The field of crustal velocity in Asia calculated from Quaternary rates of slip on faults

Philip England¹ and Peter Molnar²

¹Department of Earth Sciences, Oxford University, Oxford OX1 3PR, UK. E-mail: philip.england@earth.ox.ac.uk
²Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139, USA

SUMMARY

Rates of Quaternary faulting, when treated as measures of strain rates within regions several times larger than the depths of brittle faulting, yield a velocity field for Asia that matches the average relative velocity between the Indian and Eurasian plates over the past 2 Myr (NUVEL-1A), and the velocity of Shanghai with respect to Eurasia, measured with Very Long Baseline Interferometry (VLBI). By using rates of slip on discrete surfaces (faults) to define average strain rates, we explicitly assume that the large-scale deformation can be treated as that of a continuum. Depending on the scales over which strain rates are averaged, 87 to 93 per cent of the assumed strain rates are fit within assigned 1-sigma uncertainties, and 97 to 99 per cent are fit within 2-sigma. The internal consistency of these data and the agreement with independent rates (NUVEL-1A and VLBI) test the validity of treating large-scale deformation of continents as that of a continuum. In the calculated velocity field, a large fraction, ~85 per cent, of India's convergence with Eurasia is absorbed by crustal, or lithospheric, thickening. Among consistently misfit assumed strain rates are those dominated by high rates of strike-slip faulting. Strain associated with 20-30 mm a⁻¹ of slip on the Altyn Tagh, Karakax, and Karakorum faults is inconsistent with the distributions of strain in surrounding regions. We conclude that these rates, all of which are based on correlations of ages of offset landforms with changes in global climate, and not on radiometric dating, have been systematically overestimated. Lateral transport out of India's northward path is relatively minor; South China is calculated to move east-southeast at less than 10 mm a⁻¹.

Key words: Asia, Quaternary, tectonics.

1 INTRODUCTION

Plate tectonics owes much of its success to the simplicity of rigid-body displacements of a small number of plates and, equally importantly, to the clear separation of the kinematic description from the dynamic processes governing the relative motions of the plates (e.g. McKenzie 1972). Continental deformation has proved less tractable than plate tectonics, first because the kinematics is not as simple as that of the few rigid plates, and secondly because the kinematics cannot be separated from the dynamics (e.g. Molnar 1988; England & Jackson 1989).

Because a major goal in continental deformation is to understand the dynamic processes, a necessary first step is to obtain measurements of the velocity field that both span the regions of interest and sample the deformation densely enough to identify the principal velocity gradients in a region. In small and accessible regions this goal may now be achieved by space geodetic techniques. One of the most important regions of continental deformation is Asia, not only because of its size, but also because there are indications that the present deformation of the region is linked to gradients of the gravitational potential energy of the lithosphere (e.g. England & Houseman 1986; Molnar & Tapponnier 1978; Molnar et al. 1993). The size and inaccessibility of Asia, however, mean that space geodetic measurements of the requisite scale and completeness are a distant prospect. The Quaternary faulting in Asia has been extensively studied, and estimates of slip rates, on the major faults at least, are constrained to a degree that justifies an attempt to estimate the velocity field of the region from these slip rates.

Our goal here is to obtain a measurement of the velocity field within Asia from observations of Quaternary faulting of the region (Fig. 1). A few of these faults are major, in the sense that their lengths (500 km or more) are not small with respect to the length scale of the deformation in Asia (~2500 km) and
their slip rates (10 mm a⁻¹ or greater) are not small with respect to the speed of India with respect to Asia (~50 mm a⁻¹).
There are many smaller faults in the region whose lengths, nonetheless, lie in the range 100–500 km, and whose slip rates are between 1 and 5 mm a⁻¹; presumably there are many more faults whose slip rates and lengths are smaller still.

The contribution of an individual fault segment to the strain of a region is proportional to its equivalent moment rate (or to the product of its slip rate with its area). The cumulative equivalent moment rate for all the faults used in this study (see Tables in Appendix) is about $2.5 \times 10^{20}$ N m a⁻¹. About $1.4 \times 10^{20}$ N m a⁻¹ is released along the Himalayan front and on the major strike-slip fault in Burma. These faults are all long, and are slipping at rates in excess of ~20 mm a⁻¹, and are major faults, in the sense just defined. The $1.1 \times 10^{20}$ N m a⁻¹ of moment rate released within Asia north of the Himalaya, however, includes slip on a mixture of large and small faults; approximately $0.4 \times 10^{20}$ N m a⁻¹ is released on faults slipping faster than 10 mm a⁻¹, but $0.7 \times 10^{20}$ N m a⁻¹, or 60 per cent of the total strain of the interior of Asia, takes place on the smaller faults, which are slipping at less than 10 mm a⁻¹. A full description of the velocity field must allow for slip on the smaller faults.

One approach to continental kinematics has been to emphasize the major faults and neglect the deformation that may be occurring either aseismically or on minor faults and thus to approximate the behaviour of the regions between the major faults as rigid (e.g. Avouac & Tapponnier 1993; Peltzer & Saucier 1996). Another approach has been to treat the deformation as being continuous, though heterogeneous, and to use faults to assign average strain rates to individual domains within deforming continental regions (e.g. Holt et al. 1991, 1995; Holt & Haines 1993).

Each approach has its drawback. To treat continents as rigid plates is to ignore the deformation that certainly does occur between major faults. Equally, continental deformation cannot be treated as strictly continuous because the dimensions of the major faults, in Asia at least, approach in scale the dimensions of the deforming region itself. The severity of either drawback depends, however, on the goal of the kinematic analysis. Our eventual aim is to investigate the dynamics of the deformation, which are governed by the relationship.
between the deviatoric stress required to deform the lithosphere and the contrasts in gravitational potential energy between lithospheric columns (e.g. Artyushkov 1973; England & McKenzie 1982; Molnar & Lyon-Caen 1988). This relationship should be simple, if analysed on a scale that is large compared with the thickness of the lithosphere, whereas it may well be obscured by detail if it is observed on a smaller scale (e.g. England & Jackson 1989). Our goal here, therefore, is to form estimates of the velocity field within Asia in which variations in velocity over distances of hundreds of kilometres are well resolved. The cost we must pay for this approach is that variations over a few tens of kilometres may not be well resolved.

Clearly, we could obtain such a velocity field by treating the deformation in Asia as though it occurred only on a few major faults, separated by rigid blocks (e.g. Avouac & Tapponnier 1993; Peltzer & Saucier 1996). This treatment would, however, be exactly equivalent to plate tectonics, which separates the kinematics from the dynamics, and precludes analysis of the distributed deformation between the major faults (see Fig. 1). We are obliged to use a continuum approach to analyse the distributed deformation of the region both because we do not yet know the details of the deformation on the minor faults in Asia (or any other deforming region), and because a continuous description is required to relate the distributed deformation to the stresses that drive it.

Our approach follows that of Haines (1982), who showed that if the (asymmetric) strain rate tensor is known everywhere throughout a region, it can be integrated across the region to yield the spatially varying velocity field, without knowledge of the (asymmetric) spin tensor. Subsequently, Haines, Holt, and their colleagues exploited this result by using the seismic moment tensors of earthquakes, or Quaternary slip rates on faults, to calculate internally consistent velocity fields of active regions (e.g. Haines & Holt 1993; Holt & Haines 1993; Holt et al. 1995; Jackson et al. 1992, 1995). In this paper we apply a modification of Haines’ and Holt’s methods to the Quaternary deformation of Asia. We employ a discretization that is designed to deal both with rapid spatial variations in strain rate, associated with major faulting, and with distributed deformation in between faults.

We first present the method, and test it by showing that it can recover velocity fields from their associated strain-rate fields. We test the method both against a continuous velocity field and against a plate-like velocity field. Haines’ (1982) result depends on the fact that the strain in one region must be consistent with the strains in surrounding regions, in the sense of St Venant compatibility (Malvern 1969, pp. 183–186). A field of strain rate is compatible, or internally consistent, if the path integral of strain rate between any two points yields the same relative velocity between the points; otherwise, the strain rates are inconsistent. The definition of compatibility continues to apply even when strain rates over finite regions are replaced by velocity differences over faults [a point that underlies Wilson’s (1965) recognition of transform faulting].

Although the distribution of deformation in a deforming continental region must be internally consistent, in the sense described above (otherwise voids would appear), there is no guarantee that estimates of strain rate will be internally consistent, and in general they will not be. We therefore next investigate the sensitivity of the method to internally inconsistent input strain rates. We determine the conditions under which such strain rates may contaminate a solution for the velocity field, and conclude that those conditions probably do not hold for the data set of Quaternary slip rates in Asia.

We then use the method to estimate the velocity field in Asia. Our data are the slip rates on major faults of the region. Our purposes are: to determine whether these rates, obtained piecemeal from widely distributed observations, form a self-consistent description of the deformation when put together; to estimate from these data the velocity field in Asia at the scale of approximately 500 km; and to compare this velocity field with fields predicted by different tectonic hypotheses for the region. In particular we shall place bounds on the rate of eastward transport of material in Asia, which is a topic of considerable disagreement among differing views of the deformation (e.g. Armijo, Tapponnier & Han 1989; Avouac & Tapponnier 1993; Davy & Cobbold 1988; England & Houseman 1986; England & Molnar 1990; Houseman & England 1993; Molnar & Gipson 1996; Peltzer, Tapponnier & Armijo 1989; Tapponnier, Peltzer & Armijo 1986).

2 OBSERVATIONS OF, AND UNCERTAINTIES IN, SLIP RATES

To infer strain rates within elements, we used measurements of various kinds and averaged over periods of time that vary from as short as a few years (from geodetic measurements, using GPS) to as long as 10–20 Myr, in one case. Moreover, in many cases we used information provided by earthquakes as a guide to styles of deformation and orientations of faults. Appendix A contains a tabulation of the data used, including a brief discussion of each rate. In all cases, entries consist of endpoints of planar faults, orientations of the fault and of the slip vector in the fault plane, and an average rate of slip.

Uncertainties in the strikes of faults are small, because the length of each fault segment is large compared with the uncertainty in the relative position of its ends. We allot an uncertainty of 5° to the strikes of the fault segments. Most major faults of the region are either nearly pure strike-slip, with dips of ~90° (for which an uncertainty of 15° would be generous), or are nearly pure dip-slip, for which the fault-plane solutions of earthquakes commonly show 30° < dip < 60°. We use 45° ± 15° for the dips of these faults; the only important exception is for the Himalaya, where dips are only ~10° for most major earthquakes. Rakes are assumed to be uncertain by ±30°, which is again conservative, given the nearly pure dip-slip or strike-slip nature of the major faults.

We also assigned an uncertainty to the slip rate. To illustrate the slip data, we discuss examples of the fastest rates, those in excess of 10 mm a⁻¹. We used widely quoted rates for most of these faults, but with some alternatives, discussed below.

For underthrusting in the Himalaya, we assumed a rate increasing from west to east: 10 ± 2 mm a⁻¹ based on surface constraints on the amount of slip since 2 Ma in the Pakistan Himalaya (Baker et al. 1988) and 18 ± 7 mm a⁻¹ in western India, where onlapping sediment in the Ganga Basin implies such an average rate since 10–20 Ma (Lyon-Caen & Molnar 1985). In western Nepal, Lavé & Avouac (private communication, 1995) report a rate of 21 ± 1.5 mm a⁻¹ from growing folds during the past 10 kyr, consistent with GPS measurements spanning four years (Bilham et al. 1997). We use a larger uncertainty of 7 mm a⁻¹, first because the estimate ignores the possibility of additional shortening within the
Himalaya, and second because we have not seen the original data. Extrapolating such a rate to the eastern end of the Himalaya suggests that underthrusting in that area approaches 25 mm a\(^{-1}\).

For the major strike-slip faults of Tibet, we adopted rates of 29.8 mm a\(^{-1}\) for the Altyn Tagh fault (Avouac & Tapponnier 1993; Peltzer et al. 1988), 30–35 mm a\(^{-1}\) for the Karakorum fault (Avouac & Tapponnier 1993; Liu 1993), and 15 mm a\(^{-1}\) for the right-lateral strike-slip faults in central Tibet (Armijo et al. 1989).

For the Tien Shan, like Avouac et al. (1993), we assumed that rates decrease eastwards from a maximum in the west; we use the rate of 20 ± 6 mm a\(^{-1}\) across the western end of the Tien Shan, obtained from repeated GPS measurements and extrapolating the GPS rate across the entire belt (Abdrakhamatov et al. 1996).

We compare the calculated velocity field, based on Quaternary slip rates, both with rates of plate motions averaged over approximately the past 2 Myr (DeMets et al. 1990, 1994) and with velocities deduced using Very Long Baseline Interferometry (VLBI) made during the past 10 years. In order to make this comparison, we assume that variations in rates over such periods are small, and that 2 Myr averages and 10 yr averages agree. This assumption, which is unavoidable in the absence of independent measurements of rates made over the same timescale as our own, is justified to some degree by agreement of rates based on plate motions with those from VLBI (e.g. Gordon & Stein 1992) or of rates based on geodetic measurements (e.g. Feigl et al. 1993) and late Quaternary slip along the San Andreas fault (Sieh & Jahns 1984). As the only geodetic measurements that we use (Abdrakhamatov et al. 1996) span distances that are large compared with those affected by the elastic strain field near major faults, the 2 yr time span of these data ought not to be biased by elastic strain accumulation over the earthquake cycle.

Systematic errors in rates of slip appear to offer a more serious limitation to the approach taken here of using rates averaged over different intervals. Many rates of late Quaternary slip on faults are based on ratios of relatively accurately measured offset features divided by poorly constrained estimates of ages of those features. The procedure used to assign ages has been, in many cases, to assume that the features offset from glacial to interglacial climates at 10–20 ka may bias estimated slip rates towards high values. Representative slip rates could be two to four times lower than are commonly assumed in such settings, particularly where glacial climates were especially cold and dry. Moreover, such a bias might not be uniform, but might differ from setting to setting, as the timing of maximum glacial advances also seems to vary from region to region (Gillespie & Molnar 1995).

To allow for a possible systematic bias in rates based on assumed ages, we also present solutions for the velocity field with a modified data set. We used slip rates on the major faults mentioned above, the Altyn Tagh, the Karakorum, and the right-lateral faults in central Tibet, three times smaller than those assigned to them by Avouac & Tapponnier (1993), Liu (1993), and Armijo et al. (1989). For that inversion, we used the same uncertainties as in the inversion with full rates. With this range of uncertainties, the three-times lower rates include rates that could be either two times or four times lower. Moreover, the lower rate assigned to the Altyn Tagh fault includes the possibility that slip on the Altyn Tagh fault occurs at only 5 mm a\(^{-1}\) (Altun Tagh Active Fault Research Group 1992). A comparison of the two calculated velocity fields provides some support for the contention that a systematic bias does exist (Section 5).

3 METHOD

3.1 Strain rates from slip rates

A result of Kostrov (1974) is commonly used to form estimates of the average strain tensor that is due to slip on faults within a volume. Kostrov investigated the case of quasi-continuous deformation, wherein a region is deformed by slip on many faults whose dimensions are small compared with those of the region. He showed that, under these conditions, the average strain, \(\varepsilon\), of a volume, \(V\), is given by

\[
\varepsilon_{ij} = \frac{1}{2\mu V} \sum_{k=1}^{K} M_{ij}^k,
\]

where \(\mu\) is the elastic shear modulus of the region, and \(M_{ij}^k\) is the \(ij\)th component of the moment tensor of the \(k\)th earthquake occurring in the volume \(V\).

As derived, Kostrov's (1974) result might seem inapplicable to the calculation of average strain of domains within Asia, because major faults are few, and, in some cases, approach in length the size of deforming Asia itself. However, Molnar (1983) showed that, even when a region is cut by a single fault, the average strain of the region is identical to Kostrov's.

We subdivide the surface of Asia into triangles (Section 3.2), and the faults in Appendix A into segments, with each segment corresponding to the portion of the fault that traverses a particular triangle. We use the result of Kostrov (1974) to calculate the average rate of strain within each triangle due to slip on the fault segments contained within it.

Each fault segment is represented by a slip-rate vector \(\mathbf{u}\), and a normal, \(\mathbf{n}\), to its surface. A moment-rate tensor for this segment, \(\mathbf{M}\), which resembles the moment tensor of an earthquake, may be formed from the shear modulus, \(\mu\), the area, \(A\), of the fault segment and the outer product of these vectors:

\[
M_{ij} = \mu A(u_i n_j + n_i u_j),
\]

where

\[
A = \frac{LW}{\sin \delta}
\]

and \(L\) is the length of the fault segment, \(W\) is the thickness of the faulted layer, and \(\delta\) is the dip of the fault.
For each triangular element the strain-rate tensor, $\dot{e}$, is given by

$$\dot{e}_{ij} = \frac{1}{2} \sum_{k=1}^{3} M_{ij}^k,$$

(4)

where $V$ is the volume of the region of interest and equals the product of its surface area $A$ and thickness $W$. Combining (2), (3), and (4) eliminates $\mu$ and $W$, and the strain rate of a region may be expressed solely in terms of quantities that can be observed at the surface:

$$\dot{e}_{ij} = \frac{1}{2A} \sum_{k=1}^{3} L_k \left( \frac{u_{i}^{n_k} + u_{j}^{n_k}}{\sin \delta^k} \right).$$

(5)

If the z-direction is taken as vertical, then only the components $\dot{e}_{xx}$, $\dot{e}_{yy}$, and $\dot{e}_{xy}$ of the strain-rate tensor depend upon horizontal components of the velocity. Therefore, although we compute vertical components of shear on vertical planes (and horizontal components on horizontal planes), for the present application we use only $\dot{e}_{xx}$, $\dot{e}_{yy}$, and $\dot{e}_{xy}$ to calculate the horizontal components of velocity.

We carry out our calculations on a Cartesian grid, converting differences in longitude and latitude to the appropriate differences in $x$ and $y$, but not otherwise accounting for spherical geometry. We expect the distortion involved in this approximation to be small, over the $2500 \text{ km} \times 2500 \text{ km}$ area of maximum strain rates, especially in comparison with the uncertainties in the strain rates.

3.2 Inversion of strain rates to yield velocities

Haines (1982) showed that, if the horizontal components of the strain-rate field are known everywhere in a region, the field of relative horizontal velocity may be recovered by integration away from a boundary, assumed fixed. We take a similar approach by dividing the region of interest into triangles and assuming that, within each triangle, the strain rate is constant, so that the velocity field varies linearly across the triangle. We can express the velocity in the interior of the triangle in terms of the velocities of its vertices:

$$U = \sum_{m=1}^{3} N^m u^m,$$

(6)

where the superscript $m$ is an index, not a power. $u^m$ is the velocity of vertex $m$, and $N^m$ are interpolation functions:

$$N^1 = a' + b'x + c'y,$$

(7)

where

$$a' = 1/3,$$

(8)

$$b' = (y' - y^k)/2\Delta,$$

(9)

$$c' = (x' - x^k)/2\Delta,$$

(9)

where $x$, $y$, $z$ label the vertices of the triangle, $\Delta$ is its area, and $x$ and $y$ are (local) Cartesian coordinates with the centroid of the triangle as their origin. Shape functions associated with nodes $j$ and $k$ are obtained by cyclic permutation of the subscripts in the order $i, j, k$ (These functions are equivalent to standard, linear 'shape functions' of finite-element theory: see, for example, Zienkiewicz & Taylor 1989, p. 47.) Strain rates within triangles can be described with derivatives of (6):

$$\frac{\partial U_i}{\partial x_j} = \sum_{m=1}^{3} \frac{\partial N^m}{\partial x_j} u^m.$$

(10)

Equating observed components, $\dot{e}_{ij}$, of strain rate in triangular regions to those calculated from (10) yields the observation equations:

$$\dot{e}_{xx} = \sum_{m=1}^{3} b^m u_{ix}^m,$$

(11)

$$\dot{e}_{yy} = \sum_{m=1}^{3} c^m u_{iy}^m,$$

(12)

$$\dot{e}_{xy} = \sum_{m=1}^{3} \frac{1}{2} \left( c^m u_{ix}^m + b^m u_{iy}^m \right),$$

(13)

or

$$Au = \dot{e},$$

(14)

where the matrix $A$ contains derivatives of the shape functions $N$ for the triangles in which the observations of strain-rate components, $\dot{e}$, have been made, and $u$ are the velocities at the vertices of these triangles. The model parameters, $u$, are estimated from the observations, $\dot{e}$, by

$$u = (A^TCA)^{-1} A^T \dot{e},$$

(15)

where $C$ is the inverse of the variance–covariance matrix of the data. The variance–covariance matrix of the estimated velocities is

$$V = (A^TCA)^{-1}.$$

(16)

The velocity on two or more of the nodes must be fixed to provide a reference frame for the solutions.

To estimate the variance and covariance of the observations we assume that each fault segment is unrelated to each other segment. We form 1000 randomly distributed realizations of the equivalent-moment tensor of each fault segment, allowing the strike, dip, rake, and slip rate of each segment to vary between the limits discussed in Section 2, and allowing the slip rate on each segment to vary between the limits shown in Appendix A. From the distribution of these 1000 values, we form the variance–covariance matrix. Where more than one fault segment lies within a triangle, we add their variance–covariance matrices.

4 Inversion of Synthetic Data

Because we use slip rates on faults as data, convert them to strain rates within elements, and then invert strain rates to obtain a velocity field, we must carry out tests that demonstrate that the method works, when data are reliable and consistent, and experiments that evaluate the effects of inaccurate data on the resulting velocity fields.

4.1 Test of method using exact data

To demonstrate that the method works, we show results for two cases: (1) rigid-plate motion with elements spanning the plate boundaries and (2) the velocity field calculated for a thin viscous sheet (England & Houseman 1989). In each case, the input data are internally consistent.

4.1.1 Rigid plates

It may seem, at first sight, that the conversion of slip rates on faults, which are inherently narrow features in the crust, to average strain rates over regions hundreds of kilometres in
dimension could lead to serious errors in the velocity field. The equivalence, discussed above, between the results of Kostrov (1974) and Molnar (1983) shows that this procedure is not erroneous. Because, however, this conversion is central to the continuum approach that we, and others (e.g. Holt et al. 1991, 1995; Haines & Holt 1993), use, an illustrative calculation seems desirable. We constructed the velocity field for two plates separated by a transform fault and a subduction zone, draped a regular network of elements over it, and calculated (by eq. 13) average strain rates within each element (Fig. 2). Obviously, strain averaged this way occurs only in elements spanning the plate boundaries, and zero strain is calculated in all other elements. The velocity field inverted for this network agrees identically with that of the plates (Fig. 2). Although the agreement should come as no surprise, it does demonstrate that for our purposes of determining a velocity field on a scale much greater than the width of a fault zone, treating slip on faults as strain rates within elements of finite area introduces no error to the calculated velocity field.

4.1.2 Continuously varying strain-rate field

To evaluate the method applied to a strain-rate field typical of that in Asia, where orientations and magnitudes of strain rates vary continuously across the region, we use strain rates calculated by England & Houseman (1989) for a thin viscous sheet with an indenting boundary condition. To estimate strain rates, we first interpolated England & Houseman's numerically calculated velocity field onto a regular mesh of nodes and triangles (Fig. 3). We then decomposed the strain rate for each triangular region into equivalent double-couple moment tensors (Houseman & England 1986, Appendix A), and we assigned each double couple the same uncertainties in strike, dip, and rake of 5°, 15°, and 30°, respectively, as assigned to real data. We assigned the corresponding rate of slip on each equivalent double couple a uniform uncertainty of 30 per cent, typical of those assigned to real data (Section 2). With velocity fixed at zero on the top boundary of the figure, the velocity field is recovered from the strain rates essentially without error, and with an uncertainty that is arbitrarily small, because it depends only upon the magnitude of uncertainty that is assigned to the input strain rates (Fig. 3).

4.2 Contamination of velocity fields by inconsistent data

The quality of agreement in the preceding experiments is not surprising because, as Haines (1982) showed, the velocity field should be recovered if the strain-rate field is known everywhere. In the examples we have just considered, an internally consistent synthetic strain-rate field covers completely the region of interest. In the real case of deformation in Asia, however, considerable uncertainty attaches to the rates of motion on all the major faults, and there is the distinct possibility that the data will not form a self-consistent set, in the sense defined in the Introduction. We may reasonably expect two types of error in using Quaternary faulting data to infer strain rates. By underestimating the slip rate on a fault or ignoring important faults in a region, we may underestimate one or more components of strain rate in a region, or we may overestimate the slip rates and corresponding strain rates. We illustrate the sensitivity of our inversion scheme to such errors by adding large horizontal simple shear strains to the internally consistent strain-rate field of Fig. 3. Experiments involving the omission of data showed qualitatively similar results, but are not illustrated.

We chose a magnitude of $10^{-7} \text{a}^{-1}$ for the additional strain rate, and applied it to eight elements within the shaded area in Fig. 4. If the dimensions of the region are scaled to those of Asia, with the sides of elements $\sim 300$ km in length, the additional strain rate used here corresponds to the addition of an east–west strike-slip fault slipping at $\sim 30 \text{mm a}^{-1}$, without any change to the surrounding strain-rate field. Because these strain rates are added only to eight adjacent elements, they must be inconsistent with the strain rates in neighbouring elements. Fitting both the internally consistent and the additional, inconsistent, data requires some kind of compromise, which manifests itself as misfits of some strain rates.

Central to these experiments are the uncertainties assigned to the various strain rates, because the inversion method weights each component of strain rate inversely proportionally to the uncertainty assigned to it. We treat the synthetic data as real fault data, characterized by strikes, dips, and rakes, respectively uncertain by 5°, 15°, and 30°, and by slip rates uncertain by amounts specified below for separate experiments.

With all strain rates, including the additional, inconsistent set, assigned the same 30 per cent uncertainty, the inversion yields velocities that differ only by a few per cent from those calculated from the internally consistent strain-rate field (Fig. 4a). The inconsistent strain rates are misfit and do not contribute significantly to the velocity solution. Fitting the added sinistral shear would require either sinistral motion beyond the edges of the shaded region, or contraction and extension at its edges. The consistent set of strain rates outside the shaded region, however, do not contain such deformation and because they outweigh the inconsistent strain rates, the latter are not fit.

The inconsistent strain rates can be fit better, but still not well, if the uncertainties in fault-slip rates in all elements are raised to 100 per cent and the uncertainties in strike, dip, and rake are all raised to 40° (not illustrated). The explanation for the continuing poor fit to the inconsistent data is that the increases in uncertainties do not represent changes in the relative weights of the inconsistent and consistent input strain rates.

Clearly, if the inconsistent slip rates were assigned a percentage uncertainty lower than the internally consistent rates, then a better fit to the inconsistent rates could be obtained. With an uncertainty assigned to the inconsistent slip rates of 3 per cent, rather than the 30 per cent assigned to the consistent data (Fig. 4b) the high, inconsistent strain rates can be fit well. With the uncertainty in the inconsistent slip rates set to 10 per cent, however, the calculated velocity field resembles the results shown in Fig. 4(a). These, and other, tests indicate that even hugely inconsistent rates, corresponding to 30 mm a$^{-1}$ of slip on a fault traversing several elements, perturb the calculated velocity field by only small amounts ($1–2 \text{mm a}^{-1}$) unless much smaller percentage uncertainties are assigned to the inconsistent rates than to rates in adjacent elements. Because the errors in real fault-slip rates are likely to be much smaller than the 30 mm a$^{-1}$ gaffe assumed here, and because we assigned no rate an uncertainty as small as 10 per cent of its value, we conclude that individual inconsistent strain rates should not perturb the calculated velocity field by more than a few millimetres per year.
Figure 2. Open arrows show velocities for a synthetic calculation in which two rigid plates are separated by a pair of faults representing a convergent and a transform margin. Slip on these faults is converted into equivalent strain rates over the triangles containing the faults. Bars show principal horizontal strain rates; thick bars correspond to contractional strains and thinner bars to extensional strains. Input strain rates are shown as open bars; strain rates calculated from the inversion are indistinguishable (less than 1 per cent different) from the input, and are shown as black symbols. The open and black symbols are drawn to the same length scale, but the black symbols are narrower, for clarity. Units of velocity and strain rate are arbitrary.

Figure 3. Open arrows show velocities from a calculation of the deformation of a thin viscous sheet (England & Houseman 1989, Fig. 7). Only the right half of the solution is shown; there is mirror symmetry about the left vertical axis. Solid arrows show the velocities recovered from the strain rates given by that calculation, using the method of this paper. The ellipses show formal uncertainties in the recovered velocities. Bars show principal horizontal strain rates; convention as in Fig. 2. Strain rates calculated from the inversion are indistinguishable (less than 1 per cent different) from the input. As the original calculations were non-dimensional, scales of velocity and strain rate are arbitrary.
Figure 4. (a) Black arrows, almost obscured, show velocities calculated from the same input data as shown in Fig. 3, except that an inconsistent simple shear at a rate of $10^{-7} \text{ a}^{-1}$ has been added to the strain rates in the shaded region. Open arrows show the solution without inconsistent data, plotted at the same scale as the solid arrows. Representations of the input and output principal strain rates are shown with symbols as in Fig. 3. Open symbols correspond to input strain-rate data; black symbols show strain rates calculated from the inversion; where only a black symbol can be seen, the input and output are indistinguishable on the scale of the figure. In the inversion, data covariance is estimated by assuming rates of slip to be uncertain by 30 per cent, strikes by $5^\circ$, dips by $15^\circ$, and rakes by 30°. (b) As (a), except that the slip rates within the shaded region are uncertain by only 3 per cent. (c) As (a), except that in any region where the strain rate falls below $10^{-7} \text{ a}^{-1}$, the uncertainty in strain rate is set to this value. (d) As (c), except that the threshold strain rate is $2 \times 10^{-7} \text{ a}^{-1}$. © 1997 RAS, GJI 130, 551–582
This test is incomplete in one respect that prevents it alone from being definitive. In the inversion, low strain rates in one region prohibit large components of some strain rates in neighbouring regions that share the same nodes. In the synthetic data, low strain rates are correctly assigned and consistent with strain rates in adjacent elements, but in parts of Asia a low strain rate may be assigned because evidence of strain is missing or has been overlooked. We must assign an uncertainty to components of strain rate in such regions to allow the possibility of unrecognized strain. The relative magnitudes of the uncertainty assigned to possibly undetected strain and the uncertainty assigned to any overestimated strain will affect the degree to which the latter can be fit in the inversion.

Accordingly, we compare calculated velocity fields for which different uncertainties are assigned to slowly straining elements. To any element where a component of the input strain rate falls below a chosen value, we set the uncertainty to that value, and we ignore any covariance between components assigned that value. With a threshold strain rate of $10^{-8}$ a$^{-1}$ (Fig. 4c), the inconsistent data are still fit poorly, but with the threshold uncertainty set to $2 \times 10^{-8}$ a$^{-1}$ (Fig. 4d), the additional, inconsistent, strain rates are fit reasonably well and contaminate the velocity field.

Although the magnitude of the augmented uncertainty in Fig. 4d ($2 \times 10^{-8}$ a$^{-1}$) is much smaller than the magnitude of the additional, inconsistent, strain rates ($10^{-7}$ a$^{-1}$), the augmented uncertainty applies to a much larger area than does the inconsistent strain rate. Consequently, the large region of augmented uncertainty can absorb the extra shear at low strain rates spread over a large area. The product of additional shear and the area over which it applies is $4 \times 10^{-7}$ a$^{-1}$, where we let areas of elements be 0.5 dimensionless units. For an augmented uncertainty of only $10^{-9}$ a$^{-1}$ (Fig. 4c) the product of augmented uncertainty with the area over which it applies is $5.5 \times 10^{-9}$ a$^{-1}$, comparable with the corresponding product for the inconsistent shear. For an augmented uncertainty of $2 \times 10^{-8}$ a$^{-1}$ (Fig. 4d), however, the product of $20 \times 10^{-7}$ a$^{-1}$ exceeds that for the inconsistent strain rates by more than five times.

This comparison, and other tests not reported, suggest a simple rule. First, note that the product of an inconsistent strain rate with its areal extent is equal to the product of the inconsistent slip rate on the fault times its fault length (see eqs 2–4). For the experiments illustrated by Fig. 4, a region with area $A$ assigned a threshold strain rate of $\delta$ will prevent contamination by an erroneous slip rate $\delta V$ on a nearby fault of length $L$ if $\delta V L \leq A \delta$. For instance, if an excessive slip rate of 10 mm a$^{-1}$ were assigned to a fault 1000 km in length, the assignment of a threshold strain rate of $10^{-8}$ a$^{-1}$ to surrounding elements with an area of 1 000 000 km$^2$ would prevent contamination by that erroneous slip rate. We experimented with different threshold strain rates and commonly used a value of $6 \times 10^{-9}$ a$^{-1}$, smaller than the $10^{-8}$ a$^{-1}$ used in these tests. Thus, inconsistent slip or strain rates will be suppressed more than is implied by the experiments described above.

We obtained results qualitatively similar to those just discussed both in the cases for which inconsistent contractional or extensional strain were added to otherwise consistent strain-rate fields, and in experiments where an underestimation of strain rates was simulated by artificially setting components of strain rates to zero.

These experiments illustrate a general feature of the problem we address: the influence exerted by a set of strain-rate measurements in one region on the calculated velocities in an adjacent region is in proportion to the relative weights of the data for the two regions, where weight can be loosely defined as the product of area and strain rate, divided by the uncertainty in the strain rate. The experiments were motivated by a particular concern, namely that the slip rates on one or more of the major faults in the region could be seriously in error (Section 2). We conclude that inaccurate strain rates associated with such an error could contribute erroneous velocities only if those strain rates were assigned an uncertainty that is proportionately lower than uncertainties assigned to the remaining strain rates, or if the threshold strain rate for regions of negligible strain is much larger than we assume here.

5 RESULTS

5.1 Implications of averaged strain rate

The calculations of the velocity field in Asia are based on strain-rate estimates made from the fault-slip rates illustrated in Fig. 1. A consistent aspect of the calculations, shown in this section, is that relatively little of the convergence between India and Asia is accommodated by the expulsion of material eastwards, in contrast with the conclusions of previous studies (Avouac & Tapponnier 1993; Molnar & Deng 1984). We begin by showing that the result is a property of the data, and not of the method of analysis, by calculating the regional strain for the whole of deforming Asia using the method of Kostrov (1974).

We convert fault-slip rates and dimensions to equivalent moment-rate tensors (eqs 2 to 4), assuming values of $3.3 \times 10^{10}$ N m$^{-2}$ for $\mu$, the shear modulus, and 10 km for $W$, the thickness of the seismogenic layer. The principal horizontal axes of the summed moment-rate tensor, using the faster slip rates assigned to the major strike-slip faults (see Section 2, and Appendix A), are $-91 \times 10^{18}$ N m a$^{-1}$ at an azimuth of 23°, and $22 \times 10^{18}$ N m a$^{-1}$ at an azimuth of 113°. The azimuth of the contractional axis lies close to that of India’s velocity with respect to Eurasia at the Himalayan front, which varies from about 0° in the west to about 18° in the east.

We interpret this summed moment tensor by rotating it so that the $y$-direction lies parallel to the India–Asia convergence direction, taken to be 10°E. If all the strain in a region is taken up on faults, then the $y'y'$ element of the moment-rate tensor is directly related to the contractional or extensional component of plate motion (Jackson & McKenzie 1988). If the length of the plate boundary to the deforming zone is $S$, and the thickness of the seismogenic layer is $W$, then the $y'$-component of relative plate velocity is given by

$$v_y = \frac{M_{yy}}{2 \mu W}.$$  

(17)

The value of $v_y$ does not depend on the assumed values of $W$ and $\mu$, which appear in the denominator of eq. (17), because these quantities also appear implicitly in the numerator (see eq. 2).

The value of $M_{yy}$ calculated from the Quaternary slip rates is $-86 \times 10^{18}$ N m a$^{-1}$, nearly the same as the principal contractional moment rate, because the orientations of the principal contractional axis and the plate convergence direction are nearly the same (Table 1). Taking the along-strike length
of the Himalaya to be 2500 km, the $M_{x'y'}$ component yields a value of 52 mm a$^{-1}$ for the average convergence rate between India and Eurasia. The average speed of India with respect to Asia at the Himalayan front is approximately 50 mm a$^{-1}$, calculated from the angular velocity of DeMets et al. (1994).

The value of $v_r$ is, however, sensitive to uncertainties in the slip rates of the faults. One of the largest uncertainties is connected with the major strike-slip faults in central and northern Tibet, which have both great length and high inferred slip rates (Section 2 and Table 1). If we use the data set containing the slower rates of slip on these faults, the value of $M_{x'y'}$ becomes $-73 \times 10^{18}$ N m a$^{-1}$, and the equivalent convergence rate becomes 44 mm a$^{-1}$. The difference between these quantities and those obtained using the faster slip rates on the strike-slip faults is small, because the orientations of these faults are at high angles to the convergence direction.

The agreement between the convergence rates calculated from plate relative motions and from the faulting implies that the data set we are using is representative of the total active deformation in Asia. If this implication is accepted, the summed moment tensors may be used to estimate the overall partitioning of convergence into shortening in the $y'$-direction (represented by $-M_{y'y'}$) and extension (or expulsion) in the $x'$-direction (represented by $M_{x'y'}$). Then the ratio $M_{x'y'}/M_{y'y'}$ represents the fraction of the shortening in the direction of convergence ($y'$-direction) that is accommodated by expulsion of material in the perpendicular ($x'$-direction). This fraction is 19 per cent for the summed moment tensors using the fast slip rates on the major strike-slip faults (Table 1), but only 5 per cent if the lower values of these slip rates are used. This result shows the fact that none of the velocity fields calculated below shows the large eastward velocities that have been suggested in the past for South China (Avouac & Tapponnier 1993; Molnar & Deng 1984).

5.2 Inversion for velocity field

To calculate the velocity field in Asia we divide the region into triangular elements whose side length, initially, is about 500 km (Fig. 5b). The mesh geometry is chosen to give resolution in regions where the fault style changes rapidly from place to place. For example, element boundaries lie north of the Himalaya, between the region of rapid north–south shortening of the Himalaya and the much slower east–west extension of the Tibetan plateau. Similarly, the aseismic Tarim Basin is separated from its deforming surroundings. We also examined the use of both coarser and finer meshes, and discuss these results below.

We fix the velocity to be zero on the three nodes at 60°N and on the node at 70°E, 50°N, which gives velocities in a reference frame fixed to Eurasia. We allow for the possibility that undetected strain may occur in regions where no faulting is described by setting the uncertainty in strain rate in these regions to $6 \times 10^{-9}$ a$^{-1}$. This uncertainty is almost certainly an overestimate; for example, if such a rate of strain were to exist over the 1500 km between the Tien Shan and 60°N, the velocity difference would be 10 mm a$^{-1}$, equivalent to 20 per cent of the rate of convergence between India and Asia. In tests of the effect of this quantity, a value of $3 \times 10^{-9}$ a$^{-1}$ yielded velocities that differed from those shown in Fig. 5 by less than the uncertainties in the latter and a value of $9 \times 10^{-9}$ a$^{-1}$ gave velocities that agree with those shown in Fig. 5 to within the combined uncertainties of the two solutions.

5.2.1 Results with high slip rates on the major strike-slip faults

We use the fault-slip rates, and uncertainties, listed in the Tables in Appendix A, employing always the higher value of slip rate if there is a choice.

97 per cent of input strain rates are fit in the solution to less than two standard errors, and 90 per cent are fit to better than one standard error. The observed distribution is, however, both more narrowly peaked, and has more outliers, than would be expected for a normal distribution (Fig. 6). Thus we cannot assume Gaussian statistics and treat the model covariance matrix (eq. 16) as an accurate measure of uncertainties. To evaluate the uncertainties we carried out Monte Carlo calculations, with all data allowed to vary within the bounds given above. Using 1000 such input strain-rate fields we calculated 1000 velocity fields, determined covariances and variances about the mean, and compared these with the estimates given in Table 1.

Because the only velocities fixed in this calculation are on the Siberian and Kazakh platforms, a check on our calculations is to compare the velocity they predict for points on the Indian shield with the velocity of India with respect to Eurasia calculated from plate tectonics. The grey arrows on Fig. 5(a) show the velocities, calculated from the India–Eurasia pole of...
Figure 5. (a) Solid arrows show velocities calculated from the strain rates shown in (b). Grey arrows show the velocity of India with respect to Asia calculated from the angular velocity of DeMets et al. (1994). The grey arrow, barely visible, at 121.5°E, 31.2°N shows the motion of Shanghai obtained from VLBI observations (Molnar & Gipson 1996). (b) Mesh of triangles used for calculating the velocity field in Asia. Strain rates calculated from the fault-slip data illustrated in Fig. 1 are shown as open symbols (convention as in Fig. 3); strain rates calculated in the inversion are shown with a black shade. Open circles at the top of the figure show the four nodes whose velocities are constrained to be zero.

© 1997 RAS, GJI 130, 551–582
The major disagreements between input and output strain rates in Fig. 5(b) lie in the northwest Tibetan Plateau, where slip rates are inconsistent with the strain rates of surrounding regions, whereas the lower slip rates are consistent.

5.2.3 Influence of constraining the motion of India

Although the calculated velocities for points on India agree, to within their uncertainties, with those of NUVEL-1A for either set of slip rates, there is a systematic difference, in that the calculated velocity for eastern India points west of the plate velocity both in Fig. 5(a) and in Fig. 7(a). This difference might suggest systematic errors in the calculated velocity field.

Indeed, calculations in which the three southernmost points in India are fixed to the velocities corresponding to NUVEL-1A yield eastward components of the movement of Tibet and South China that are 4–6 mm a\(^{-1}\) larger than their equivalents without constraints on the motion of India (compare Figs 5a, 7a, and 8). Even with this additional constraint, however, the eastward motion of South China remains lower than 10 mm a\(^{-1}\), and comparable with the rate inferred from VLBI. The misfits for the two cases are \(X^2 = 156\) for the fast slip rates and \(X^2 = 86\) for the slower rate, very close to the values obtained from the calculations with India unconstrained (above). This correspondence demonstrates that the extra constraint exposes no additional inconsistency in the input strain rates; in particular, the imposition of this constraint does not change the conclusion of the previous subsection, that the higher slip rates on faults in northern and eastern Tibet are inconsistent with the strain of their surroundings.

5.2.4 Influence of different grid sizes

Our conclusions about the velocity field are insensitive to the size of the grid we choose. The grid illustrated in Fig. 5 has 76 triangles and 47 nodes (four of which are fixed), yielding 228 data with 86 unknowns, and thus 142 degrees of freedom. A grid of triangles with roughly twice the side length of those used for Fig. 5 yields velocities that agree, within their uncertainties, with those calculated on a grid twice as fine as that used for Fig. 5 (Fig. 9a). These two velocity calculations are each consistent with that shown in Fig. 5. For the fine grid, 89 per cent of input strain rates are fit to within one standard error and 97 per cent to within two standard errors (Fig. 9b, 154 triangles, 87 nodes, four fixed nodes, 296 degrees of freedom). For the coarse grid, 87 per cent of input strain rates are fit to within one standard error and 98 per cent to within two standard errors (Fig. 9c, 44 triangles, 30 nodes, four fixed nodes, 80 degrees of freedom).

5.2.5 Motion of South China

When large strike-slip faults were recognized in western China and adjacent areas, they were first interpreted as evidence that material is translated eastwards along them and out of India’s northward path (e.g. Molnar & Tapponnier 1975; Tapponnier & Molnar 1976). Later when slabs of plasticine indented by a rigid die and constrained to deform in plane strain revealed a similar pattern of strike-slip shear, Peltzer & Tapponnier (1988; Tapponnier et al., 1982, 1986) interpreted not only the present kinematic pattern, but also finite strain in Asia as evidence for such eastward ‘propagating extrusion’ of China. Attempts to quantify the current rate of eastward transport, however, have been few. Molnar & Deng (1984) used seismic moments of
Figure 7. Velocity and strain-rate solutions for calculations in which the slip rates on the major strike-slip faults in northern and central Tibet, and the Karakoram fault are reduced to one-third of the values given in Table A1. These faults are marked by the symbol † in the Table. Velocities calculated with the same mesh as in Fig. 5 are shown in (a); strain rates are shown in (b).
large earthquakes to suggest that South China moves east-southeast at \( \sim 20 \text{ mm a}^{-1} \) with respect to Asia. Later, Avouac & Tapponnier (1993) treated part of Tibet as a rigid block and with a number of additional assumptions suggested that South China probably moves eastwards at a minimum rate of \( \sim 0.15 \text{ mm yr}^{-1} \) relative to Siberia. More recently, Peltzer & Saucier (1996) expanded Avouac and Tapponnier's approach by dividing Asia into numerous blocks separated by faults with assigned slip rates. With a finite-element approach, they minimized simultaneously elastic strain within blocks and misfits to slip rates, calculating a velocity field that resembles those in Figs 5 and 9, but with \( \sim 15 \text{ mm a}^{-1} \) of ESE movement of South China.

In contrast with these views, numerical experiments on a homogeneous, thin viscous sheet within a gravity field and indented by a rigid die repeatedly showed only a small amount of lateral transport; instead most of the shortening of the sheet was accommodated by thickening of it in front of the rigid die (England & Houseman 1986; Houseman & England 1986, 1993). Thus, measuring the speed with which South China moves east or southeast with respect to Eurasia provides a test of the extent to which the plane strain and rigid blocks, or the thin viscous sheet apply to the current deformation of Asia.

In our calculations, when India's velocity with respect to Eurasia is not constrained, South China is calculated to move south- to south-southwestwards at less than 10 mm a\(^{-1}\). When India's velocity is constrained to agree with NUVEL1-A, however, South China is calculated to move eastwards at 5 to 9 mm a\(^{-1}\), depending upon which data set is used. As uncertainties in such velocities are \( \sim 10 \text{ mm a}^{-1} \), they do not permit estimated speeds of South China in excess of 15 mm a\(^{-1}\), but only with small probability. Thus our calculations favour only limited 'extrusion' and the thin viscous sheet passes a test.

Holt et al. (1995) derived strain estimates from the moment tensors of this century's earthquakes in Asia and found two acceptable solutions for the motion of South China with respect to Eurasia. In one they obtained a velocity for South China that was comparable with the velocity we obtain (10 mm a\(^{-1}\) or less in a roughly easterly direction), but were unable to fit the style of deformation in eastern Eurasia. In the other they could fit the style of deformation but obtained eastward velocities of 20 mm a\(^{-1}\) or more for South China.

The principal difference between our solutions and those of

---

**Figure 8.** As Figs 5(a) and 7(a), except the three southernmost nodes in India are constrained to have velocities relative to Eurasia given by NUVEL1-A (DeMets et al. 1994). White arrows are calculated using the full slip rates for all faults in Table A1, black arrows are calculated with slip rates on the major strike-slip faults in northern and central Tibet, and the Karakoram fault reduced to one-third of the values given in Table A1.

© 1997 RAS, GJI 130, 551–582
Figure 9. (a) Velocities calculated for the two meshes shown in (b) and (c). White arrows are for the coarser mesh, black arrows for the finer mesh. The arrows are scaled so that one degree of latitude is equivalent to 10 mm a$^{-1}$. (b) Input and output strain rates for calculation with the fine mesh. Symbols as Fig. 5(b). (c) As (b), but for the coarser mesh.
Holt et al. (1995) results from the use of different data. The moment-rate tensor calculated from the earthquakes shows, when rotated into the orientation of India-Asian convergence (Table 1), an x'y'-component (corresponding to dextral shear parallel to convergence or sinistral shear perpendicular to convergence) that is equal in magnitude to the y'y'-component (corresponding to shortening in the direction of convergence). In contrast, the x'y'-component in the tensor calculated from the Quaternary faulting is less than one third of the magnitude of the y'y'-component (Table 1). It is inevitable, because of this contrast, that there should be significant differences between the velocity fields calculated from the two sets of data and that x'y'-shear should be more prominent in the solution based on this century's earthquakes.

Two factors account for the difference between our calculated rate and the rapid rates of eastward transport of South China inferred by Armijo et al. (1989), Avouac & Tapponnier (1993), Peltzer & Tapponnier (1988) and Tapponnier et al. (1986). First, the rates of strike-slip faulting inferred by these authors may well be in error, as we discuss above. Secondly, their route from inferred slip rate on individual faults to large-scale tectonic transport was through plate-like models. Even where rigid bodies can be recognized in regions of diffuse continental deformation, plate-like calculations of relative motions from slip vectors are not valid where rigid bodies are embedded in broad deforming zones (e.g. England & Jackson 1989). To estimate rates of relative motion in such cases, one must take account of rotations about vertical axes, which are not observable from faulting alone. If these rotations are ignored, velocity fields inferred from faulting may be seriously in error, as we now discuss (see also McKenzie & Jackson 1983; England & Molnar 1990).

5.2.6 Rotations about vertical axes

An important part of the case for rapid eastward transport of crustal material in Asia has been the zone of left-lateral strike-slip faulting in eastern Tibet. Originally these faults (see Fig. 1) were seen as traces of a broad left-lateral shear zone accommodating a southward increase of east-to-southeastward velocity with respect to Eurasia (Molnar & Tapponnier 1975). Later, it became clear that the sense of slip on these faults could result from their rotating clockwise within a zone of roughly north–south right-lateral shear (Cobbold & Davy 1988; England & Molnar 1990).

The velocity fields calculated here allow a test of these ideas. From the velocity field, which is based on an internally consistent strain field, we may calculate an internally consistent spin field, where spin is defined as

$$\Omega = \frac{1}{2} \left( \frac{\partial u_2}{\partial x_1} - \frac{\partial u_1}{\partial x_2} \right).$$  \hspace{1cm} (18)

The spin represents the average rate of clockwise rotation, about a vertical axis, of an infinitesimal volume. The rate of rotation of crustal blocks in a zone of distributed deformation depends, in general, on the shape of the blocks, and on their orientation with respect to the shear (Lamb, 1987). The rates should lie close to \( \Omega \) for blocks that are roughly equant in map view (McKenzie & Jackson, 1983), and, for long thin blocks, should lie between 0, for blocks aligned with the simple-
shear direction, and $2\Omega$, for blocks lying transverse to the shear (Lamb 1987).

An example of the spin field, shown in Fig. 10, based on the velocity field calculated on the fine grid and shown in Fig. 9(b), shows several expected features. Clockwise spin occurs in Burma and western Tibet, associated with shear along the Sagaing and Karakorum faults, respectively. Counterclockwise spin in northern Tibet is associated with shear along the Altyn Tagh fault system. Clockwise spin of the Tarim Basin at a nearly constant rate (about 1° Ma$^{-1}$) is associated with its rigid-body rotation with respect to Eurasia (e.g. Avouac & Tapponnier 1993).

The spin in eastern Tibet is clearly clockwise, which is at first sight inconsistent with the sinistral motion on the major faults of the region. Clockwise spin is to be expected, however, if the region is accommodating roughly north–south right-lateral shear resulting from the difference in northward motion between central Tibet and South China (Cobbold & Davy 1988; England & Molnar 1990). The magnitude of the spin in this region (1–2° Ma$^{-1}$) is consistent with that estimated by England & Molnar (1990). If the left-lateral faults of this region are rotating at 1° Ma$^{-1}$ or more then they are accommodating right-lateral north–south shear rather than left-lateral east–west shear (see England & Molnar 1990, Fig. 2).

5.2.7 Consistency of velocity field with orientation of faulting

Faults in the upper crust are shear fractures and represent, therefore, lines parallel to which there is no local change of length. Haines & Holt (1993) suggested that, if the brittle upper crust accommodates, by faulting, distributed deformation in the lower lithosphere, then lines of no length change in the continuous velocity field should be aligned with the orientations of faults in the brittle layer. Because, in this study, the strain rates used to calculate the velocity field are derived from faulting, we should expect coherence between these two directions. Our calculation, however, involves replacing discrete slip on faults by strain averaged over triangles of side 300 km or longer, which is essentially a process of smoothing. If this smoothing were to distort the nature of the strain, then we should expect that the lines of no length change would be rotated with respect to the orientations of the faults. It is worthwhile, therefore, to determine whether the calculated orientations of faulting are consistent with the input orientations.

The orientations, $\phi$, of lines of no length change are given by

$$\tan \phi = -\frac{\varepsilon_{xx} + \sqrt{\varepsilon_{xx}^2 - \varepsilon_{xx} \varepsilon_{yy}}}{\varepsilon_{yy}},$$

(19)

where $\phi$ is measured counterclockwise from the $x$-axis. This equation has no solution when $\varepsilon_{xx}^2 < \varepsilon_{xx} \varepsilon_{yy}$, which corresponds to the condition that both principal horizontal strain axes have the same sign (either net dilatation or net contraction). In such cases, no preferred orientation of faulting can be chosen without making additional assumptions.

The directions of no length change calculated from the
Crustal velocity field in Asia

569

Figure 11. Orientations of lines, in the velocity field of Fig. 9(a), that have no rate of change of length. Orientations of these lines are calculated for each triangular element in Fig. 9(b). Black bars correspond to orientations of lines across which there is sinistral shear, as well as shortening or extension. Grey symbols correspond to orientations of lines across which there is dextral shear, as well as shortening or extension. Where black and grey bars are parallel, faulting would be dip-slip; where the two bars are perpendicular, the faulting would be purely strike-slip. Otherwise, oblique faulting is implied. Lines indicate the locations of faults shown in Fig. 1.

velocity field shown in Fig. 9(a,b) agree with the orientations of, and sense of slip on, the faults in Asia (Fig. 11). In nearly all cases, one or other of the lines of no length change lies parallel to the nearby faults, and the sense of shear (given by the shading of the symbol) agrees with that observed.

Note that misalignment between the faults and the directions of no length change in the velocity field can result either from our having over-smoothed the strain field, or from the local orientation of a fault being controlled by pre-existing weakness of the crust. In view of the close agreement in general between the orientations of faults and of lines of no length change in the continuous velocity field, particularly noticeable for the larger faults, we conclude that the process of smoothing we employ here does not distort the strain field.

6 CONCLUSION

A stable large-scale feature of all the velocity fields calculated here is that the convergence between India and Asia is absorbed primarily by north-south shortening north of India, in the Himalaya, Tibet, the Tien Shan, and the Nan Shan. Fast rates of slip (20–30 mm a⁻¹) on major strike-slip faults in central and northern Tibet are incompatible with the strain in surrounding regions, but rates of slip 2–4 times lower on these faults are consistent with the strain in the surrounding region. The calculated rate of eastward transport of South China is, in all cases, lower than 10 mm a⁻¹, smaller than that proposed by Avouac & Tapponnier (1993), Molnar & Deng (1984), or Peltzer & Saucier (1996), and consistent with the measurements using VLBI. This rate, however, lies within the range of 0 to 10 mm a⁻¹ estimated by England & Molnar (1990), who, like Cobbold & Davy (1988), inferred that the sinistral strike-slip faulting in eastern Tibet reflects north-south dextral shear and clockwise rotation of the faults, rather than irrotational eastward transport of material.

One view of continental deformation is that the motions are fundamentally discontinuous (e.g. Avouac & Tapponnier 1993; Peltzer & Saucier 1996; Peltzer et al. 1988). An alternative view is that the discontinuous deformation at the surface reflects localization, within the brittle upper crust, of continuous strain in the lower, ductile, parts of the lithosphere (e.g. Molnar 1988; England & Jackson 1989). The calculations of this paper show that the observations of Quaternary faulting in Asia can be fit by a continuous velocity field sampled on the scale of ~300 km. This field is consistent with gross constraints on the deformation, in that it yields rates of motion between India and Eurasia that differ by at most a few millimetres a year from those calculated from plate motions since 2 Ma. The field is also consistent with the detailed constraints implied by the observed orientations of faulting (Fig. 11). These agreements suggest that the discontinuous surface deformation in Asia can be reconciled with a continuous distribution of deformation at depth. The similarity between the velocity field calculated here, and that of Peltzer & Saucier (1996), calculated by minimizing the strain in blocks separating a few major faults, demonstrates, however, that...
References

Abdrakhmatov, K.Ye. et al. 1996. Relatively recent construction of the Tien Shan inferred from GPS measurements of present-day crustal deformation rates, Nature, 384, 450–453.

Allen, C.R., Gillespie, A.R., Han, Y., Sieh, K.E., Zhang, B., & Zhou, H., 1997. Estimation of slip rates in the southern Tien Shan using cosmic ray exposure dates of abandoned alluvial fans, Geol. Soc. Am. Bull., submitted.

Burchfiel, B.C., Zhang, P., Wang, Y., Zhang, W., Song, F., Deng, Q., Molnar, P. & Royden, L., 1991. Geology of the Haiyuana fault zone, Ningxia-Hui Autonomous Region, China, and its relation to the evolution of the northeastern margin of the Tibetan plateau, Tectonics, 10, 1091–1110.

Burchfiel, B.C., Chen, Z., Liu, Y. & Royden, L.H., 1996. Tectonics of the Longmen Shan and adjacent regions, Central China, Int. Geol. Rev., 37, 661–735.

Burchfiel, B.C., Chen, Z., Liu, Y. & Royden, L.H., 1996. Tectonics of the Longmen Shan and adjacent regions, Central China, Int. Geol. Rev., 37, 661–735.

Butler, R.W.H. & Prior, D.J., 1988. Tectonic controls on the uplift of the Nanga Parbat massif, Pakistan Himalayas, Nature, 333, 247–250.

Chen, W.-P. & Kao, H., 1996. Seismotectonics of Asia: some recent progress, in Tectonic Evolution of Asia, ed. Yin, A. & Harrison, T.M., Cambridge University Press, Cambridge.

Chen, W.-P. & Molnar, P., 1997. Seismic moments of major earthquakes and the average rate of slip in central Asia, J. geophys. Res., 82, 2945–2969.

Chen, W.-P. & Molnar, P., 1990. Source parameters of earthquakes and intraplate deformation beneath the Shilong Plateau and northern Indoburman Ranges, J. geophys. Res., 95, 12527–12552.

Chen, W.-P. & Nábelek, J., 1995. Seismogenic strike-slip faulting and the development of the North China Basin, Tectonics, 7, 975–989.

Cobbold, P.R. & Davy, Ph., 1988. Indentation tectonics in nature and experiment. 2. Central Asia, Bull. Geol. Inst. Uppsala, NS, 14, 143.

Curran, J.R., Moore, D.G., Lawver, L.A., Emmel, F.J., Raitt, R.W., Heany, M. & Kieckhefer, R., 1978. Tectonics of the Andaman Sea and Burman, in Geological and Geophysical Investigations of Continental Margins, eds Watkins, J.S., Montardert, L. & Dickerson, P., Mem. Am. Assoc. Petrol. Geol., 29, 189–198.

Davy, P. & Cobbold, P., 1988. Indentation tectonics in nature and experiment. 1. Experiments scaled for gravity, Bull. Geol. Inst. Uppsala, 14, 129–141.

DeMets, C., Gordon, R.G., Argus, D.F. & Stein, S., 1990. Current plate motions, Geophys. J. Int., 101, 425–478.

DeMets, C., Gordon, R.G., Argus, D.F. & Stein, S., 1994. Effect of recent revisions to geomagnetic reversal timescale on estimates of current plate motions, Geophys. Res. Lett., 21, 2191–2194.

Deng, Q. et al., 1984. Active faulting and tectonics of the Ningxia-Hui Autonomous Region, China, J. geophys. Res., 89, 4427–4445.

England, P.C. & Houseman, G.A., 1986. Finite strain calculations of continental deformation 2. Comparison with the India–Asia collision zone, J. geophys. Res., 91, 3664–3676.

England, P.C. & Houseman, G.A., 1989. Extension during continental convergence with application to the Tibetan plateau, J. geophys. Res., 94, 17561–17579.

England, P.C. & Jackson, J.A., 1989. Active deformation of the Continents, Ann. Rev. Earth Sci., 17, 197–226.

England, P.C. & McKenzie, D.P., 1982. A thin viscous sheet model for continental deformation, Geophys. J. R. Astr. Soc., 70, 295–321.

England, P. & Molnar, P., 1990. Right-lateral shear and rotation as
the explanation for slip-strike faulting in eastern Tibet, Nature, 344, 140–142.

England, P. & Molnar, P., 1997. The active deformation in Asia: from kinematics to dynamics, Science, submitted.

Feigl, K.L., et al., 1993. Space geodetic measurement of crustal deformation in central and southern California, 1984–1992, J. geophys. Res., 98, 21 677–21 712.

Gaudemer, Y., Tapponnier, P., Meyer, B., Peltzer, G., Guo, S., Chen, Z., Dai, H. & Cifuentes, I., 1995. Partitioning of crustal slip between linked, active faults in the eastern Qilian Shan, and evidence for a major seismic gap, the 'Tianzu gap', on the western Haiyuan fault, Gansu (China), Geophys. J. Int., 120, 599–645.

Gillespie, A. & Molnar, P., 1995. Asynchronous maximum advances of mountain and continental glaciers, Rev. Geophys., 33, 311–364.

Gordon, R.G. & Stein, S., 1992. Global tectonics and space geodesy, Science, 256, 333–342.

Guseva, T.G., 1986. Contemporary Movements of the Earth's Crust in the Transition Zone from the Pamir to the Tien Shan, Inst. Phys. Earth, Akad. Sci., Moscow (in Russian).

Haines, A.J., 1982. Calculating velocity fields across plate boundaries from observed shear rates, Geophys. J. R. astr. Soc., 68, 203–209.

Haines, A.J. & Holt, W.E., 1993. A procedure to obtain the complete horizontal motion within zones of distributed deformation from the inversion of strain rate data, J. geophys. Res., 98, 12 057–12 082.

Holt, W.E. & Haines, A.J., 1993. Velocity fields in deforming Asia from the inversion of earthquake-released strains, Tectonics, 12, 1–20.

Holt, W.E., Ni, J.F., Wallace, T.C. & Haines, A.J., 1991. The active tectonics of the eastern Himalayan syntaxis and surrounding regions, J. geophys. Res., 96, 14 595–14 632.

Holt, W.E., Li, M. & Haines, A.J., 1995. Earthquake strain rates and instantaneous relative motions within central and eastern Asia, Geophys. J. Int., 122, 569–593.

Houseman, G.A. & England, P.C., 1986. Finite strain calculations of continental deformation 1. Method and general results for convergent zones, J. geophys. Res., 91, 3651–3663.

Houseman, G. & England, P., 1993. Crustal thickening versus lateral expulsion in the India-Asia continental collision, J. geophys. Res., 98, 12 233–12 249.

Huang, J. & Chen, W.-P., 1986. Source mechanisms of the Mogod earthquake sequence of 1967 and the event of 1974 July 4 in Mongolia, Geophys. J. R. astr. Soc., 84, 361–379.

Jackson, J.A. & McKenzie, D.P., 1988. The relationship between plate motions and seismic moment tensors, and the rates of active deformation in the Mediterranean and the Middle East, Geophys. J. R. astr. Soc., 93, 45–75.

Jackson, J., Haines, A.J. & Holt, W.E., 1992. The horizontal velocity field in the deforming Aegean Sea region determined from the moment tensors of earthquakes, J. geophys. Res., 97(B12), 17 657–17 684.

Jackson, J., Haines, A.J. & Holt, W.E., 1995. The accommodation of Arabia–Eurasia plate convergence in Iran, J. geophys. Res., 100, 15 205–15 220.

Johnson, M.R.W., 1986. The structural evolution of the Kumaon Konopaltsev, I.M., 1971. Measurement of movement of the earth's crust and faults, Izv. Acad. Sci. USSR Phys. Solid Earth, 1, 23–44.

Kuchai, V.K. & Trifonov, V.G., 1977. Young left-lateral strike-slip along the zone of the Darvaz-Karakul fault, Geotektonika, 3, 91–105 (in Russian).

Kurushin, R.A., Bayasgalan, A., Elzybat, M., Enhuvshin, B., Molnar, P., Bayarsayhan, Ch., Hudnut, K.W. & Jian, L., 1996. The surface rupture of the 1957 Gobi-Altay (Ih Bogd), Mongolia, Earthquake, Geol. Soc. Am. Spec. Paper submitted.

Kurushin, R.A., Bayasgalan, A., Oltibat, M., Enhuvshin, B., Molnar, P., Bayarsayhan, Ch., Hudnut, K.W. & Jian, L., 1997. The Surface Rupture of the 1957 Gobi-Altay, Mongolia, Earthquake, Geol. Soc. Am. Spec. Paper 320, in press.

Lamb, S.H., 1987. A model for tectonic rotations about a vertical axis, Earth planet. Sci. Lett., 84, 75–86.

Le Dain, A.Y., Tapponnier, P. & Molnar, P., 1984. Active faulting and tectonics of Burma and surrounding regions, J. geophys. Res., 89, 453–472.

Lillie, R.J., Johnson, G.D., Yousuf, M., Zamin, A.H.S. & Yeats, R.S., 1987. Structural development within the Himalayan foreland fold-thrust belt of Tibet, in Sedimentary Basins and Basin-Forming Mechanisms, pp. 379–392, eds Beaumont, C. & Tankard, A.J., Can. Soc. Petrol. Geol, Mem. 12.

Liu, Q., 1993. Paléoclimat et contraintes chronologiques sur les mouvements récents dans l'Ouest du Tibet: Failles du Karakorum et de Longmu Co-Gozha Co, lacs en pull-apart de Longmu Co et de Sumxi Co, PhD thesis, Université Paris VII.

Logatchev, N.A., Zorin, Y.A. & Rogozhina, V.A., 1983. Baikal rift: active or passive?—Comparison of the Baikal and Kenya rift zones, Tectonophysics, 94, 223–240.

Lyon-Caen, H. & Molnar, P., 1984. Gravity anomalies and the structure of western Tibet and the southern Tarim Basin, Geophys. Res. Lett., 11, 1251–1254.

Lyon-Caen, H. & Molnar, P., 1985. Gravity anomalies, the Indian plate, and the structure, support and evolution of the Himalaya and Gangsa Basins, Tectonics, 4, 513–538.

McKenzie, D.P., 1972. Active tectonics of the Mediterranean region, Geophys. J. R. astr. Soc., 30, 109–185.

McKenzie, D. & Jackson, J., 1980. The relationship between strain rates, crustal thickening, paleomagnetism, finite strain and fault movements within a deforming zone, Earth Planet. Sci. Lett., 65, 182–202.

Malvern, L.E., 1969. Introduction to the Mechanics of Continuous Medium, Prentice-Hall, Englewood Cliffs, NJ.

Meyer, B., 1991. Mécanismes des Grands Tremblements de Terre et du Raccourcissement Crustal Oblique au Bord Nord-est du Tibet, PhD thesis, Université Paris VI.

Meyer, B., Tapponnier, P., Gaudemer, Y., Peltzer, G., Guo, S. & Chen, Z., 1996. Rate of left-lateral movement along the easternmost segment of the Altyn Tagh fault, east of 96°E (China), Geophys. J. Int., 124, 29–44.

Molnar, P., 1987. Inversion of profiles of uplift rates for the geometry of the Altyn Tagh fault, Nature, 331, 663–670.

Molnar, P., 1988. Continental tectonics in the aftermath of plate tectonics, Nature, 335, 131–137.

Molnar, P., 1990. A review of the seismicity and the rates of underthrusting and deformation at the Himalaya, J. Himalayan Geol., 1(2), 131–154.

Molnar, P., 1992. A review of seismicity, recent faulting, and active deformation of the Tibetan Plateau, J. Himalayan Geology, 3(1), 43–78.

Molnar, P. & Chen, W.-P., 1983. Focal depths and fault plane solutions

© 1997 RAS, GJI 130, 551–582
of earthquakes under the Tibetan plateau, J. geophys. Res., 88, 1180–1196.
Molnar, P. & Deng, Q., 1984. Faulting associated with large earthquakes and the average rate of deformation in central and eastern Asia, J. geophys. Res., 89, 6203–6227.
Molnar, P. & Gipson, J.M., 1996. A bound on the rheology of continental lithosphere using very long baseline interferometry: The velocity of south China with respect to Eurasia, J. geophys. Res., 101, 545–553.
Molnar, P. & Lyon-Caen, H., 1988. Some simple physical aspects of the support, structure and evolution of mountain belts, in Processes in Continental and Lithospheric Deformation, ed. Clark, S.P., Jr, Geol. Soc. Am. Spec. Pap. 218.
Molnar, P. & Lyon-Caen, H., 1989. Fault plane solutions of earthquakes and active tectonics of the northern and eastern parts of the Tibetan plateau, Geophys. J. Int., 99, 123–153.
Molnar P. & Pandey, M.R., 1989. Rupture zones of Great Earthquakes in the Himalayan region, in Frontiers of Seismology in India, ed. Brune, J.N., Proc. Indian Academy of Sciences, 98, 61–70.
Molnar, P. & Tapponnier, P., 1978. Active tectonics of Tibet, J. geophys. Res., 83, 5361–5375.
Molnar, P., Freedman, D. & Shih, J.S.F., 1979. Lengths of intermediate and deep seismic zones and temperatures in downgoing slabs of lithosphere, Geophys. J. R. astr. Soc., 65, 41–54.
Molnar, P., Burchfiel, B.C., Liang, K. & Zhao, Z., 1987. Geomorphic evidence for active faulting in the Altyn Tagh and northern Tibet and qualitative estimates of its contribution to the convergence of India and Eurasia, Geology, 15, 249–253.
Molnar, P. & England, P. & Martinod, J., 1993. Mantle dynamics, uplift of the Tibetan plateau, and the Indian monsoon, Rev. Geophys., 31, 357–396.
Nábelek, J., Chen, W.-P. & Ye, H., The Tangshan earthquake sequence and its importance for the evolution of the North China basin, J. geophys. Res., 92, 12 615–12 628.
Nelson, M.R., McCaffrey, R. & Molnar, P., 1987. Source parameters for 11 earthquakes in the Tien Shan, Central Asia, determined by P and SH waveform inversion, J. geophys. Res., 92, 12 629–12 648.
Ni, J. & York, J.E., 1978. Cenozoic extensional tectonics of the Tibetan Plateau, J. geophys. Res., 83, 5377–5387.
Norin, E., 1946. Geological Explorations in Western Tibet, Reports from the Scientific Expedition to the Northwestern Provinces of China under the Leadership of Dr. Sven Hedin, Publ. 29, (Ill), Geological Explorations in Western Tibet, Reports from the Scientific Expedition to the Northwestern Provinces of China under the Leadership of Dr. Sven Hedin, Publ. 29, (Ill), Geology 7, Tryckeri Aktiebolaget, Thule, Stockholm.
Oldham, R.D., 1899. Report on the Great Earthquake of 12th June 1897, Mem. Geol. Surv. India, 29, Geol. Surv. India, Calcutta (reprinted 1981).
Peltzer, G. & Saucier, F., 1996. Present-day kinematics of Asia derived from geologic fault rates, J. geophys. Res., 101, 27 943–27 956.
Peltzer, G. & Tapponnier, P., 1988. Formation and evolution of strike-slip faults, rifts, and basins during the India–Asia collision: an experimental approach, J. geophys. Res., 93, 15 085–15 117.
Peltzer, G., Tapponnier, P., Zhang, Z. & Xu, Z., 1984. Neogene and Quaternary faulting in and along the Qinling Shan, Nature, 317, 500–505.
Peltzer, G., Tapponnier, P., Gaudemer, Y., Meyer, B., Guo, S., Yin, K., Chen, Z. & Dai, H., 1988. Offsets of late Quaternary morphology, rate of slip, and recurrence of large earthquakes on the Chang Ma fault (Gansu, China), J. geophys. Res., 93, 7793–7812.
Peltzer, G., Tapponnier, P. & Armijo, R., 1989. Magnitude of Late Quaternary left-lateral displacements along the north edge of Tibet, Science, 246, 1285–1289.
Ritz, J.F., Brown, E.T., Bourlos, D.L., Philip, H., Schlupp, A., Raisbeck, G.M., Yiou, F. & Ehnthuvinsh, B., 1995. Slip rates along active faults estimated with cosmic ray exposure dates: Application to the Bogd fault, Gobi-Altay, Mongolia, Geology, 23, 1019–1022.
Rothery, D.A. & Drury, S.A., 1984. The neotectonics of the Tibetan Plateau, Tectonics, 3, 19–26.
Seever, I. & Armbruster, J., 1981. Great detachment earthquakes along the Himalayan arc and long-term forecasts, in Earthquake Prediction: An International Review, Maurice Ewing Series 4, eds Simpson, D.W. & Richards, P.G., Am. geophys. Un., Washington, DC.
Shackleton, N.J. et al., 1984. Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region, Nature, 307, 620–623.
Shedlock, K.M., Hellinga, S.J. & Ye, H., 1985. Evolution of the Xiaoliao Basin, Tectonics, 4, 170–185.
Sherman, S.I. & Ruzhich, V.V., 1973. Folds and faults of the basement: West Pribaikalia, Khamar-Daban and North Mongolia, in Tectonics and Volcanism of the Southwest Part of the Baikal Rift Zone, pp. 24–35, Moscow, Nauka (in Russian).
Sich, K. & Jahn, R., 1984. Holocene activity of the San Andreas fault at Wallace Creek, California, Bull. Geol. Soc. Am., 95, 883–896.
Tapponnier, P. & Molnar, P., 1976. Slip-line field theory and large scale continental tectonics, Nature, 294, 319–324.
Tapponnier, P. & Molnar, P., 1977. Active faulting and tectonics in China, J. geophys. Res., 82, 2905–2930.
Tapponnier, P. & Molnar, P., 1979. Active faulting and Cenozoic tectonics of the Tien Shan, Mongolia and Baykal Regions, J. geophys. Res., 84, 3425–3459.
Tapponnier, P., Peltzer, G. & Armijo, R., 1986. On the mechanics of the collision between India and Asia, in Collision Tectonics, pp. 115–157, eds Coward, M.P. & Ries, A.C., Geol. Soc. Spec. Publ., 19, London.
Tapponnier, P. et al., 1990. Active thrusting and folding in the Qilian Shan, and decoupling between upper crust and mantle in north-eastern Tibet, Earth Planet. Sci. Lett., 97, 382–403.
Thomas, J.C., Perroud, H., Cobbold, P.R., Bazhenov, M.L., Burtman, V.S., Chauvin, A. & Sadybakasov, E., 1993. A paleomagnetic study of Tertiary formations from the Kyrgyz Tien-Shan and its tectonic implications, J. geophys. Res., 98, 9571–9589.
Voitovich, V.S., 1969. The Nature of the Dzungarian Deep Fault, Trans. Geol. Inst., Acad. Sci. USSR, 183, Nauka, Moscow (in Russian).
Wang, E., 1994. Late Cenozoic Xianshuihe/Xiaojiang and Red River fault systems of southwestern Sichuan and central Yunnan, China, Phd thesis, MIT.
Ward, S.N., 1994. Constraints on the tectonics of the central Mediterranean from Very Long Baseline Interferometry, Geophys. J. Int., 117, 441–452.
Wessel, P. & Smith, W.H.F., 1995. New version of the Generic Mapping Tools released, EOS, Trans. Am. geophys. Un., 76, 329.
Wilson, J.T., 1965. A new class of faults and their bearing on continental drift, Nature, 207, 343–347.
Ye, H., Shedlock, K.M., Hellinga, S.J. & Slater, J.G., 1985. The North China Basin: an example of a Cenozoic rifted intraplate basin, Tectonics, 4, 153–169.
Zhang, W., Jiao, D., Zhang, P., Molnar, P., Burchfiel, B.C., Deng, Q. & Wang, Y., 1987. Displacement along the Haiyuan fault associated with the great 1920 Haiyuan, China, earthquake, Bull. seism. Soc. Am., 77, 117–123.
Zhang, P. et al., 1988. Bounds on the Holocene slip rate of the Haiyuan fault, north-central China, Quat. Res., 30, 151–164.
Zhang, P. et al., 1990. Late Cenozoic tectonic evolution of the Ningxia-Hui Autonomous Region, China, Geol. Soc. Am. Bull., 102, 1484–1498.
Zhang, P. et al., 1991. Amount and style of late Cenozoic deformation in the Liupan Shan area, Ningxia Autonomous Region, China, Tectonics, 10, 1111–1129.
Zienkiewicz, O.C. & Taylor, R.L., 1989. The Finite Element Method, Vol. 1, McGraw-Hill.
Zonenshain, L.P. & Savostin, L.A., 1981. Geodynamics of the Baikal rift zone and plate tectonics of Asia, Tectonophysics, 76, 1–45.
Zorin, Y.A., 1981. The Baikal rift: an example of the intrusion of astheneospheric material into the lithosphere as the cause of disruption of lithospheric plates, Tectonophysics, 73, 91–104.
APPENDIX A: DISCUSSION OF SLIP RATES ON FAULTS IN DIFFERENT REGIONS

Baikal-Hövsgöl region

Deformation occurs largely by normal faulting, with extension approximately perpendicular to the local trends of the Baikal and Hövsgöl rift systems and left-lateral strike-slip faulting on the east–west segment between them. We use average extension rates of $2 \pm 1$ mm a$^{-1}$ at the northeast end of the rift zone and $2 \pm 1$ and $3 \pm 1.5$ mm a$^{-1}$ where Lake Baikal fills the rift valley. These are based on estimates of the amount of opening (15 km to perhaps as much as 25 km) since some time in the Pliocene Epoch (2–5 Ma) (e.g. Logatchev, Zorin & Rogozhina 1983; Zorin 1981) and from seismicity in this century (Chen & Molnar 1977; Molnar & Deng 1984). As neither estimate is well constrained, we assign large uncertainties. [See Baljinnyam et al. (1993) for additional discussion.] The lower rate at the northeast end seems required by the disappearance of the rift east of the Muya graben (e.g. Zonenshain & Savostin 1981). Following Sherman & Ruzhich's (1973) estimate of roughly 11 km of left-lateral slip and 11 km of vertical slip along the east–west-trending Tunka graben, just west of the lake, we assume components of both north–south crustal extension and east–west left-lateral shear at rates of $2 \pm 1$ mm a$^{-1}$. This strike-slip component seems to serve as a transform fault between the Baikal rift and three grabens in northern Mongolia that form the north–south-trending Hövsgöl graben system. We assume a rate of $1.0 \pm 0.7$ mm a$^{-1}$ of east–west opening for each. These rates, however, are guesses based largely on their being consistent with the other rates nearby.

| Coordinates | Coordinates | Fault-plane solution | Rates of movement |
|-------------|-------------|----------------------|-------------------|
| Lat. | Long. | Strike | dip | slip | Horizontal | Slip vector |
| N | E | ° | ° | ° | mm a$^{-1}$ | mm a$^{-1}$ |
| 57.2 | 120.8 | 116.0 | 116.0 | 051 | 45 | 270 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 56.6 | 116.0 | 110.0 | 047 | 45 | 270 | 1.0 ± 0.5 | 1.4 ± 0.7 |
| 56.3 | 112.3 | 109.0 | 041 | 45 | 270 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 54.0 | 109.0 | 105.0 | 046 | 45 | 270 | 3.0 ± 1.5 | 4.2 ± 2.1 |
| 51.8 | 105.0 | 99.5 | 090 | 45 | 000 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 51.8 | 105.0 | 99.5 | 090 | 45 | 270 | 2.0 ± 1.0 | 2.0 ± 1.0 |
| 51.8 | 100.5 | 100.0 | 008 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |
| 51.8 | 99.3 | 99.0 | 005 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |
| 51.8 | 98.5 | 98.0 | 008 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |

Western Mongolia: Bulnay fault, Gobi-Altay and Mongolian Altay

Baljinnyam et al. (1993) inferred rates of strike-slip faulting in Mongolia assuming that most topographic features that have been offset along the active faults formed some time since the last global glacial period. To be consistent with data from other regions, we use these rates, but our uncertainties are meant to allow for the possibility that these rates could be wrong by a factor of two to four (Gillespie & Molnar 1995).

Baljinnyam et al. (1993) inferred E–W left-lateral slip on the Bulnay fault at $5 \pm 3$ mm a$^{-1}$. Dip slip is clearly minor in this area.

For the Hangay, a high area in western Mongolia, where deformation is mild, we include roughly conjugate segments, 200–300 km in length, with $0.5 \pm 0.3$ mm a$^{-1}$ of slip, corresponding to NE–SW shortening and NW–SE extension at comparable rates. These are meant to permit a variety of styles of relatively slow deformation: (a) normal faulting, largely, but not entirely, associated with NW–SE extension; (b) right-lateral slip on northerly trending planes as for the 1967 Mogod earthquake (e.g. Huang & Chen 1986; Tapponnier & Molnar 1979); and (c) thrust faulting on NW planes, as also observed for the Mogod earthquake sequence and other small to moderate earthquakes.

For the ESE–WNW Gobi-Altay belt in southern Mongolia, we assumed both left-lateral and thrust slip on faults trending 100°, parallel to the surface rupture of the 1957 Gobi-Altay earthquake: left-lateral slip at $4 \pm 2$ mm a$^{-1}$, and $2 \pm 1$ mm a$^{-1}$ of convergence, as indicated by the existence of thrust or reverse components of slip in this area (Baljinnyam et al. 1993; Kurushin et al. 1997; Ritz et al. 1995). Using the strike-slip rate of 1.2 mm a$^{-1}$ of Ritz et al. (1995) for the Bogd fault, we assume that the area affected by this fault is roughly 1/3 to 1/4 of the Gobi Altay and that NNE–SSW shortening occurs at roughly half the rate of strike slip (Kurushin et al. 1997).

From a crude estimate of the history of faulting in the last 500 years, Baljinnyam et al. (1993) inferred that the dominant deformation in the Mongolian Altay of western Mongolia is right-lateral strike slip at a rate of about roughly 10 mm a$^{-1}$; we assumed strike slip at $6 \pm 4$ mm a$^{-1}$ and $4 \pm 2$ mm a$^{-1}$ on roughly parallel segments along the NE and SW margins of the belt. The topography and some faulting imply that crustal shortening perpendicular to the range also occurs. We assumed a total convergence rate of 5 mm a$^{-1}$, again dividing this into contributions of $3 \pm 2$ mm a$^{-1}$ and $2 \pm 1$ mm a$^{-1}$ from the two sides of the belt.

© 1997 RAS, GJI 130, 551–582
belt. For an average thickness of crust of 40 km outside the range, and mean elevation of about 2 km for the range, there should be an extra thickness of crust of about 12 km beneath it. For an average width of 250 km, this implies 75 km of shortening. If we assumed 75 km of shortening since 15 Ma, this would imply an average of 5 mm a\(^{-1}\).

Finally, we also include a long zone, nearly 1000 km in length, of modest to insignificant N–S shortening, 1 ± 1 mm a\(^{-1}\), across the NW end of the Mongolian Altay and eastwards, where elevations are high and the fault-plane solution of one earthquake suggests thrust faulting (e.g. Baljinnyam et al. 1993; Tapponnier & Molnar 1979). The 100 per cent uncertainty is meant to include all reasonable possibilities.

### Coordinates

| Coordinates | Fault-plane solution | Rates of movement |
|-------------|----------------------|-------------------|
|             | Strike  | dip  | slip | Horizontal | Slip vector |
|             | \(^\circ\) | \(^\circ\) | \(\text{mm a}^{-1}\) | \(\text{mm a}^{-1}\) |
| Lat. Long.  | Strike  | dip  | slip | Horizontal | Slip vector |
| 49.5 93.0   | 094     | 90   | 000  | 5.0 ± 3.0  | 5.0 ± 3.0   |
| 48.0 96.0   | 119     | 45   | 090  | 0.5 ± 0.3  | 0.7 ± 0.4   |
| 46.0 99.0   | 049     | 45   | 270  | 0.5 ± 0.3  | 0.7 ± 0.4   |
| 45.0 94.5   | 100     | 90   | 000  | 4.0 ± 2.0  | 4.0 ± 2.0   |
| 45.0 94.5   | 100     | 45   | 090  | 2.0 ± 1.0  | 2.0 ± 1.0   |
| 50.5 91.0   | 157     | 90   | 180  | 6.0 ± 4.0  | 6.0 ± 4.0   |
| 50.5 91.0   | 157     | 45   | 090  | 3.0 ± 2.0  | 4.2 ± 2.1   |
| 50.0 84.5   | 128     | 90   | 180  | 4.0 ± 2.0  | 4.0 ± 2.0   |
| 50.0 84.5   | 128     | 45   | 090  | 1.0 ± 1.0  | 1.4 ± 1.4   |
| 50.5 84.0   | 070     | 45   | 090  | 1.0 ± 1.0  | 1.4 ± 1.4   |
| 52.0 92.0   | 096     | 45   | 090  | 1.0 ± 1.0  | 1.4 ± 1.4   |

### Tien Shan

We assume that most deformation occurs by crustal shortening perpendicular to the range, and that rates are roughly proportional to the width of the range (Avouac et al. 1993). Thus, the rate is slowest at the eastern end and increases to a maximum in the west. The tightest constraint on the rate is provided by repeated GPS measurements across the northern 70 per cent of the western segment of the belt, which when extrapolated across the entire Tien Shan yields a rate of 20 (±6) mm a\(^{-1}\) (Abdrakhmatov et al. 1996). In addition, we assume smaller shortening farther west, 6 ± 3 mm a\(^{-1}\), north of the Talas-Ferghana fault where the belt narrows. Assuming that the rate of 20 mm a\(^{-1}\) decreases to zero east of the eastern end of the Tien Shan, we assigned linearly decreasing rates to segments proportionally towards the east of that western segment. In the central segment, field investigations also constrain the rate, although only weakly. On the south flank, Burchfiel et al. (in preparation) estimated a Quaternary shortening rate of 4.5 to 14 mm a\(^{-1}\). This ignores shortening within the range or on its north flank. Farther east, Avouac et al. (1993) inferred shortening at 3 ± 1.5 mm a\(^{-1}\) across anticlines along the northern flank near 86\(^\circ\)E, and, doubling that rate to account for thrust faulting on the southern margin, they suggested shortening at 6 ± 3 mm a\(^{-1}\) across the entire belt. Burchfiel et al. (in preparation) inferred a rate of 2 ± 1 mm a\(^{-1}\) for some of the same anticlines, but deformation along parallel series of anticlines allows additional shortening. On the south flank, however, Brown et al. (1997) dated fans offset by scarps 8–43 m high, and deduced that shortening faster than 2 mm a\(^{-1}\) seems unlikely there. Although the sum of the rates estimated by Brown et al., Burchfiel et al., and Avouac et al. (1993) implies a total convergence rate of only 4–6 mm a\(^{-1}\), allowing for internal deformation of the range permits a higher convergence rate. In fact, the variety of styles of deformation along the belt requires such internal deformation (Brown et al. 1997; Burchfiel et al. in preparation). We assume shortening at 10 ± 5 mm a\(^{-1}\) in these segments.

Two major strike-slip faults cross parts of the Tien Shan, and both terminate in thrust zones where additional crustal thickening seems to occur. The right-lateral Talas-Ferghana fault bounds the western end of the Tien Shan, separating it from the Ferghana basin and the Chatkal Ranges (e.g., Burtman 1963, 1964, 1975). Measured offsets of gullies and ridges and radiocarbon ages of material associated with them imply slip at 10 (±3) mm a\(^{-1}\) (Burtman et al. 1996). Slip along the Talas-Ferghana fault must be absorbed at both ends of the fault, where it clearly terminates. Like Burtman et al. (1996) and Thomas et al. (1993), we assume that the Ferghana valley west of the Talas-Ferghana fault rotates about an axis near the SW end of the Chatkal ranges. We assume further that crustal shortening in the Chatkal ranges absorbs the full 10 ± 3 mm a\(^{-1}\) of slip on the Talas-Ferghana fault and decreases to nil at the SW end, yielding an average of 5 ± 3 mm a\(^{-1}\) of convergence. For the Dzhangarian fault, Voitovich's (1969) estimate of 7 km of right-lateral slip since Pliocene time implies a rate of 3 ± 1 mm a\(^{-1}\). This fault seems to die out in a range of mountains just west of the fault, the Dzhangarian Alatau, across which we assume N–S shortening at 1.4 ± 0.7 mm a\(^{-1}\). Farther northeast, right-lateral slip associated with the Zaysan earthquake of 1990 (Chen & Kao 1996) suggests the presence of another right-lateral strike-slip fault, within an en echelon zone of east–west-trending high topography. We assume...
north-south shortening at $2 \pm 1 \text{ mm a}^{-1}$ across this mountainous region, approximately 350 km long, between the Dzungarian Alatau and the Mongolian Altay.

| Co ordinates | Co ordinates | Fault-plane solution | Rates of movement |
|--------------|--------------|----------------------|------------------|
| Lat. Long. | Lat. Long. | Strike dip slip | Horizontal Slip vector |
| °N °E | °N °E | ° ° | mm a$^{-1}$ | mm a$^{-1}$ |
| 43.5 95.0 | 43.7 90.0 | 091 45 090 | 5.0 $\pm$ 2.5 | 7.1 $\pm$ 3.6 |
| 43.7 90.0 | 43.3 85.0 | 084 45 090 | 10.0 $\pm$ 5.0 | 14.1 $\pm$ 7.1 |
| 43.3 85.0 | 42.2 79.0 | 079 45 090 | 15.0 $\pm$ 8.0 | 21.2 $\pm$ 11.3 |
| 42.2 79.0 | 41.0 75.0 | 079 45 090 | 20.0 $\pm$ 6.0 | 28.3 $\pm$ 8.5 |
| 42.1 75.0 | 42.1 72.0 | 090 45 090 | 6.0 $\pm$ 3.0 | 8.5 $\pm$ 4.2 |
| 42.2 71.4 | 41.5 73.3 | 116 90 180 | 10.0 $\pm$ 3.0 | 10.0 $\pm$ 3.0 |
| 41.5 73.3 | 40.6 74.7 | 130 90 180 | 10.0 $\pm$ 3.0 | 10.0 $\pm$ 3.0 |
| 40.5 70.0 | 42.0 72.0 | 045 45 090 | 5.0 $\pm$ 3.0 | 7.1 $\pm$ 4.2 |
| 44.2 83.6 | 46.5 80.1 | 132 90 180 | 3.0 $\pm$ 1.0 | 3.0 $\pm$ 1.0 |
| 44.3 78.5 | 45.3 81.8 | 066 45 090 | 1.4 $\pm$ 0.7 | 2.0 $\pm$ 1.0 |
| 47.6 81.3 | 47.0 86.0 | 099 45 090 | 2.0 $\pm$ 1.0 | 2.8 $\pm$ 1.4 |

Pamir and South Tien Shan

Fault-plane solutions of earthquakes imply that the Ferghana basin is being underthrust southwards beneath the south Tien Shan along faults dipping gently south (e.g. Burtman & Molnar 1993; Nelson, McCaffrey & Molnar 1987). We assume relatively slow convergence ($3 \pm 1 \text{ mm a}^{-1}$) along the margin of the south Tien Shan and Ferghana basin.

Following Burtman & Molnar (1993), we assume that roughly half of the 44 mm a$^{-1}$ of convergence (DeMets et al. 1990) occurs in the zone spanning the Alai Valley and the Trans-Alai Range at the northern edge of the Pamir (22 $\pm$ 6 mm a$^{-1}$). This is justified by rapid slip on the left-lateral Darvaz strike-slip fault (Kuchai & Trifonov 1977), which follows the western edge of the Pamir, and by rapid, geodetically measured, convergence across the Garm region spanning the northern edge of the Pamir (Guseva 1986; Konopaltsev 1971a,b). (See Burtman & Molnar 1993 for further details.)

For the northeastern margin of the Pamir, in western China, fault-plane solutions of a few earthquakes imply NE–SW shortening (Burtman & Molnar 1993). We arbitrarily assume 5 $\pm$ 5 mm a$^{-1}$ of such convergence. For the strike-slip component, we consider two possibilities: rates of 15 to 22 mm a$^{-1}$ estimated by Liu (1993) from offset features mostly recognized on the SPOT imagery and ages of 10 kyr assigned to them, and lower, but comparably uncertain, rates of 5 to 7.3 mm a$^{-1}$.

| Coordinates | Coordinates | Fault-plane solution | Rates of movement |
|--------------|--------------|----------------------|------------------|
| Lat. Long. | Lat. Long. | Strike dip slip | Horizontal Slip vector |
| °N °E | °N °E | ° ° | mm a$^{-1}$ | mm a$^{-1}$ |
| 40.0 70.0 | 40.4 74.0 | 082 10 090 | 3.0 $\pm$ 1.0 | 3.0 $\pm$ 1.0 |
| 39.0 70.0 | 39.5 74.5 | 080 10 090 | 22.0 $\pm$ 6.0 | 22.0 $\pm$ 6.0 |
| 39.4 75.1 | 37.0 77.5 | 141 45 090 | 5.0 $\pm$ 5.0 | 7.1 $\pm$ 1.0 |
| 39.3 74.2 | 38.2 75.4 | 140 90 180 | 15.0 $\pm$ 10.0 | 15.0 $\pm$ 10.0 |
| 38.1 75.2 | 36.9 75.5 | 168 90 180 | 22.0 $\pm$ 10.0 | 22.0 $\pm$ 10.0 |
| 36.9 75.5 | 36.0 76.5 | 138 90 180 | 22.0 $\pm$ 10.0 | 22.0 $\pm$ 10.0 |

† Rate one third as large also used in some calculations

The Alty n Tagh fault and surroundings

Most of the deformation along the northern margin of the Tibetan Plateau is strike-slip, but convergence also occurs in both the western and eastern parts.

In the western part, we assume a convergence rate $8 \pm 4 \text{ mm a}^{-1}$, which seems reasonable given the many tens to perhaps more than 100 km of shortening (Bourjot 1991; Lyon-Caen & Molnar 1984) and the likelihood that the shortening is young (Avouac & Peltzer 1993; Norin 1946). Avouac & Tapponnier (1993) deduced a virtually identical rate of 7.9 mm a$^{-1}$. Nevertheless, the large uncertainty allows for anything from negligible active shortening to a rate comparable with that of most ranges in Asia.

The westernmost segment of the Alty n Tagh fault, through the Karakax valley, appears to be the best studied (Avouac 1991; Peltzer et al. 1989). This strand continues to the NW and curves to become nearly parallel to the Karakorum fault in that area (Liu 1993), a puzzling juxtaposition of adjacent, nearly parallel right- and left-lateral faults. Avouac & Tapponnier (1993) reported a rate of 19.9 mm a$^{-1}$ for the Karakax segment, but assigned it a very large uncertainty of 20 mm a$^{-1}$. We do the same.

From topographic features offset by the main ENE segment of the Alty n Tagh fault, Peltzer et al. (1989) inferred a rate of 20–30 mm a$^{-1}$, assuming that the features formed since the last glaciation. Avouac (1991) regrouped the measured offsets and

© 1997 RAS, GJI 130, 551–582
deduced a higher rate, 29.8 ± 10.0 mm a⁻¹ (Avouac & Tapponnier 1993). We also experimented with a rate three times lower. At the eastern end of the Altyn Tagh fault, the slip rate must decrease. We used 9 ± 6 mm a⁻¹ for the segment between the Altyn Tagh and the northwest end of the Nan Shan. The large uncertainty permits both overestimates and underestimates. For the easternmost part of the main strand, we use 4 ± 2 mm a⁻¹, given by Meyer et al. (1996). In addition, Peltzer et al. (1989) suggested ~5 mm a⁻¹ for a splay emanating from the Altyn Tagh fault and curving into the Nan Shan; we assigned it an uncertainty of 3 mm a⁻¹.

N–S shortening also occurs across the Altyn Tagh, which Molnar et al. (1987) crudely estimated to occur at 2 to 10 mm a⁻¹; 6 ± 4 mm a⁻¹ permits both a negligible value and as much as 10 mm a⁻¹, which, nevertheless, is less than the 13.5 mm a⁻¹ deduced by Avouac & Tapponnier (1993). Farther east, we allowed N–S shortening of 3 ± 3 mm a⁻¹ across this eastern portion of the Altyn Tagh fault, which also permits negligible or modest shortening.

### The Tibetan Plateau

#### Northern Tibet

In westernmost Tibet, Avouac (1991) and Avouac & Tapponnier (1993) reported a system of strike-slip and normal faults that connect the Altyn Tagh and Karakorum faults. They estimated 7 mm a⁻¹ of left-lateral slip on the ENE-trending Longma-Gozha Co fault zone, which is interrupted by grabens along its eastern end. To account for this deformation, we used 8.5 ± 5.0 mm a⁻¹ of left-lateral slip on an ENE-trending fault ~300 km in length and a northeast-trending zone of extension ~200 km in length with opening at the same rate. The large uncertainties reflect our inability to evaluate the inferred rates.

Studies of earthquakes within northern Tibet suggest that deformation is roughly equally partitioned into normal and strike-slip faults, with the T-axes oriented approximately N115° (Molnar & Chen 1983; Molnar & Lyon-Caen 1989; Molnar & Tapponnier 1978, Ni & York 1978), which accords with Rothery & Drury's (1984) interpretation of the Landsat imagery. Rather than assign a rate to every minor graben or strike-slip fault in this region, we used regional trends of grabens or linear fault zones shown by Armijo et al. (1988) to define zones of relatively slow normal or strike-slip faulting: 1.0 ± 0.7 mm a⁻¹ for horizontal components on each. The less prominent faulting recognized on the Landsat imagery of northern compared with southern Tibet (e.g. Armijo et al. 1986, 1989; Tapponnier et al. 1986) favours a lower rate than that of 10 to 18 mm a⁻¹ across southern Tibet.

| Coordinates | Fault-plane solution | Rates of movement |
|-------------|---------------------|------------------|
| Lat. Long. | Strike | dip | slip | Horizontal | Slip vector |
| 34.2 | 78.0 | 34.6 | 81.0 | 080 | 90 | 000 | 8.5 ± 5.0 | 8.5 ± 5.0 |
| 34.6 | 81.0 | 36.0 | 82.0 | 030 | 45 | 270 | 8.5 ± 5.0 | 12.0 ± 7.1 |
| 34.5 | 81.6 | 32.7 | 83.0 | 000 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |
| 35.3 | 84.2 | 32.5 | 85.5 | 000 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |
| 35.3 | 84.2 | 32.5 | 85.5 | 158 | 90 | 180 | 1.0 ± 0.7 | 1.0 ± 0.7 |
| 35.9 | 85.4 | 33.3 | 86.7 | 165 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |
| 35.5 | 85.5 | 32.1 | 87.0 | 000 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |
| 32.2 | 87.5 | 33.6 | 88.8 | 000 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |
| 34.2 | 87.5 | 33.6 | 88.8 | 119 | 90 | 180 | 1.0 ± 0.7 | 1.0 ± 0.7 |
| 33.4 | 89.5 | 31.6 | 94.0 | 114 | 90 | 000 | 1.0 ± 0.7 | 1.0 ± 0.7 |
| 31.5 | 90.5 | 32.3 | 92.0 | 000 | 45 | 270 | 1.0 ± 0.7 | 1.4 ± 1.0 |

† Rate one third as large also used in some calculations

© 1997 RAS, GJI 130, 551–582
Karakorum fault

A prominent NW strike-slip fault, the Karakorum fault, can be traced as a nearly linear feature on the satellite imagery for about 700 km and seems to mark the western end of normal faulting within the high terrain of the Tibetan Plateau. From topographic features offset along the fault and assuming an age of 10 ka for them, Liu (1993) inferred right-lateral slip rates of 22 mm a\(^{-1}\) in the NW segment, where the fault is relatively inaccessible, 35 ± 10 mm a\(^{-1}\) in the central segment where the fault is particularly clear, and 30 ± 10 mm a\(^{-1}\) in the SE segment, where a component of normal faulting has left a long, narrow, NW-trending valley. Avouac & Tapponnier (1993) assumed 32 ± 8 mm a\(^{-1}\) for slip on this fault. As discussed below with normal faulting in southern Tibet, we also include a modest component of normal faulting suggested by Armijo et al. (1986). We ignore the discrepancy between the rate of extension perpendicular to the Karakorum fault implied by the rate and orientation of relative movement postulated by Liu (≈10–15 mm a\(^{-1}\) of extension) and the much lower rate suggested by Armijo et al. (1986) of only 1–2 mm a\(^{-1}\).

These rates of strike slip in excess of 30 mm a\(^{-1}\) are much higher than the bound of 9 ± 4 mm a\(^{-1}\) deduced by Molnar & Lyon-Caen (1989) from the smooth variation in orientations of underthrusting along the Himalaya. Part of the difference between this bound and the 30–35 mm a\(^{-1}\) estimate of Liu (1993) can be reconciled by allowing for decreasing rate towards the southeast end of the Karakorum fault with strain absorbed within Tibet, as Armijo et al. (1989) imagined. Nevertheless, the 30 mm a\(^{-1}\) reported by Liu for the SE portion of the fault requires a large right-lateral component of slip parallel to the Himalaya east of the fault. We used both Liu’s rates and those reduced by three times.

### Coordinates

| Lat. | Long. | Strike | Dip | Slip | Horizontal | Slip vector |
|------|-------|--------|-----|------|------------|-------------|
| °N   | °E    | °      | °   | °    | mm a\(^{-1}\)| mm a\(^{-1}\) |
| 36.0 | 76.5  | 149    | 90  | 180  | 22.0 ± 10.0 | 122.0 ± 10.0 |
| 34.2 | 78.1  | 142    | 90  | 180  | 35.0 ± 10.0 | 135.0 ± 10.0 |
| 32.2 | 79.9  | 120    | 90  | 180  | 30.0 ± 15.0 | 130.0 ± 15.0 |

\(^{\dagger}\) Rates one third as large also used in some calculations

### Central Tibet

Armijo et al. (1989) mapped ESE right-lateral fault zones across central Tibet. One ruptured in a major earthquake in 1951, which left a fresh, unambiguous scarp. They visited two others, to its east and west, and documented clear evidence for right-lateral slip on them as well. The two westernmost faults were inferred largely from their expression on the Landsat imagery. Armijo et al. (1989) deduced right-lateral slip rates of 10–20 mm a\(^{-1}\) on them, from offset features to which they assigned ages of thousands of years, 10 ka (last deglaciation), and 125 ka (penultimate deglaciation). Molnar (1992) suggested that these ages might have been underestimated by a factor of 5 to 10, and he suggested that these offset features might instead be assigned to the last deglaciation, to the last interglacial, and to roughly 1 Ma. He reasoned that strike-slip rates of 10–20 mm a\(^{-1}\) that terminate near the ends of the graben systems in southern Tibet are incompatible with rates of extension of only 1–2 mm a\(^{-1}\). We considered rates of 15 ± 5 mm a\(^{-1}\) and 5 ± 5 mm a\(^{-1}\).

### Southern Tibet

Following Armijo et al. (1986) we assigned a rate of extension of 1.6 ± 0.8 mm a\(^{-1}\) to each of the main grabens in southern Tibet and an average orientation of extension of N095\(^{\circ}\). This rate and the orientation are consistent with those inferred by Molnar & Lyon-Caen (1989) from earthquake slip vectors and their estimated rate of convergence at the Himalaya.
The Himalaya

We divided most of the Himalaya into segments 300–400 km in length and assigned rates increasing from west to east. We also included a short NNE segment of thrust faulting at 3.5 ± 3.5 mm a⁻¹ within the Nanga Parbat region (Butler & Prior 1988).

The most precisely measured segment of underthrusting of India beneath the Himalaya appears to be from the NW. Measured offsets along a cross-section across the Salt range, south of the Himalayan front, using seismic reflection profiles (e.g. Lillie et al. 1987), and the dates of the deformed rock (Johnson, Raynolds & Burbank 1986), give a rate of underthrusting of 10 ± 2 mm a⁻¹ of India beneath the southern margin of the western Himalaya (Baker et al. 1988).

From the ages of sediment in drill holes in the foreland basin in front of the Himalaya (the Ganga basin), and the assumption that the flexural rigidity of the Indian Plate had not changed with time, Lyon-Caen & Molnar (1985) obtained a rate of 10 to 25 mm a⁻¹ for the underthrusting of the Indian Shield beneath the front of the Himalaya. Strictly, this range is an average for the last 15–25 Myr, although it is consistent with the historic record of seismicity and crude estimates of seismic moments (e.g. Molnar 1990; Molnar & Pandey 1989; Seeger & Armbruster 1981). Moreover, it applies most reliably to the western part of the Himalaya, where boreholes are numerous. To allow for internal deformation within the Himalaya, which has manifested itself as folding of the rock units within the Lesser Himalaya (e.g. Johnson 1986), Molnar (1987) suggested adding a few mm a⁻¹, making 18 ± 7 mm a⁻¹ a convenient summary that embraces all supposed ignorance. The orientation of underthrusting, as indicated by slip vectors of earthquakes, is perpendicular to the chain except at the easternmost end (Baranowski et al. 1984; Molnar & Lyon-Caen 1989).

Lavé and Avouac have measured late Quaternary shortening across anticlines of the sub-Himalaya of Nepal at 21.5 ± 1.5 mm a⁻¹ (J.-P. Avouac, private communication, 1995). We assign an uncertainty of 7 mm a⁻¹, in part to allow for deformation farther north within the Himalaya, but also because we have not seen the data on which this inference is based. In addition, using GPS measurements 4 years apart, Bilham et al. (1997) measured convergence across the Nepali Himalaya ~100–200 km farther east, and, with assumptions about elastic strain accumulation, their measurements place a tight lower bound of 20 ± 2 mm a⁻¹ on shortening across this part of the Himalaya. Taking into account the results of Lavé & Avouac and of Bilham et al. (1997), we assume a roughly linearly increasing rate of convergence along the Himalaya: from 18 mm a⁻¹ where the ages of material at the base of boreholes suggest such a rate (Lyon-Caen & Molnar 1985), to 22 mm a⁻¹ where Lavé & Avouac worked and 23 ± 7 mm a⁻¹ where Bilham et al. (1997) reported a minimum of 20 ± 2 mm a⁻¹, to 25 ± 7 mm a⁻¹ across the eastern 500 km of the belt.

We also include modest N–S shortening at the southern edge of the Shillong Plateau (3.0 ± 3.0 mm a⁻¹), where seismicity and fault-plane solutions require active thrust faulting (e.g. Chen & Molnar 1990; Oldham 1899).
Nan Shan (or Qilian Shan)

For rates of thrust slip, we rely on the work of Meyer (1991) and Tapponnier et al. (1990) and assume shortening to be perpendicular to the mean orientation of fault-bounded ranges in this area. Meyer (1991) and Tapponnier et al. (1990) estimated rates of vertical movement at the edges of ranges of ~2 mm a⁻¹. More recently, Gaudemer et al. (1995) studied the SE end of this zone and deduced a rate of 4 ± 2 mm a⁻¹. Meyer (1991) reported several other ranges with comparable thrust faults bounding them. We assume 4 ± 2 mm a⁻¹ of shortening for the range-bounding system on the northeast (Gaudemer et al. 1995) and 30 ± 15 mm a⁻¹ for five other ranges within the Nan Shan. In addition, we allow for 2 ± 1 mm a⁻¹ of shortening within the Qaidam basin. Where all thrust systems overlap, we allow for 21 mm a⁻¹ of NNE-SSW shortening, decreasing to 14 mm a⁻¹ at the southeast end of the region, consistent with Meyer's (1991) southeastward decrease in total shortening across the region.

In addition to the shortening across the Nan Shan, left-lateral strike-slip faulting also occurs. For the oblique strike-slip faulting along the segment of the 1932 Chang-ma earthquake rupture zone, we used 5.0 ± 1.5 mm a⁻¹, based on the detailed study by Peltzer et al. (1988) and augmented by Meyer (1991). In addition, the Haiyuan fault, discussed below, extends westwards into this region (Gaudemer et al. 1995; Meyer 1991).

| Coordinates | Coordinates | Fault-plane solution | Rates of movement |
|-------------|-------------|----------------------|-------------------|
| Lat. Long.  | Lat. Long.  | Strike dip slip      | Horizontal Slip vector |
| °N °E       | °N °E       | ° ° °              | mm a⁻¹ mm a⁻¹      |
| 40.0 97.0   | 37.6 103.0  | 115 45 090         | 4.0 ± 2.0 5.6 ± 2.8|
| 39.8 96.6   | 37.8 101.8  | 115 45 090         | 3.0 ± 1.5 4.2 ± 2.1|
| 39.6 96.0   | 38.5 99.0   | 114 45 090         | 3.0 ± 1.5 4.2 ± 2.1|
| 39.5 95.5   | 38.5 97.7   | 120 45 090         | 3.0 ± 1.5 4.2 ± 2.1|
| 39.3 95.0   | 38.1 96.9   | 128 45 090         | 3.0 ± 1.5 4.2 ± 2.1|
| 39.1 93.0   | 37.2 98.0   | 113 45 090         | 3.0 ± 1.5 4.2 ± 2.1|
| 37.2 98.0   | 37.2 101.0  | 090 45 090         | 3.0 ± 1.5 4.2 ± 2.1|
| 37.2 101.0  | 36.8 102.3  | 110 45 090         | 2.0 ± 1.0 2.8 ± 1.4|
| 37.0 97.6   | 36.0 103.0  | 103 45 090         | 3.0 ± 1.5 4.2 ± 2.1|
| 39.0 92.0   | 36.5 97.0   | 120 45 090         | 2.0 ± 1.0 2.8 ± 1.4|
| 36.5 97.0   | 35.5 101.8  | 104 45 090         | 2.0 ± 1.0 2.8 ± 1.4|
| 39.6 97.4   | 39.5 98.2   | 279 73 144         | 5.0 ± 1.5 6.2 ± 1.8|

The Kunlun fault system

Kidd & Molnar (1988) measured a 30 km offset of till that they assumed to date from 2.4 Ma, the age of the oldest ice-rafted debris in the North Atlantic and the date when late Cenozoic continental glaciation is presumed to have developed in Europe (Shackleton et al. 1984). A somewhat greater date of rapid change in δ¹⁸O suggests a somewhat greater age, ~3 Ma, for the onset of cooling and glaciation in the Northern Hemisphere. To include such a possibility, we assume an uncertainty of 5 mm a⁻¹. We extrapolated the fault east of the section, where it is clear on the Landsat imagery, to pass through an earthquake with an appropriate fault-plane solution (Molnar 1988) and assumed a slower rate (6 ± 1 mm a⁻¹) for this less clear segment.

The Haiyuan fault

A major strike-slip fault extends from the southern Nan Shan eastwards to the Liupan Shan in western Gansu. A major earthquake in 1920, near the city of Haiyuan gives the fault its name. For the westernmost segment, Meyer (1991) suggested a rate of 18 ± 8 mm a⁻¹ from an offset of 235 ± 40 m and an assumed postglacial age, but he also permitted a rate of only a few mm a⁻¹, if features offset only 40 m are postglacial in age. To allow for both possibilities, we used 10 ± 10 mm a⁻¹. For the central segment, Gaudemer et al. (1995) inferred a rate of 11 ± 3 mm a⁻¹ near 101°E, but because offset features are not well dated, we use a larger uncertainty, 6 mm a⁻¹. Gaudemer et al. (1995) also reported 4.3 ± 2.1 mm a⁻¹ of strike slip on the more easterly trending Gulang.
fault. The eastern segment of the Haiyuan fault not only is clear on the satellite imagery (e.g. Tapponnier & Molnar 1977), but 8 m offsets associated with the 1920 Haiyuan earthquake (Zhang et al. 1987) define a sharp trace. Using 10–15 km offsets of late Pliocene/early Quaternary and older formations, Burchfiel et al. (1991) determined an average rate of 5–10 mm a⁻¹ for the Quaternary period. From offset stream channels dated with radiocarbon, Zhang et al. (1988) obtained a late Quaternary rate of 8 ± 2 mm a⁻¹.

Farther north, another strike-slip strand seems to slip at a lower rate (Zhang et al. 1990); we use 3.0 ± 1.5 mm a⁻¹. This strike-slip component becomes one of thrust slip farther east, and we assume a comparable rate of shortening (Zhang et al. 1991).

At the eastern end of the Haiyuan fault, strike slip is converted into thrust slip along the Liupan Shan. For a rate of shortening, we use the same rate of strike-slip faulting on the Haiyuan fault, 8 ± 3 mm a⁻¹, because cross-sections across the Liupan Shan suggest that all of the strike slip on the Haiyuan fault is absorbed by shortening (Zhang et al. 1991). The larger uncertainty allows for a poorly constrained orientation of shortening.

The Xianshuihe fault and Longmen Shan

Allen et al. (1991) measured late Quaternary left-lateral offsets along the main, NW-striking strand of the Xianshuihe fault, and estimated a rate of 15 ± 5 mm a⁻¹ using relevant, published radiocarbon and thermoluminescence dates. This rate includes that obtained from offsets associated with earthquakes in the twentieth century (Molnar & Deng 1984). We assumed 15 ± 5 mm a⁻¹ for the NNW–SSE-trending, southeast continuation of the fault. Left-lateral slip also continues to the northwest on parallel fault strands, which can be seen on the Landsat imagery (Tapponnier & Molnar 1977), but rates seem to be lower (Wang 1994). We assume 5 ± 3 mm a⁻¹.

Convergence clearly occurs across the Longmenshan, but recent deformation appears to be mild (Burchfiel et al. 1996), and recent GPS measurements corroborate low rates (King et al. 1997). We assume shortening at 2.5 ± 2.0 mm a⁻¹ to include possibilities both of negligibly small and of moderate rates. In addition, the orientations of faults and folds seen on the Landsat imagery strongly suggest a component of right-lateral slip parallel to the range (Tapponnier & Molnar 1977), and we include a similarly small and uncertain strike-slip component (2.5 ± 2.0 mm a⁻¹).

| Coordinates | Fault-plane solution | Rates of movement |
|-------------|----------------------|------------------|
| Lat. Long.  | Strike dip slip      | Horizontal slip vector |
| °N °E       | ° ° °               | mm a⁻¹ mm a⁻¹    |
| 38.5        | 97.9 38.4 100.1      | 092 90 000 10.0 ± 10.0 |
| 38.4        | 100.1 37.1 103.0     | 119 90 000 11.0 ± 6.0 |
| 37.1        | 103.0 36.7 105.0     | 103 90 000 8.0 ± 2.0 |
| 36.7        | 105.0 36.1 106.0     | 121 90 000 8.0 ± 2.0 |
| 37.3        | 102.3 37.3 103.5     | 090 90 000 4.3 ± 2.1 |
| 37.5        | 103.3 37.4 105.2     | 090 90 000 3.0 ± 1.5 |
| 37.4        | 105.2 36.6 106.0     | 141 45 090 3.0 ± 1.5 |
| 36.1        | 106.0 34.7 106.2     | 353 45 090 8.0 ± 3.0 |

Strike-slip, thrust, and normal faulting in Burma and Yunnan

Faulting in this area includes a spectrum of styles and in general none of the deformation is quantified well. For the southern continuations of the Xianshuihe fault, which trend roughly N–S, we use the rate of 5 mm a⁻¹ given by Wang (1994), but with a large uncertainty (3 mm a⁻¹), because he relied on qualitative estimates of ages.

For the normal faulting, both near Kunming and farther northwest, we use rates similar to those of Armijo et al. (1986) for grabens in Tibet. Because the graben systems in Yunnan are spread over areas roughly twice as wide as those in Tibet, we assume rates double those in Tibet, but total rates of 3.0 ± 1.5 mm a⁻¹ across each clearly are guesses.

We divided the Red River fault into three segments and use 4 ± 2 mm a⁻¹, based on the range of 2 to 5 mm a⁻¹ estimated by Allen et al. (1984). We also assumed a strike-slip continuation from the NW set of grabens in Yunnan to the southeast end of the
right-lateral fault mapped by Armijo et al. (1989) near the eastern Himalayan syntaxis. For a rate, we used 5 ± 3 mm a⁻¹, a compromise between rates for the Red River fault and the right-lateral faults in Tibet.

Deformation in Yunnan west of the Red River fault seems to occur primarily by left-lateral slip on ENE faults (e.g. Le Dain, Tapponnier & Molnar 1984). Fault-plane solutions of earthquakes confirm such slip (Holt et al. 1991). We assume 3.0 ± 2.0 mm a⁻¹ of such slip on five such sub-parallel faults, using the interpretation of Landsat imagery by Le Dain et al. (1984) to define the locations of such faults. As Holt & Haines (1993; Holt et al. 1991) showed, these faults seem to absorb right-lateral shear on a roughly north–south zone.

Most of the right-lateral slip of India past South China, however, occurs along the Sagaing fault, the most important fault in this region. We use a rate of 40 (±10) mm a⁻¹ for it, based on the current spreading rate in the Andaman Sea reported by Curran et al. (1978) and updated by J. R. Curran (private communication 1996). Moreover, we extend this fault north of 26°N, where it is clear on the Landsat imagery, assuming that such rapid slip must end near the syntaxis. Fault-plane solutions indicate strike-slip faulting along a subparallel fault that marks the eastern edge of the Indo-Burman ranges and the wide basin of central Burma (Le Dain et al. 1984). We assume a very uncertain rate of 5.0 ± 5.0 mm a⁻¹ for slip on it.

We also assume westward overthrusting of the Indo-Burman ranges onto eastern India and Bangladesh, at a modest, uncertain rate of 5.0 ± 5.0 mm a⁻¹. Fault-plane solutions show deformation within the Indian plate, beneath the Indo-Burman ranges, not underthrusting of India beneath the Indo-Burman ranges (e.g. Chen & Molnar 1990; Le Dain et al. 1984). Because seismicity gives no indication of such underthrusting, we allow for a negligible rate of slip.

### Coordinates and Fault-plane solutions

| Lat. °N | Long. °E | Strike ° | dip ° | Slip °/mm a⁻¹ |
|-------|---------|---------|------|--------------|
| 29.2  | 102.2   | 180     | 90   | 000          |
| 28.0  | 102.2   | 154     | 90   | 000          |
| 26.3  | 103.1   | 180     | 90   | 000          |
| 24.0  | 102.9   | 016     | 45   | 270          |
| 26.0  | 100.0   | 143     | 90   | 180          |
| 23.5  | 102.0   | 116     | 90   | 180          |
| 22.9  | 103.3   | 129     | 90   | 180          |
| 26.7  | 100.5   | 010     | 45   | 270          |
| 27.2  | 100.2   | 010     | 45   | 270          |
| 29.9  | 95.5    | 123     | 90   | 180          |
| 23.8  | 97.0    | 078     | 90   | 000          |
| 23.0  | 97.0    | 075     | 90   | 000          |
| 23.5  | 99.0    | 050     | 90   | 000          |
| 21.7  | 98.0    | 070     | 90   | 000          |
| 20.8  | 98.5    | 077     | 90   | 000          |
| 20.0  | 98.0    | 070     | 90   | 000          |
| 20.0  | 96.0    | 000     | 90   | 180          |
| 24.0  | 96.1    | 015     | 90   | 180          |
| 20.0  | 94.3    | 000     | 90   | 180          |
| 24.0  | 94.4    | 025     | 90   | 180          |
| 27.6  | 96.4    | 050     | 10   | 090          |
| 26.0  | 94.0    | 030     | 10   | 090          |
| 23.5  | 93.0    | 345     | 10   | 090          |

### Normal faulting surrounding the Ordos Plateau and strike-slip faulting in North China

The Ordos Plateau is surrounded by grabens, with clear evidence of normal faulting, on nearly its complete perimeter. In the northwest, the rupture of a segment of the Great Wall of China indicates a right-lateral strike-slip component, but only approximately one quarter of the vertical component (e.g. Deng et al. 1984). For the other grabens, we assume pure normal faulting. The grabens are aligned northeast–southwest, and using the regional strikes recognized on the Landsat imagery, we divided them into individual segments. We take 2 ± 1 mm a⁻¹ as the rate for the horizontal component at most segments, assuming that these grabens are slightly larger than those in southern Tibet, studied by Armijo et al. (1986), but smaller than those in the Baikal Rift Zone. For the northeasternmost and southernmost segments, where relief is high and faulting is especially impressive, we allow for a larger rate: 3.0 ± 1.5 mm a⁻¹. In addition, we assume a rate of 5 ± 3 mm a⁻¹ along a clearly active strike-slip fault south of the graben system (e.g. Tapponnier & Molnar 1977). Peltzer et al. (1984) suggested a larger offset than that determined by Burov et al. (1991) for the Haiyuan fault, which might suggest a higher slip rate. The large uncertainty is meant to allow for this possibility.

© 1997 RAS, GJI 130, 551-582
To constrain rates and orientations of deformation in North China, we use analyses of earthquakes by Chen & Nábelek (1988; Nábelek, Chen & Ye 1987) as a guide. NE-trending right-lateral strike-slip faulting seems to be the dominant style of deformation, but extension presumably with a roughly N–S orientation also occurs (e.g. Tapponnier & Molnar 1977). We assumed N–S extension for two normal fault segments. We took the rates of 2.5 (±1) mm a⁻¹ of right-lateral shear, estimated by Chen & Nábelek (1988) from a summation of seismic moments of earthquakes. This rate is comparable with the average rate obtained from estimates of 30 per cent extension of the region since Eocene time (Shedlock, Hellinger & Ye 1985; Ye et al. 1985). Given that there may have been an acceleration in extension (and subsidence) in Quaternary time (Shedlock et al. 1985; Ye et al. 1985), we assigned uncertainties large enough to include this possibility.

| Coordinates | Fault-plane solution | Rates of movement |
|-------------|----------------------|-------------------|
| Lat. | Long. | Strike | dip | slip | Horizontal | Slip vector |
| °N | °E | ° | ° | ° | mm a⁻¹ | mm a⁻¹ |
| 38.0 | 106.0 | 019 | 45 | 259 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 40.2 | 106.0 | 050 | 45 | 270 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 40.6 | 108.8 | 075 | 45 | 270 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 39.3 | 112.3 | 055 | 45 | 270 | 3.0 ± 1.5 | 4.2 ± 2.1 |
| 39.0 | 112.3 | 060 | 45 | 270 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 36.3 | 111.5 | 034 | 45 | 270 | 2.0 ± 1.0 | 2.8 ± 1.4 |
| 34.0 | 109.0 | 052 | 45 | 270 | 3.0 ± 1.5 | 4.2 ± 2.1 |
| 34.3 | 106.0 | 099 | 90 | 000 | 5.0 ± 3.0 | 5.0 ± 3.0 |
| 35.1 | 114.1 | 013 | 90 | 180 | 2.5 ± 1.0 | 2.5 ± 1.0 |
| 38.2 | 115.0 | 104 | 45 | 270 | 2.5 ± 1.0 | 3.5 ± 1.4 |
| 38.0 | 116.0 | 031 | 90 | 180 | 2.5 ± 1.0 | 2.5 ± 1.4 |
| 38.6 | 117.0 | 104 | 45 | 270 | 2.5 ± 1.0 | 3.5 ± 1.4 |
| 38.2 | 119.0 | 041 | 70 | 207 | 2.5 ± 1.0 | 2.8 ± 1.1 |

© 1997 RAS, GJI 130, 551–582