W strike. The ruptures seen at the surface are consistent with hanging wall deformation above a buried thrust or reverse fault of this orientation.

A.2 1974 July 4 (740704), $M_w$ 6.4

This earthquake occurred in SW Mongolia and is known as Tahiin Shar earthquake (see Figs 2 and 5). Khilko et al. (1983) described
This study

Figure A12. Comparison of the minimum misfit solution of the event 880 723 with the Harvard CMT solution. For the Harvard CMT solution the strike, dip, rake and depth were fixed, and $M_o$ was free to vary. The $M_o$ decreased from the reported CMT value $9.0 \times 10^{17}$ N m at a depth of 18 km. The display convention is the same as in Fig. A2.

the coseismic ruptures produced during this event as a system of ENE ruptures with an overall length of more than 17 km. The width of the zone reaches up to 15 m. The configuration of en-echelon mole-tracks and tension gashes suggests left-lateral strike-slip displacements of typically 0.3–0.4 m. Along the base of a hill, vertical displacements up to 0.4–0.5 m were also observed, with the southern end up. Most likely Huang & Chen (1986), who modelled the $P$ waves of this earthquake, did not know of the existence of these ruptures. They came to the conclusion that the event consisted of two almost pure strike-slip subevents separated by 3 s, but as their modelling suggested that the second event was located 12 km south of the first event, they concluded that the event occurred on a N–S right-lateral fault.

The results of our modelling (Fig. A3) are similar to those of Huang & Chen’s (1986), in that we also find two subevents separated by 3 s, both with pure strike-slip motion. However, we find an adequate fit to the seismograms with the second subevent about 9 km to the southwest (210°) of the first, consistent with SW propagation of a left-lateral strike-slip rupture, which is more consistent with the surface ruptures.

The first two lines of Fig. A4 show the contribution of each subevent to the waveforms. In the third line, the solution is held fixed to the orientation and time function in line 2, but with the second subevent offset 9 km to the NE (030°) instead of SW. Although the fit to the seismograms is marginally worse at some stations, the offset of the second subevent is not well resolved by the data. The offset to the SW, used to produce Fig. A3, was chosen to be consistent with the surface ruptures, but is not required by the seismological data.

Fig. A5 shows a comparison of the minimum misfit solution in Fig. A3 with the solution reported by Huang & Chen (1986). The $P$ waveforms are fit quite well by both solutions, but the $SH$ waveforms (which Huang & Chen did not use) at SNG and GDH are matched rather better by the solution in Fig. A3 than by Huang & Chen’s (1986).

A.3 1987 September 18 (870918), $M_o$ 5.2

This earthquake occurred in the western Mongolian Altay, near the northern end of the Fu Yun Fault, which produced a $M_s > 8$ earthquake on 1931 August 10 (Figs 2 and 4). The minimum misfit solution is pure strike-slip event with right-lateral motion on NW plane (Fig. A6). This was the first event near Mongolia to be recorded by...
This study

HRV CMT

Figure A14. Comparison of the minimum misfit solution of the event 890513 with the Harvard CMT solution. Display convention is the same as in Fig. A2.

911227 Busiin Gol

Figure A15. $P$ and $SH$ waveforms for the 911 227 earthquake west of lake Hövsgöl in northern Mongolia. The display convention is the same as in Figs A1 and A11.

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Figure A16. Comparison of the minimum misfit solution of the event 911 227 with Harvard CMT solution. Display convention is the same as in Fig. A2.

950622 Tsagaan Shuvuut
89/31/82/11/1.576E17

Figure A17. P and SH waveforms for the 950 622 earthquake in the northern Mongolian Altay, south of the east–west Tsagaan Shuvuut range. Display convention is the same as in Figs A1 and A11.
The synthetic $SH$ seismogram for the CMT source orientation is a worse fit to the observed seismogram at COL, but similar at other stations.

A.4 1987 December 22 (871222), $M_w$ 5.3

This earthquake occurred in eastern Tien Shan (Fig. 2). The minimum misfit solution shows thrust faulting with a roughly N–S strike (Fig. A8).

Our minimum misfit solution is very similar in orientation to the Harvard CMT solution (Fig. A9). However, the depth is greater in our solution (30 km against 19 km) with a smaller scalar seismic moment of $M_w$ (0.97 \times 10^{17}$ N m, compared with 2.1 \times 10^{17}$ N m). To see whether our solution is correct, we did a depth test, holding the depth fixed at 10, 20, 30 and 40 km and allowing the other parameters to change freely during the inversion (Fig. A10). The main effect is that the source time function in the minimum misfit solutions at 10 and 20 km depth have two separate pulses, where a single pulse is required at greater depths. Patton (1998) analysed the source mechanisms of several earthquakes in Tien Shan using the amplitude spectra of regional Love and Rayleigh waves recorded at WMQ (Urumqi), and for this event he estimated the focal depth as 20 km.

Unfortunately, for this interesting event the available digital seismograms for $P$ and $SH$ are rather limited. Most of the stations have only a long-period response with little high-frequency content, and are not very sensitive to depth (Fig. A10). On the grounds of simplicity, and the marginally better fit at BCAO (SH) and KEV (SH), the short time function and a depth of 30 km is perhaps preferable, though we can not rule out a shallower source ~20 km with a more complex time function, which would be in closer agreement with Patton’s (1998) results. Note, however, that this earthquake occurred in a part of the Tien Shan where at least one earthquake has a confirmed depth of ~40 km (Fig. 2; Nelson et al. 1987).

A.5 1988 July 23 (880723), $M_w$ 5.7

This earthquake occurred in the Mongolian Altay and is known as the Tsambagarav earthquake (Figs 2 and 3). The minimum misfit solution shows right-lateral strike-slip faulting on the northwest plane (Fig. A11). This earthquake is located in the central part of the Hovd fault system on the eastern edge of the Mongolian Altay range. A Holocene fault scarp was reported just north of the epicentre by Khilko et al. (1983) and Baljinnyam et al. (1993). The west-facing scarp varies in height from 0.5 to 2.0 m and the
configuration of tension cracks suggests right-lateral slip on the NNW-SSE nodal plane (Fig. 3). The vertical component of slip on the scarp is anomalous, in that it is up on the east side, opposite to the overall topography.

The minimum misfit solution is very similar to the Harvard CMT solution. The minimum misfit solution fits the $SH$ waves at TOL and KEV rather better, though the $P$ waveform at KMI is better fit by the Harvard CMT solution (Fig. A12).

A.6 1989 May 13 (890513), $M_w$ 5.5

This event occurred in northern Mongolia near $50^\circ$N $105^\circ$E (Fig. 2) and is known as the B"uteeliin earthquake. This is an almost pure strike-slip event (Fig. A13), but it is not clear which nodal plane is the fault plane. Although, the B"uteeliin range is located on the eastern continuation of the left-lateral Bulnay fault, which produced the great 1905 earthquake, the NE-striking nodal plane for 890513...
This study

HRV CMT

Figure A21. Comparison of minimum misfit solution of the event 950 629 with Harvard CMT solution. Display convention is the same as in Fig. A2.

030213 Junggar basin

94/26/84/19/8.095E16

Figure A22. P and SH waveforms for the 030 213 earthquake on the southern margin of the Junggar Basin. The display convention is the same as in Figs A1 and A11.

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Figure A23. Comparison of minimum misfit solution of the event 030 213 with Harvard CMT solution. Display convention is the same as Fig. A2. In line 2, the strike, dip, rake and depth were fixed to the values of the Harvard solution. Only the time function was free to change.

Figure A24. Free-air gravity (top) and topography (bottom) of the region outlined in the rectangular box in Fig. 1, used for the 2-D spectral analysis of the admittance and the estimate of $T_e$ in central Mongolia. The data were provided by Derek Fairhead of GETECH, and are described in more detail in McKenzie & Fairhead (1998).
is right-lateral, not left-lateral. Both our minimum misfit solution and Harvard CMT solution require either right-lateral slip on a NE plane or left-lateral slip on a NW fault plane. An earlier earthquake (570206) occurred close to this Buteeliin event and has very similar first-motion fault-plane solution, reported by Petit et al. (1996) (Table 3 and Fig. 2).

The reported Harvard CMT seismic moment was $3.6 \times 10^{17}$ Nm with a fixed depth at 15 km (Fig. A14). An inversion with the strike, slip and rake fixed to the Harvard CMT solution finds a minimum misfit at the depth 4 km with a seismic moment of $1.9 \times 10^{17}$ and a complex source time function with three separate impulses, but the fit to the P waveforms at MAJO and KEV is not good. Our minimum misfit solution (Fig. A13) has a centroid at 8 km depth with a scalar moment smaller than the reported Harvard CMT moment, and with a slightly different strike ($+4^\circ$), dip ($+13^\circ$) and rake ($-10^\circ$).

A.7 1991 December 27 (911227), $M_w$ 6.3

This earthquake occurred in northern Mongolia, west of lake Hövsgöl (Figs 2 and 6). It is located close to the Busiin Gol graben, and the minimum misfit solution is almost pure strike-slip (Fig. A15).

The Harvard CMT solution reported a scalar seismic moment of $3.8 \times 10^{18}$ Nm with a fixed depth at 15 km. The inversion with the strike, dip and rake fixed to the Harvard CMT values gives a depth of 12 km and a slightly larger seismic moment of $4.3 \times 10^{18}$ Nm. Our minimum misfit solution with 13 km depth and seismic moment at $3.9 \times 10^{18}$ Nm is only slightly different in strike ($-2^\circ$), dip ($-8^\circ$) and rake ($-7^\circ$) from the Harvard CMT solution (Fig. A16).

A.8 1995 June 22 (950622), $M_w$ 5.4

This earthquake occurred in the northern Mongolian Altay (Figs 1 and 3) and involved thrusting on an east-west trending fault (Fig. A17), similar to the Üüreg Nuur event of 1970 May 15. The minimum misfit solution is also confirmed by the unfiltered broadband records (Fig. A18), although these seismograms were not included in the inversion process.

Harvard reported a CMT solution with a scalar seismic moment of $1.7 \times 10^{17}$ Nm at a fixed depth at 15 km. An inversion with the strike, dip and rake fixed to the Harvard CMT values (Fig. A19) obtains a depth of 12 km with a slightly smaller seismic moment ($1.5 \times 10^{17}$ Nm). Our minimum misfit solution in Fig. A17 and the Harvard CMT solution are not significantly different (Fig. A18).

A.9 1995 June 29 (950629), $M_w$ 5.7

This earthquake occurred in the E–W Tunka depression, within the southern Baykal region (Figs 2 and 6). If the ENE–WSW nodal plane is the fault plane, the minimum misfit solution implies left-lateral strike-slip on a fault dipping 40° SSE (Fig. A20).

The Harvard CMT solution is similar to our minimum misfit solution, though the CMT depth which was fixed at 15 km and the moment was $5.2 \times 10^{17}$ Nm. An inversion with the strike, dip and rake fixed to the Harvard CMT values shows little change in depth and scalar seismic moment from our minimum misfit solution (Fig. A21).

A.10 2003 February 13 (030213), $M_w$ 5.2

This earthquake was a thrust, on the southern margin of the Junggar basin (Fig. 2). The waveforms are sufficient to show that the depth was $\sim 19$ km (Fig. A22). This is substantially shallower than the reported Harvard CMT depth of 45 km, which does not provide a satisfactory fit to the waveforms (Fig. A23). The earthquake is $\sim 200$ km west of one with a confirmed depth of 44 km (Fig. 2, Table 1), and in the same region as two other earthquakes in 1983 reported at 55–65 km by Harvard and Engdahl et al. (1998), marked in Fig. 2 and identified in Table 2. The waveforms from these two events were, however, too small for us to verify their depths independently.