FOCAL DEPTHS AND FAULT PLANE SOLUTIONS OF EARTHQUAKES UNDER THE TIBETAN PLATEAU

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Abstract. We compare synthetic and recorded P waveforms to place constraints on the focal depths and fault plane solutions of 16 crustal earthquakes beneath the highest parts (>4000 m) of the Tibetan plateau. Fault plane solutions for all 16 events show combinations of normal and strike slip faulting with T axes oriented approximately east-west. None of these solutions show thrust faulting. Thus the data corroborate previous inferences that the active tectonics are dominated by east-west extension. Focal depths for all 16 events are less than 15 km and appear to be between 5 and 10 km. This style of deformation and these depths of faulting are similar to those in the Basin and Range province of the western United States. Two intermediate depth events below the crust of southern Tibet also show primarily normal faulting with east-west T axes. The solution for one, discussed by Chen et al. (1981), is unambiguous. The solution for the other, the event of August 1, 1973 (27.59°N, 89.17°E, 85.10 km, m0 = 4.9) is less certain. Both apparently occurred in the mantle beneath a thick, aseismic lower crust, and their occurrence suggests that brittle deformation occurs there in response to a stress field similar to that operating at shallow depths beneath Tibet.

Introduction

With a mean elevation of over 4500 m for an area more than 600 x 1000 km², the Tibetan Plateau dwarfs all other continental plateaus. It follows that an understanding of the evolution of Tibet is essential for an understanding of how such plateaus evolve. Our purpose here is to present seismic data that constrain the depth and style of active tectonics of Tibet [see also Molnar and Tappinconier, 1978; Hj and York, 1978]. Some of these results are remarkably similar to those from the Basin and Range province, and an underlying prejudice of this study is that these two regions are experiencing different stages of a similar tectonic history.

Many recent studies have demonstrated the utility of synthesized seismograms to study either the details of the earthquake sources or the properties of the media through which the waves pass [e.g., Cipar, 1981; Cormier and Richards, 1977; Heimberger, 1968; Kanamori and Stewart, 1975; Rial, 1978]. Most of these studies have considered either subtle details of earth's structure not resolvable with travel times alone or complicated rupture processes of large events. Insofar as is possible, we have studied moderate (m0 = 5.5 to 6.5), simple earthquakes for which the waveforms are shaped primarily by the delays between reflected phases and the orientations of nodal planes [e.g., Jackson and Fitch, 1981; Langston and Heimberger, 1975].

Method

We studied all events between 1962 and 1976 beneath the Tibetan Plateau that were large enough that we could digitize at least one P waveform recorded at an epicentral distance between 30° and 80° by a long-period seismograph of the World-Wide Standardized Seismograph Network (WWSSN). The synthetic seismograms were computed by an algorithm implemented by J. Nabelek following Bouchon [1976]. A short discussion of the procedure was given by Chen et al. [1981]. In all cases we synthesized seismograms using the direct P phase and the surface reflections, PP and SP. In a few cases we experimented with layered structures, but we found that the effects of layering were small compared with other free parameters. Here we present results only for calculations that ignore layering of the earth near both source and receiver. Most events are small and signals are simple, so that we could use a point source with a finite source time function. In all cases we used symmetrical trapezoidal or triangular pulses for far field source time functions. For the three largest events (March 6, 1966; July 14, 1973 [0451]; January 19, 1975), the signals were too complicated to allow a point source with a simple time function, and we present results employing multiple point sources.

We assumed a causal attenuation operator of the form $\exp[-\omega t/2]$, with $t^* \equiv$ travel time/quality factor = 1 s [e.g., Carpenter, 1967; Futterman, 1962]. We did not experiment with different values of $t^*$ because of the obvious trade-off with the assumed values of the pulse width. In all cases we began with fault plane solutions from Molnar and Tappinonier [1978], sometimes more tightly constrained by the addition of seismograms unavailable to them. We then experimented by trial and error with ranges of pulse widths, depths of foci, and orientations of nodal planes. For the simpler events, our goal was not to match the recorded waveforms perfectly but to use both good and bad matches to constrain the ranges of possible source parameters. From the ratios of amplitudes between observed and
synthetic seismograms, we estimated the seismic moments.

For simple sources, given the range of acceptable fault plane solutions allowed by the first motions of P waves, the depth of focus affected the shape of the synthetic waveforms most. Only rarely do the synthetic P waveforms yield the uncertainties in the orientations of the nodal planes to less than 20°-30°. The principal exceptions are for events that predominantly strike slip faulting on steeply dipping planes, for which the P wave first motions already constrain the nodal planes well. For these cases, the sensitivity of the waveforms to the orientations of the planes can be so large as to give the false impression that the uncertainties in their orientations can be reduced to a few degrees. Given the likely heterogeneity in the earth structure, we doubt that such uncertainties can be less than about 5°. With one exception, the event of May 22, 1971, discussed below, we did not find that the synthetic P waveforms allowed a significant reduction in the uncertainties in the fault plane solutions beyond those provided by P wave first motions.

The virtue of comparing synthetic and recorded waveforms for small events is to place tight constraints on depths of foci (±2-4 km). The major source of uncertainty is from a trade-off between the source time function and the depth. The durations and amplitudes of each half cycle of the waveforms depend upon both the duration of the pulse radiated by the source and the intervals of time among the direct and reflected phases [e.g., Jackson and Pritch, 1981; Langston and Helsenger, 1975]. An additional small error in the depth arises from the fact that the fractional error in the assumed velocities of the P and S waves in the source region. We arbitrarily used 6.0 and 3.6 km/s.

For multiple events, the number of parameters that can be assumed multiples rapidly, and the uncertainties in each increase accordingly. We include discussions of three events for which multiple sources are required. Since our emphasis is not on the details of earthquake sources, however, we have not made an exhaustive study of the range of possible rupture histories of them. In all cases we assumed that the fault plane solutions and depths of foci of each subevent were the same. We found that unless the sources were much more complex than we have assumed, the waveforms placed relatively tight constraints on the depths of foci. For a simple, single pulse recorded by a long-period WWSSN instrument, the first half cycle is much larger than the second. A second pulse of opposite sign and within a few seconds of the first is necessary to cause the second half cycle to be bigger than the first. Hence, when the second half cycle is larger than the first, probably P or S to arrives within a few seconds of the direct P phase and has the opposite polarity of it. Because this relationship between the first and second half cycles is consistently observed for the three largest events, which clearly are multiple events (March 6, 1966; July 14, 1973 (0451); January 19, 1975), we are convinced that the first subevent occurred at shallow depths. The factors that affected most the latter parts of synthetic waveforms for multiple events were the relative amplitudes and the lengths of the time intervals among subevents. Here we present only one synthetic seismogram for each recorded trace and simply state our qualitative estimates of the uncertainties in the depths and orientations of the nodal planes. To present enough synthetic seismograms to substantiate these assertions would require an excessively large number of illustrations.

Finally, for each event we calculated the geometric mean of the seismic moments from different stations to obtain the values listed in Table 1. Most are probably uncertain by about a factor of 2. Estimates from station to station are in general more consistent than this factor, but significant trade-offs among pulse widths, depths of foci, and seismic moments introduce additional errors. Moreover, for the largest events, both the uncertainty of how to describe the rupture history and the relatively poor fits of synthetic to recorded waveforms offset the advantage of having many stations and a good azimuthal distribution.

Results

Figures showing the fault plane solutions and synthetic seismograms for the events studied here (Table 1, Figure 1) are presented in Appendix A with discussions of each event in the appropriate caption.

Crustal Earthquakes

All of the 16 crustal events that we studied occurred at shallow depths: less than 15 km and probably between 5 and 10 km. This suggests that the crust below 10-15 km is essentially aseismic (Figure 2), as it is in most of the western United States. The depths obtained here agree with those obtained by Romanovitz [1981] for two events that she also studied with body waves. We found, as she did, that her depths based on surface waves were greater than those from the synthesis of body waves.

The fault plane solutions are slightly better constrained than those given by Molnar and Tappinonier [1978], partly because for the more recent events we were able to analyze seismograms from a few key stations that they could not. In a few cases, the comparison between synthetic and recorded seismograms reduced the uncertainties further. Nearly all solutions show a large component of normal faulting (Figure 1). For two events (May 22, 1971, and May 5, 1975), primarily strike slip motion occurred on steeply dipping planes. In all cases the T axis is more nearly horizontal than the P axis, and the T axes are consistently oriented approximately east-west (±45°). Thus these data do not alter the deduction that the upper crust of Tibet is actively extending in an east-west direction [Molnar and Tappinonier, 1975, 1978; Nl and York, 1978]. Together with the more recent analysis of normal faulting in southern Tibet [Tappinonier et al., 1981], one must conclude that the north-south shortening of the upper crust of Tibet occurs by thrust faulting, it must do so in a manner that leaves little evidence for such a process.
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<th>Longitude</th>
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<th>Strike</th>
<th>Dip</th>
<th>Slip</th>
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<td>180</td>
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As support for this statement, it is important to point out that Molnar and Tapponnier [1978] inferred a thrust solution for one event in Tibet, that of May 22, 1971. We found that recorded seismograms at two stations, the only ones with a sufficiently large signal-to-noise ratio, could not be matched to the synthetic seismograms using a thrust solution (Figure 4A). A strike slip solution allowed a satisfactory match. Reanalysis of several seismograms (e.g., at stations COL and KEV) showed that the abrupt clear upward signals that Molnar had picked as the first motions, in fact, arrived a few seconds after the small but clear arrivals of the P phase recorded by the short-period seismograms. We infer that rays to these stations were radiated near nodal planes. Thus the one thrust solution that Molnar and Tapponnier [1978] had reported is incorrect. The preferred solution includes a large component of strike slip motion and is roughly consistent with the orientation and sense of surface faulting observed by Tapponnier et al. [1981] for the nearby earthquake of November 18, 1951.

Finally, as discussed in Appendix B, average rates of east-west extension in Tibet of approximately 0.3 to 1% per m.y. are plausible for the last few m.y.

Intermediate Depth Earthquakes

Chen et al. [1981] studied one event (September 14, 1976) in southern Tibet that occurred at a depth of 90 km. The fault plane solution shows normal faulting with an east-west trending T axis. Since that study, we found a second event with a comparable depth (August 1, 1973). Clear later phases seen on short-period seismograms attest to a depth of 85±10 km (Figure
of development. In both, the crust probably was thickened by crustal shortening, and now it is thinning by crustal extension. Gravitational potential energy stored in regions of high elevations, and for Tibet in the crustal root as well, would provide the energy to drive the extension and crustal thinning [e.g., Molnar and Tapponnier, 1978]. The greater elevation and thicker crust of Tibet suggests that less extension of it has occurred than of that in the Basin and Range province, which apparently has extended by 30 to 100% of its original width [e.g., Wernicke et al., 1982]. The high Andes of Peru and Bolivia, where extension is still small and localized [e.g., Dalmayrac and Molnar, 1981; Suarez et al., 1983], might be in a more youthful stage than Tibet.

Conclusions

Fault plane solutions of 16 crustal events and two intermediate depth events beneath the highest parts of the Tibetan plateau show large components of normal and strike slip faulting with east-west trending T axes. None of these events exhibit components of thrust faulting.

Focal depths of all of these crustal events are less than 15 km and appear to be between 5 and 10 km. The two intermediate depth events occurred at depths of 85 and 90 (±10) km in the mantle below Tibet’s thick crust. Thus, brittle deformation occurs at shallow depths in the crust and in the uppermost mantle, but the lower crust seems to be aseismic.

Appendix A

Figures A1-A17 show fault plane solutions and comparisons of synthetic and recorded seismograms of the long period vertical component of P waves for events listed in Table 1. See Chen et al. [1981] for a discussion of event 18 (September 14, 1976).

In all cases, equal-area projections of the lower hemisphere of the focal sphere are shown with solid circles as compressional and open circles as dilatational first motions. Smaller symbols indicate less reliable readings, and in the case of the event of August 1, 1973, clear readings are from short-period seismograms. Arrows indicate S wave first motions and crosses show weak signals inferred to be nodal.

Time scales and shapes of far field source time functions (trapezoids or triangles) are shown below most synthetic seismograms.
Fig. A1. March 6, 1966; 31.49°N, 80.5°E, 8±5 km. (a) Fault plane solution. With dilatational first motions at all but two stations, the fault plane solution is not well constrained but clearly requires a large component of normal faulting [Fitch, 1970]. (b) Comparison of recorded (dashed lines) and synthetic (solid lines) P waveforms. The amplitude scales of those at BAG, PMG, and CTA were reduced 0.5 times. From the recorded signals it is clear that a simple time function is inadequate to model the waveforms. The shape of the signal suggests that two subevents occurred a few seconds apart. By trial and error we examined numerous possible pairs of sources before settling upon those shown here. The fault plane solutions and the locations for both subevents are assumed to be the same (Table 1). The seismic moment for the second subevent is assumed to be 2.5 times larger than that of the first, and the interval between subevents is 8 s. We experimented with different depths (between 6 and 15 km), different strikes (from 0 to 50°), different dips (between 40° and 55°), different slip angles (240°, 270°, and 300°), and different time functions (with durations from 4 to 8 s), and found that most values within these ranges yielded similar synthetic signals. Except that there clearly is a large component of normal faulting in an east-west direction, we cannot constrain the fault plane solution very well. The uncertainties in the strike, dip, and slip are at least 30°, 10°, and 30°, respectively. A focal depth of 6 to 10 km yields the best fits. Short (4 s) and long (8 s) time functions do not provide good fits, at least for the ranges of other parameters considered. We used a trapezoid 6 s long (1 s, 4 s, 1 s). Given normal faulting at about 8 km, the most dramatic changes to the observed seismograms were affected by varying the relative sizes of the seismic moments and the interval of time between them. The second subevent must be at least twice as large as the first but not much more than 3 times larger. The uncertainty in the delay between them is about 1 s. We did not try different epicenters or depths or different fault plane solutions for the two events because we do not think that the data will allow resolution of them.
Fig. A2. October 14, 1966; 36.45°N, 87.43°E, 8±1 km. (a) Fault plane solution. First motions constrain the strike and dip of the southeast dipping nodal plane within a few degrees and require a large component of normal faulting, but the strike and dip of the other nodal plane are uncertain by tens of degrees. (b) Comparison of synthetic and recorded P wave forms. We experimented with different pulse durations and different fault plane solutions. With the low signal-to-noise ratio of the recorded waveforms we could not discriminate among pulse widths differing by as much as several seconds, and we could not discriminate among essentially all fault plane solutions allowed by the first motions.

Fig. A3. May 3, 1971; 30.79°N, 84.33°E, 824 km. (a) Fault plane solution. First motions clearly do not constrain the nodal planes tightly but require a large component of normal faulting. (b) Comparison of synthetic and recorded seismograms. We found only one seismogram large enough to digitize. We did not carry out an extensive search among possible time functions (here we use 0.5, 2.6, 0.5) or fault plane solutions because it did not seem to be worthwhile with only one seismogram with a small signal-to-noise ratio.
Fig. A4. May 22, 1971; 32.39°N, 92.12°E, 8±2 km. (a) Fault plane solution. The first motions allow both strike-slip solutions similar to those shown and thrust solutions with P axes trending approximately north-south. Weak signals at stations north of the epicenter and stronger later phases, likely to be sP, are more easily reconciled with the strike-slip solution. (b) Comparison of synthetic and recorded waveforms. With only two stations, the fault plane solution cannot be tightly constrained, but the short separation of the two upward pulses requires a strike-slip and prohibits a thrust solution. The three top synthetic waveforms are for a strike-slip solution. The one at the bottom is for the thrust solution given by Molnar and Tapponnier [1978] and clearly does not fit.

Fig. A5. July 22, 1972; 31.43°N, 91.49°E, 8±4 km. (a) Fault plane solution. The first motions constrain both planes well; if the reading at HKC is correct, the orientations of both are uncertain by only about 10°. Large components of both normal and strike-slip faulting are required. The strikes, dips, and slip angles for solution 1 (solid lines) are in Table 1; for solution 2 (dashed lines) they are 222°, 65°, and 340°; and for solution 3 (solid and dotted lines) they are 212°, 65°, and 270°. We include 3 in order to show that such a solution would be allowed by the waveforms if the first motion at HKC were unreliable. (b) Comparison of synthetic and recorded waveforms. Vertical lines show the arrival times of P phases on short-period seismograms at AAE, ATU, and IST. Most synthetic waveforms use solution 1; asterisk and double asterisk correspond to solutions 2 and 3. The waveforms do not discriminate among the solutions very well except that solution 3 does not yield a waveform at POO that fits as well as those for the other solutions. The uncertainties in propagation paths to POO, which is unusually close to the earthquake (Δ = 20.5°), renders this evidence weaker than if the distance were greater than 30°.
Fig. A6a. July 14, 1973; (0451), 35.18°N, 86.48°E, 6±5 km. (a) Fault plane solution. The P wave first motions constrain the nodal planes better than they do for most of the earthquakes in this study. Allowable strikes and dips lie within ranges of 30° and 10°, respectively, and require large components of normal and strike slip faulting with the T axis plunging gently to the northwest or southeast.  

(b) Comparison of recorded (dashed lines) and synthetic (solid lines) waveforms. P phases are clear but not too large at many stations. Because of complexities in the signals, however, we have not been able to obtain a unique set of parameters that describe the faulting history and that allow a good fit of synthesized to recorded seismograms. The difficulty is most clearly shown by a comparison of signals at AAE or NAI with other stations. To fit the recordings at these stations seems to require two events at shallow depths and with very short durations (symmetric triangular pulses, 1 s in duration here) and about 6 s apart. We found that larger durations (2 s or more) and greater depths (8 or 10 km) led to poor fits at these two stations. At the same time, such short pulses could not be used to synthesize the recordings at other stations. The recordings at most stations clearly require the superposition of at least two simple pulses, but the durations must be much greater than 1 s. For stations to the west and north, two trapezoidal pulses each with a duration of 5 s (1, 3, 1) and separated by varying intervals could allow an adequate match to the recordings. Here we show synthetic waveforms at the following stations (with intervals between subevents): EIL (4.5 s), ATU (4.2 s), IST (4.0 s), TRI (4.2 s), STU (4.2 s), KON (4.0 s), NUR (3.0 s), KEV (3.5 s), KBS (3.5 s), and GGD (3.5 s). At COL, the initial half cycle requires a long time function, but to fit the step in the second half cycle the durations of the pulses must be less than those for stations in Europe. Here we use 3 s (1, 1, 1), with an interval of 3 s between the separate pulses. At stations to the east, once again pulses with a long duration (5 s) but with a short interval between them seem to be necessary, which is equivalent to one pulse of very long duration (6-8 s). Here we used symmetric trapezoidal pulses 5 s in duration (1, 3, 1) with different delays: ADE (3.5 s), PMG (3.5 s), BAG (3.5 s), and GUA (3.0 s). The general decrease in the assumed intervals between the two events, from 6 s toward the southwest (NAI and AAE) to 3 s toward the north (COL) and east (GUA) suggests that a second event occurred northeast of the first. The large variations in the lengths of time functions used here, however, suggest that such a simple description of the source is inadequate. We conclude that there must be at least two subevents, but without a great deal more work, we cannot place tight constraints on their relative positions.
Fig. A7. July 14, 1973; (13 39 hours), 35.26°N, 86.60°E, 7±3 km. (a) Fault plane solution. First motions constrain the dip and strike of the southeast dipping nodal plane with uncertainties of 5° and 15°, respectively. The orientation of the other nodal plane is more uncertain, but clearly a large component of normal faulting is required. Parameters for the planes with solid lines are in Table 1, and those with dashed lines correspond to a solution: strike = 052°, dip = 70°, slip = 304°. (b) Comparison of synthetic and recorded seismograms. The very short pulses at 14 stations (those shown plus ATE, IST, JER, KON, KBS, NAI, NUR, STU, and TRI) require a short time function and a shallow depth. The synthetic waveforms are not very sensitive to the orientations of the nodal planes; the dashed planes in Figure A7a were used to calculate the top waveforms, designated by an asterisk, and the solid lines in Figure A7a show the planes used for the other synthetic waveforms. We also tried different orientations of the west dipping plane, corresponding to a smaller strike slip component, but the synthetic waveforms are not very sensitive to the orientation of this plane.

Fig. A8. August 1, 1973; 29.59°N, 89.17°E, 85±10 km. (a) Fault plane solution. The small magnitude (m = 4.9) of this event does not permit a well-constrained solution to be obtained. Large symbols show two first motions from long-period seismograms. The remaining readings are from short-period instruments. In particular, readings from ADE and MUN are not certain, and a thrust solution is possible. One reason for preferring the normal fault solution, despite the inconsistency of ADE, is the observation of a clear S wave at JER, with an eastward initial motion and without a clear north-south component. (On the focal sphere, the first motion points west-northwest.) For a thrust solution, the initial motion at JER would be westward, and on the focal sphere it would point toward the east or northeast. While we prefer a large component of normal faulting, we recognize that these data cannot prove such a solution, as they do for the event of September 14, 1976. (b) Examples of short-period seismograms. Seismograms from CCL, EIL, and UME show emergent P waves but clear later phases, probably pF. That from MAT shows a clear first motion for P. For intervals between pP and P of 24±0.5 s at KOD, 25±1 s at EIL, 24.5±0.5 s at NUR, 25±1 s at UME, 26±2 s at KTG, and 26.5±0.5 s at CCL and assuming a crustal thickness of 70 km, we infer a depth of 85±10 km. The plots of the P wave are in error by ±0.2 s.
Fig. A9. August 16, 1973; 33.24°N, 86.84°E, 8±4 km. (a) Fault plane solution. First motions require a substantial component of strike slip faulting and probably some component of normal faulting, but the dips of the nodal planes are uncertain by 10° or more and the strikes by more than 20°. Other possible solutions are tested in Figure A9b using synthetic and recorded seismograms at JER. (b) Comparison of synthetic and recorded seismograms. This event is small, and the signal-to-noise ratio (even at the stations considered here) is small. The symmetric triangular pulse with a duration of 2 s seems to fit the observed waveforms better than longer pulses, but little effort was made to constrain the waveforms. The parameters for the strike, dip, and slip are given in Table 1 for 2 (solid lines) and 3 (dot-dashed lines), and are for 1 (dashed lines) 150°, 55°, and 195°; for 3 (dot-dashed lines), 180°, 55°, and 155°; and for 4 (dotted lines), 180°, 60°, and 216°.

Fig. A10. September 8, 1973; 33.29°N, 86.82°E, 9±4 km. (a) Fault plane solution. First motions constrain the orientations of both nodal planes with uncertainties of only about 10° and require large components of both normal and strike slip faulting. Solid lines show the nodal planes used for most of the synthetic waveforms. Synthetic seismograms were calculated for the station MAT using three different east-dipping planes labeled 1 (slip angle = 195°), 2 (slip angle = 199°), and 3 (slip angle = 203°) to show the effect of a change in the orientation of this plane. (b) Comparison of synthetic and recorded seismograms. We experimented a little with the length of the source time function and found that a short pulse (2 s) gave synthetic waveforms that were too sharp. Comparisons of synthetic and recorded seismograms at eight of nine stations were good, but the depth is probably less than 15 km and could be as shallow as 5 km (9±4 km). Possibly because of the low signal-to-noise ratios, or for otherwise inexplicable reasons, the fit of BAG is poor for all depths. On the right, synthetic waveforms of MAT are shown for three solutions shown in Figure A10a, that differ only slightly from one another. Clearly, the ray to MAT must make an angle of only a couple of degrees with one nodal plane.
Fig. All a. January 19, 1975; 32.39°N, 78.50°E, 9±4 km. (a) Fault plane solution. With only one compressional first motion, the first motions do not constrain the nodal planes tightly, but they do require a large component of normal faulting. (b) Comparison of synthesized and recorded P waveforms. The seismic moment for this event is the largest of those studied here, and correspondingly, the signals are very large. For presentation, the ratio of the amplitude to time scale is reduced 0.3 times. The signals are very complicated and clearly require a multiple source. Matching the waveforms was a tedious and expensive process. Had we been concerned with details of the faulting process, we would have tried harder to constrain the parameters for individual parameters more tightly. Here we present synthetic seismograms for sources consisting of four subevents, all with the same depth (9 km), with the same fault plane solution (pure normal faulting on a plane striking north-south and dipping either 40° east or 50° west), with the same symmetrical trapezoidal time function 3.5 s in duration (1, 1.5, 1), and with relative amplitudes of 1.0, 0.8, 0.8, and 0.5. We then adjusted the time intervals between subevents to match the recorded waveforms to some level of arbitrary but low level of satisfaction. For stations to the southeast, we used the delays between events that were greater than those for stations to the east. This suggests that later events occurred west or northwest of the initial event, but because the fits are not very good and could be improved by the addition of yet more subevents, we do not consider this result to be required. We also experimented with fault plane solutions and found little sensitivity to the orientations of planes (±40°). We found that depths as deep as 15 km or as shallow as 3 km could be eliminated, but otherwise the depth is poorly constrained. The intervals between the first and later subevents used here are (in seconds): COL (3.6, 7.5, 11.5), GDH (3.0, 7.5, 11.3), NUR (2.7, 7.7, 11.5), STU (2.6, 7.0, 10.8), IST (2.5, 6.5, 10.5), ATU (2.2, 6.0, 10.0), NAI (3.5, 7.5, 10.5), ADE (4.5, 8.5, 12.5), PMG (4.0, 8.0, 12.5), and RAB (5.3, 9.0, 13.0).
Fig. A12. April 28, 1975; 35.82°N, 79.92°E, 7±3 km. (a) Fault plane solution. The first motions tightly constrain the orientations of the nodal planes, so that neither is likely to be in error by more than 10°. Large components of both normal and strike slip faulting are required. (b) Comparison of synthetic and recorded seismograms. We experimented with small changes in the orientations of the nodal planes in order to obtain satisfactory matches between synthetic and recorded waveforms, but we did not try to reduce the uncertainties in the orientations of the planes. The short characteristic periods of the recorded phases require a relatively short source time function.

Fig. A13. May 5, 1975; 33.09°N, 92.92°E, 7±2 km. (a) Fault plane solution. The first motions tightly constrain the orientations of the two nodal planes; the orientation of neither is likely to be in error by more than 10°. (b) Comparison of synthesized and recorded waveforms. Having varied the orientations of the nodal planes within the limits imposed by the first motions, we found that a unique source time function did not allow a fit of the synthetic waveforms with the recordings at all stations. For most stations, including three not shown here (IST, JER, and NAI), a short triangular pulse allows a good fit to the recordings for a depth of about 7 km. Depths even as large as 10 km seem to be excessive. For the recording of BAG, a longer pulse appears to be required. The broader pulse at BAG, to the east of the event, may suggest a westward rupture propagation along the east-northeast trending fault, but the limited amount of data is not adequate to prove this.
Fig. A14. May 19, 1975; 35.16°N, 80.80°E, 8 ± 4 km. (a) Fault plane solution. The first motions tightly constrain the northwest dipping nodal plane but place only a weak constraint on the other plane. Clearly, a large component of normal faulting is required, and a large strike slip component is possible. (b) Comparison of synthetic and recorded P phases at three stations. Clearly, the event is small, and few stations recorded signals large enough to be synthesized. The relatively low frequencies prohibit a short time function. Keeping the same northwest dipping plane, we calculated synthetic seismograms at these stations and found them to be virtually identical, regardless of whether the other nodal plane dipped more to the north than is shown or dipped southeast, corresponding to pure dip-slip motion.

Fig. A15. June 4, 1975; 35.87°N, 79.85°E, 9 ± 4 km. (a) Fault plane solution. The dip of the plane that dips west is well constrained, but its strike and the orientation of the other plane are constrained mostly by the compressional first motion at SHL. With that station, the dips and strikes of the planes are uncertain by only about 5° and 10°, respectively. We computed synthetic seismograms for other fault plane solutions to examine whether they could add an additional constraint if that reading at SHL were in error. The southeast dipping plane could be rotated to dip east but probably not northeast. The match of synthetic to observed phases is not sensitive to the orientation of this plane. (b) Synthetic and recorded P phase for five stations, four north and west of the event and one at GUA, to the east of it. The relatively smooth recorded phases suggest that the time function is relatively long. We experimented with a number of time functions and found that none could be used to fit P phases at all 11 stations for which we digitized seismograms. Here we use a symmetric triangular pulse with a duration of 4 s, which also works well at AQU, GDR, KEV, and KQN. (c) Synthetic and recorded seismograms at BAG and MAT, stations east of the event. Longer time functions than those used in Figure A15b were needed to obtain an adequate match between synthetic and recorded seismograms. Here we used a symmetric trapezoidal time function with a duration of 7 s. The longer durations at these stations than at stations east of the event may indicate a westward rupture propagation along the plane that strikes more nearly east-west, but this limited amount of data is not enough to be convincing.
Fig. A16. July 19, 1975; 31.92°N, 78.61°E, 6±3 km. (a) Fault plane solution. The compressional first motion at SHL constrains the solution to include large components of both normal and strike slip faulting. Only if the readings there or at NDI were in error could the dips and strikes of the nodal planes be altered by more than 10° and 20°, respectively. (b) Synthetic and recorded seismograms. The short characteristic periods of the observed signal implies a short time function. The signals are too small to be used as an additional constraint on the fault plane solution, but they do require a shallow source.

Fig. A17. July 29, 1975; 32.56°N, 78.46°E, 8±3 km. (a) Fault plane solution. First motions of P waves do not constrain the nodal planes tightly. They require normal faulting on a plane that must dip 30° to 65°. The strikes of the nodal planes, however, are uncertain by tens of degrees, and a large strike slip component is possible. (b) Synthetic and recorded seismograms for P phases at four stations. We found a slight sensitivity to the orientations of the nodal planes but not enough to constrain the orientations convincingly. As usual there is a trade-off between the duration of the source time function and the depth assumed in calculating the synthetic seismograms. For symmetric trapezoidal time functions of two different durations, a depth of 8±3 km appears to be appropriate. For most of the synthetic waveforms we used pulses with a duration of 3.5 s, but for those identified by an asterisk, the duration is 2.0 s.
Appendix B

Rough Estimates of Rates of Deformation in Tibet

Using the fault plane solutions and seismic moments in Table 1 for the shallow events, we can calculate the corresponding moment tensor for each. Kostrov [1974] showed that by summing the moment tensors \( M_{ij} \) one can obtain an estimate of the strain tensor \( \epsilon_{ij} \) caused by slip during earthquakes [e.g., Chen and Molnar, 1977]:

\[
\epsilon_{ij} = M_{ij} / 2 \mu V
\]

(1)

where \( \mu \) is the shear modulus (= 3.3 x 10^11 dyne/cm²) and \( V \) is the volume of the region. Let us assume, [that a layer 10 km thick and an area of 6 x 10^23 m² are involved in this deformation. If we divide \( \epsilon_{ij} \) by the duration of time considered (15 years), we obtain an estimate of the average strain rate tensor."

With the 1 direction pointing east, the 2 direction pointing north, and the 3 direction pointing up, we obtain for the cumulative seismic moment tensor:

\[
M_{ij} = ( -8.99, 0.41, 3.84 ) \times 10^{25} \text{ dyne cm}
\]

The large values of \( M_{ij} \) correspond to large values of extension in the east-west direction and contraction in the vertical direction. The intermediate (negative) value of \( M_{33} \) corresponds to a substantial contribution of left-lateral, east-west shear or right-lateral, north-south shear. Such a style of deformation is characteristic of the surrounding regions [Tappanier and Molnar, 1977].

The elements of this cumulative moment tensor, however, are very uncertain. They are dominated by the contributions of the three largest events (March 8, 1966; July 14, 1973; and January 19, 1975), for which both the fault plane solutions and the scalar values of the seismic moment are the most uncertain. If we consider the cumulative moment tensor from the remaining 13 shallow events, we obtain:

\[
M_{ij} = ( -1.90, -2.02, 0.50 ) \times 10^{25} \text{ dyne cm}
\]

For this case, the amounts of contraction in the north-south and vertical directions are comparable. Thus, if this tensor were proportional to the rate of change of the strain field for Tibet, the east-west extension of Tibet would be balanced by approximately equal amounts of north-south shortening and crustal thinning as well as left (or right) lateral, east-west (or north-south) shear.

If we use all 16 shallow events to estimate the average rate of east-west extension, then with \( u = 3.3 \times 10^{11} \text{ dyne/cm²} \), \( V = 6 \times 10^{23} \text{ km}^3 \), and \( t = 15 \text{ years} \), from (1) and (2) we obtain \( \epsilon_{11} = 4.4 \times 10^{-12} \text{ yr}^{-2} \), or 0.442 m/y. For a region 1000 km wide in an east-west direction, this corresponds to 4.4 mm/yr of easterly relative motion of eastern Tibet with respect to western Tibet.

The values obtained here are only of qualitative interest for we cannot assign a sensible quantitative value to their uncertainties. We include discussion of these estimates of seismic slip in order to compare them with the estimates implied by different directions of underthrusting beneath the Himalaya [Molnar and Chen, 1982; J. Armbruster et al., manuscript in preparation, 1983]. The directions of underthrusting in the eastern and western Himalaya differ by 10°. Therefore, for rates of underthrusting of 10 or 20 mm/yr, eastern and western Tibet move apart at 5 to 10 mm/yr. These portions of the Himalayan belt are about 1000 km apart, so we estimate rates of extension of 0.5 to 1.0 m.y. values comparable to that obtained above from the seismic moments of earthquakes between 1962 and 1976.

Notice that at rates of 0.3 to 1.0 m.y., if all of the east-west extension were accommodated by crustal thinning, then after 5 to 10 m.y., a crust originally 70 km thick would thin by 1 to 7 km. Clearly, the present crustal thickness of about 65 to 70 km [e.g., Chen and Molnar, 1981; Romanovitz, 1982] would allow such a range of values of crustal thinning. Thus if east-west extension of Tibet has been occurring for 10 m.y. or less over most of the plateau, it is not likely to have thinned the crust much.

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