Abstract. We combine observations of group and phase velocity dispersion of Rayleigh waves, of the waveform of a long-period P phase, of $P_n$ and $S_n$ velocities from unrefracted refraction profiles using earthquakes, and of teleseismic S-P travel time residuals to place bounds on the seismic wave velocity structure of the crust and upper mantle under Tibet. From surface wave measurements alone, the Tibetan crustal thickness can be from 55 to 85 km, with corresponding uppermost mantle shear wave velocities of about 4.4 to 4.9 km/s, respectively. The $P_n$ and $S_n$ velocities were determined to be 8.12\(\pm\)0.06 and 4.8\(\pm\)0.1 km/s, respectively, using travel time data at Lhasa from earthquakes in and on the margins of Tibet. Combining these results, the crustal thickness is most likely to be between 65-80 km with an average shear wave velocity in the upper crust less than 3.5 km/s. A synthesis of one $P_n$ waveform does not provide an additional constraint on the velocity structure but is compatible with the range of models given above. In contrast to observations obtained for eight earthquakes in the Himalaya, measurements of both teleseismic S and P wave arrival times for nine earthquakes within Tibet show unusually large intervals between P and S compared with the Jeffreys-Bullen Tables. Thus the $P_n$ and $S_n$ velocities apparently do not reflect high velocities in the mantle to a great depth beneath Tibet. From the dependence of the seismic velocities of olivine on pressure and temperature and from the similarity of the measured $P_n$ and $S_n$ velocities beneath Tibet and beneath shields and platforms, the velocities at the Moho beneath Tibet are compatible with the temperature being 250°-300° higher than beneath shields and platforms, i.e., 750°C if the temperature beneath the platforms is close to 500°C. Such a high temperature could reach or exceed the solvus of the lower crust. Simple one-dimensional heat conduction calculations suggest that the volcanic activity could be explained by the recovery of the geotherm maintained by a mantle heat flux of about 0.9 HPF at the base of the crust. If the distribution of radioactive heat production elements were not concentrated at the top of the crust, radioactive heating could also contribute significantly to the recovery of the geotherm and thus lower the required mantle heat flow. Thus the idea of a thickened crust in response to horizontal shortening is compatible both with these data and with these calculations.

Introduction

With an average elevation of about 5 km above sea level over an area of 7 x 10^5 km^2, the Tibetan Plateau is one of the most conspicuous topographic features on earth. To the south and southwest, it is bounded by the Himalayan convergent zone, where the continental collision between India and Eurasia plates manifests itself predominantly by low-angle thrust faulting [e.g., Fitch, 1970; Molnar et al., 1973, 1977]. The Altny Tagh and the Kun Lun left-lateral strike slip faults form its northern boundary [e.g., Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1977]. In contrast, recent normal faulting [e.g., Molnar and Tapponnier, 1975, 1978; Nl and York, 1978] and volcanism are widespread over the plateau [Burke et al., 1974]. With its complex variations of active tectonic styles and its massive volume, the evolution of the Tibetan Plateau has been recognized as a key to the understanding of continent-continent convergent processes [e.g., Dewey and Burke, 1973; Molnar and Tapponnier, 1975, 1978; Toksöz and Bird, 1977].

Primarily because of its inaccessibility, geological data available about Tibet are very limited. Upper Cretaceous limestones seem to cover much of southern Tibet [Hennig, 1915; Norin, 1946]. This observation places a lower bound on the time of uplift, but we are not aware of evidence that constrains the timing more tightly. Volcanism, which is apparently quite young is widespread on the plateau [e.g., Burke et al., 1974; Kidd, 1975]. Recently, Deng [1978] reported Quaternary calc-alkaline to alkaline volcanics in northern Tibet at sites close to those of the earlier findings of Norin [1946]. Similar rock types have also been reported along the southern part of Tibet [Chang and Cheng, 1973; Hennig, 1915; Kidd, 1973]. Gravity data are also very limited. Ambold [1948] reported one gravity measurement within the plateau that is consistent with isostatic equilibrium. Chang and Cheng [1973] also inferred isostatic equilibrium from more recent but unpublished gravity data.

Given the unusual elevation and the limited geologic data, the crustal and upper mantle structures should place important constraints on the geologic evolution of the plateau. Almost all the previous seismic studies of Tibet are based on group velocity dispersion curves of surface waves [Bird, 1976; Bird and Toksoz, 1977; Chun and Yoshi, 1977; Patton, 1978; Tung and Teng, 1974; Tseng and Sung, 1963]. In general, these investigators concluded that the crust is about 70 km thick with a relatively low upper mantle shear wave velocity (near 4.5 km/s). The mantle velocities, however, cannot be well constrained with surface wave data alone [Der et al., 1970]. The possible trade-off between upper mantle velocities and crustal thickness was not investigated in most of these studies, and little attention has been given to the uncertainties and bounds of allowable velocity structures. We do
not consider the results in the published studies to be conclusive evidence for crustal thicknesses of 70 km, and consequently, we have carried out a study of not only surface wave dispersion but also of other seismic phases.

Considering that the crust is thick now but probably was not thick before the collision between India and Eurasia in the early Tertiary, there have been two extreme mechanisms proposed for the thickening of the crust and the uplift of the plateau. Dewey and Burke (1973), Toksöz and Bird (1977), and probably others suggested that Tibet formed by relatively uniform crustal shortening and thickening in response to horizontal compression. Others [e.g., Argand, 1924; Powell and Conaghan, 1973, 1975] inferred that low-angle underthrusting of one crustal block (India) beneath another (Tibet) is the dominant mechanism for creating the high plateau. Both models require a decrease, if only temporary, in the crustal temperature gradient due to the crustal thickening. Therefore a more quantitative treatment of either situation is required to explain how volcanism could occur on the plateau.

The existence of the volcanoes and the inference of relatively low seismic wave velocities led logically to the suggestion that the upper mantle beneath Tibet is relatively hot and light. Part of the high elevation but not the crustal thickening could then be a consequence of thermal expansion of the underlying mantle. Although these various opinions diverge considerably and the specific physical processes are vaguely described at best, the ultimate cause of Tibet's elevation, thick crust, and volcanism is virtually unanimously attributed to the India-Eurasia collision.

The existing data for Tibet are too scarce to discriminate among the various proposed evolutionary mechanisms, but tectonic models of the evolution of the India-Eurasia continental collision must provide satisfactory explanations for the structure of the plateau. Furthermore, studying the geophysical characteristics of the Tibetan plateau and the active Himalaya probably provides basic information that constrain the processes of large-scale continental tectonics in general.

The purpose of this study is to present a comprehensive seismological data set that places improved constraints on the seismic velocity structure of the crust and upper mantle beneath the Tibetan plateau. We discuss separately the four relatively independent seismic methods employed: surface wave dispersion, synthesis of a $P_v$ phase, $P_n$ and $S_n$ refraction profiles, and teleseismic $S$ wave travel time residuals from earthquakes in and near Tibet. Constraints on possible models and discussions of them are also presented. At the end, we present an interpretation of all the observations and discuss their tectonic implications.

None of the studies of individual seismic phases requires a unique velocity structure, but a consistent model of the crustal thickness and uppermost mantle velocities can be obtained from the combined observations. In particular, the nonuniqueness of the velocity structure has been reduced considerably by the determination of the uppermost mantle velocities from refraction profiles.

Surface Wave Dispersion

Dispersion Curves

We analyzed long-period seismograms for nine earthquakes in and near Tibet recorded at one or more of the four stations of the World Wide Standardized Seismograph Network (WWSSN) just south of the Himalaya (Table 1). More than 80% of each of the great circle paths is within the plateau. We chose not to use longer paths that cross much of Asia because of the difficulties in deciding how to regionalize them and because of the likelihood of marked lateral refraction at the boundaries of the Tibetan plateau and elsewhere. We trust more the results obtained from paths confined as much as possible to Tibet. Thus, the range of epicentral distances is limited to about 20° (Figure 1). The multfiltering technique [Dziwonski et al., 1969] was used to obtain the group velocity dispersion curves. We analyzed only seismograms that have high signal-to-noise ratios and for which a minimum amount of multipathing was apparent (Figure 2). For only two paths were

<table>
<thead>
<tr>
<th>Date</th>
<th>Number</th>
<th>Epicenter Location Lat. (°N)</th>
<th>Long. (°E)</th>
<th>Origin Time, UT</th>
<th>Stations</th>
<th>Seismic Phases</th>
</tr>
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<tbody>
<tr>
<td>April 19, 1963</td>
<td>I</td>
<td>35.7</td>
<td>96.9</td>
<td>0735:22.7</td>
<td>NDI</td>
<td>LR, P_L</td>
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<tr>
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<td>II</td>
<td>38.6</td>
<td>73.4</td>
<td>1540:00.8</td>
<td>SHL</td>
<td>LR</td>
</tr>
<tr>
<td>March 16, 1964</td>
<td>III</td>
<td>36.95</td>
<td>95.50</td>
<td>0105:19.8</td>
<td>NDI</td>
<td>LR, LG</td>
</tr>
<tr>
<td>Oct. 14, 1966</td>
<td>IV</td>
<td>36.45</td>
<td>87.43</td>
<td>0104:42.9</td>
<td>NDI</td>
<td>LR, LG</td>
</tr>
<tr>
<td>Aug. 30, 1967</td>
<td>V</td>
<td>31.57</td>
<td>100.31</td>
<td>1108:50.0</td>
<td>LAH</td>
<td>LG</td>
</tr>
<tr>
<td>March 24, 1971</td>
<td>VI</td>
<td>35.46</td>
<td>98.04</td>
<td>1354:21.0</td>
<td>NDL</td>
<td>LR</td>
</tr>
<tr>
<td>Aug. 30, 1972</td>
<td>VII</td>
<td>36.65</td>
<td>96.35</td>
<td>1514:07.5</td>
<td>NDL</td>
<td>LR, LG</td>
</tr>
<tr>
<td>Aug. 30, 1972</td>
<td>VIII</td>
<td>36.36</td>
<td>96.35</td>
<td>1847:40.3</td>
<td>NDL</td>
<td>LR, LG</td>
</tr>
<tr>
<td>Feb. 7, 1973</td>
<td>IX</td>
<td>31.50</td>
<td>100.33</td>
<td>1606:25.8</td>
<td>NDI</td>
<td>LG</td>
</tr>
</tbody>
</table>

Hypocenters and origin times as reported by ISC after 1964, by the Coastal and Geodetic Survey before 1964.
Fig. 1. Map of Tibet showing earthquake (dots) - station (triangles) combinations used for surface wave dispersion and the $P_L$ phase observation. Events I and II generated large long-period Rayleigh waves. Note the difference in the paths to stations from these two events. Event I was used for the $P_L$ phase observation also. Major strike slip and thrust faults (see legend) were also plotted following Molnar and Tapponnier [1975].

A combination of a mislocation in epicenter of 10 km and a 2-s error in origin time can cause an error of 0.8% in determinations of both group and phase velocities at a distance of 20°. An uncertainty in the group arrival time or phase delay of one eighth of the period at periods near 80 s leads to an uncertainty of 1.6% in group or phase velocity at this distance. We assume a total combined maximum error of up to 3% in both is possible for the long-period end of the dispersion curves. The uncertainty in the group velocity at short periods should be less. However, lateral heterogeneity will have a much larger effect on relatively shorter-wavelength components than on the longer-wavelength components. This might be part of the reason why there is a fairly large scatter in the data at short periods.

For shallow events, the source depth is not important at long periods. Because both events I and II have nearly vertical strike slip fault plane solutions, obtained primarily from $P$ wave first motion polarities [Molnar et al., 1973; Tapponnier and Molnar, 1977], the uncertainty in the calculated phase delay is small. At extreme conditions, assuming an error of 0.2 cycles in various phase delays, an error of 3.4% in phase velocity is possible if all the errors conspire together for periods near 80 s and a distance of 20°. The relatively large error in the phase delay assumed here takes into account the effect of a possible nonsteplike source time function.

Note that the two Rayleigh wave seismograms with large signal-to-noise ratios at long periods (I-NDI and II-SHL) involve quite different paths (Figure 1). The difference between the two observations at periods longer than 60 s could be simply due to observational errors, but it might actually reflect real differences in average velocity structures for different paths.
Velocity Models

Before interpreting the data, we urge the readers who are not familiar with the resolving power of surface wave dispersion on the velocity structure to examine the work by Der et al. [1970]. With group velocity observations of fundamental Rayleigh and Love waves only for a limited frequency band and with fairly large scatter (Figure 3) and with phase velocity measurements of fundamental Rayleigh mode between 60 to 90 s only (Figure 4), we did not perform a formal inversion. Instead, we examined a range of plausible models, and then from a comparison with the observation, we assigned bounds to the range of possible structures.

In constructing these models, we made the following assumptions:

1. The velocity structure below 90 km for most models and density structure for all models are the same as Gutenberg's continental model [Dorman et al., 1960].
2. The ratio of P wave velocity to S wave velocity is fixed at 1.73 in the crust and 1.80 in the uppermost mantle.
3. The shear wave velocity of the top 3.75 km is fixed at 2.55 km/s [Chen and Molnar, 1975]. We did not consider the period range 5 to 10 s here.
4. No more than three layers in the crust, excluding the top sedimentary layer, are included, and the mantle between the Moho and a depth of 90 km is assumed to be homogeneous.

Since the resolution of P wave velocities is very poor compared with the resolution of S wave velocities, we examined dependencies of the dispersion curves only on $V_p$ in the mantle, the crustal thickness, and the shear wave velocity structure in the crust. Theoretical dispersion curves were compared primarily with Rayleigh wave
Fig. 3a. Group velocity dispersion curves of Rayleigh waves. No clear signal was observed between 30- to 60-s periods. A 3% maximum error in group velocities would be about 0.1 km/s.

Fig. 3b. Group velocity dispersion curves of Love waves.
Fig. 4. Phase velocity dispersion curves for Rayleigh waves. Observed values for I-NDI path (crosses) and II-SHL path (dots) were plotted together with three theoretical curves calculated for models F4, S7, and S4. Model SE generated nearly identical phase velocity dispersion curve as that of model S7 at the period range shown. Error bar shown is approximately ±1.4% maximum error in phase velocity. See Figures 5 and 6 for model parameters.

group velocity observations, but for selected structures, comparison was also made with Rayleigh wave phase velocity observations. Love wave group velocity observations served only as a loose constraint. As long as theoretical values fall within the scatter of these data, no further comparison with them was made. Surface wave dispersion at the short period range (<30 s) does not constrain the crustal thickness or the upper mantle shear wave velocity much. Little attention will be paid to these short-period range observations in the following discussion of these two subjects.

Crustal thickness and upper mantle S wave velocities. Rayleigh wave dispersion curves for three different crustal thicknesses (55, 70, and 85 km) were calculated for a range of uppermost mantle shear wave velocities ($\beta_m$) from 4.4 to 4.9 km/s (Figure 5). At long periods, higher values of $\beta_m$ cause higher Rayleigh wave group velocities, and greater crustal thicknesses cause lower group velocities (Figure 5). For a crustal thickness of 55 km, only the model with the lowest value of $\beta_m$ (4.4 km/s, model F4) fits the upper curve of the Rayleigh wave group velocity observations (Figure 5a). For a 70 km thick crust (Figure 5b), however, the $\beta_m$ seems to be constrained to be near 4.7 km/s (Model S7) or higher. Values of $\beta_m$ as small as 4.4 km/s seem to be incompatible with a 70 km crust (Figure 5b). If we further increase the crustal thickness to 85 km (Figure 5c), the observations constrain $\beta_m$ to be 4.8 or even 4.9 km/s (Figure 5c), but the fit between the observed and theoretical curves appears to us to be worse than for the previous cases of 55 and 70 km thick crust. These results confirm Der et al.'s [1970] numerical experiments that show that surface wave dispersion alone is inadequate to constrain both the crustal thickness and the upper mantle shear wave velocity tightly.

Phase velocity dispersion curves for four velocity models are plotted with the two observed dispersion curves in Figure 4. The curve for model F4 (55-km-thick crust, $\beta_m = 4.4$ km/s) lies within the scatter of the observations. Both model S7 (70-km-thick crust, $\beta_m = 4.7$) and model S8 (70-km-thick crust, $\beta_m = 4.8$, with the lowermost crustal velocity lower than that of model S7) fit the observed phase velocities for the I-NDI path well and were within the uncertainties of the observations for the II-SHL path. Model S4 (70-km-thick crust, $\beta_m = 4.4$ km/s) does not fit the II-SHL observed phase velocities and is systematically lower than the data for I-NDI. It turns out that if $\beta_m > 4.8$ km/s, however, the phase velocity data cannot rule out a crustal thickness of 85 km.

Velocity structure of the crust. The crustal structure is controlled primarily by the shorter-period portion of the dispersion curves. This portion of the dispersion curve depends primarily on the mean shear wave velocity in the crust and in its gradient. Figure 6 shows the theoretical Rayleigh wave group velocity dispersion curves for five different models with a fixed crustal thickness of 70 km and $\beta_m = 4.7$ km/s. If the acceptable values of the calculated Rayleigh wave group velocities are those that lie within the scatter of our data, then for such a crustal thickness, an average crustal shear wave velocity ($\bar{\beta}$) of 3.5 km/s seems to be too high (Figure 6a). It is found that a single layer between the top low-velocity surface layer and the Moho will not fit the data unless $\bar{\beta}$ is much less than 3.5 km/s. By adding a higher-velocity layer at the base of the crust so as to cause a velocity gradient in the crust and maintaining a constant $\bar{\beta}$, dispersion curves can be constructed that agree well with the data (Figure 6b). To obtain such a fit, however, requires a relatively low-velocity upper crust so that $\bar{\beta}$ is still less than 3.5 km/s.

The long-period end of the dispersion curve, however, is sensitive to the shear wave velocity in the lower crust. For instance, the addition of a relatively high-velocity layer (such as $V_s = 3.9$ km/s) between 60- and 70-km depth has the same effect on the long-period portion of the dispersion curve that an increase of $\beta_m$ (such as from 4.7 to 4.8 km/s) would have (Figure 6b). These variations are, of course, only a few examples of many other possibilities. Thus there are trade-offs between the lower crustal shear wave velocity and $\beta_m$ which is in turn coupled with the crustal thickness.

For a crustal thickness of 85 km, the constraints on the crustal structure placed by surface waves will be even looser than the previous discussions for 70-km-thick crust cases. We did not investigate any detailed models for this crustal thickness. Velocity models with a crustal thickness of 55 km, $\bar{\beta} < 3.5$ km/s, and $\beta_m < 4.4$ km/s could also generate group velocity dispersion curves which are consistent with observations. These models are not discussed here for reasons that will become apparent later in this paper.

The most important constraint that surface wave dispersion places on the crustal structure is that the average crustal shear wave velocity is low. Clearly, for the upper 40-50 km, $\bar{\beta} < 3.5$ km/s. Since the data seem to be better fit by a series of layers of increasing velocity than by a single layer of uniform velocity in the crust below sediments, the lower crust could have a shear wave velocity greater than 3.5 km/s, but it appears that even with this increase, $\bar{\beta}$ is less
Fig. 5a. Theoretical Rayleigh wave group velocity dispersion curves (top) for three models (bottom) with a crustal thickness of 55 km for different uppermost mantle shear wave velocities ($\beta_m$). The two long-period observations are shown as solid triangles for comparison. Also shown is the theoretical Love wave dispersion curve for model F4.

than 3.5 km/s for the entire crust. The crustal thickness and uppermost mantle shear wave velocity combinations obtained earlier are not very sensitive to the possible variations in the crustal structure.

Synthesis of Rayleigh Wave Signals

The group and phase velocity dispersion curves contain very little information about the amplitude of the surface waves other than the fact that there must exist a substantial amount of Fourier amplitude (spectral density) at a particular frequency in order to obtain a reliable determination of the group arrival time. As a further check on the velocity structure, we have used the summation of normal modes to generate synthetic seismograms of vertical components of Rayleigh waves to investigate the possibility of detecting structural differences from these synthetic seismograms. The generation of synthetic seismograms requires a knowledge of the source depth, the fault plane solution, the source time function, the seismic moment, and parameters describing attenuation and a velocity structure. Usually, one assumes a knowledge of several of the factors and tries to deduce the others from the agreement of the synthetic
seismogram with the observed one [Kanamori, 1970].

We summed fundamental Rayleigh modes with periods between about 20 to 120 s and generated synthetic seismograms for I-NDI and II-SHL observations (Figures 1 and 2). Both the synthetic seismograms and the observed seismograms (paths I-NDI and II-SHL, Figure 2) were low-pass-filtered with a cut-off frequency at 0.04 Hz (25 s) to eliminate excess high-frequency oscillations that are superimposed on the long-period signals of interest. The finite filter length tends to introduce different effects on the ends of the synthetic seismograms from those of the observations and is a shortcoming for relatively short distances, as in our cases. The abrupt beginnings of the synthetic seismograms results from truncating the earlier part of the time series, which is not of interest here. The following discussion deals with simple experiments of the effects of the different parameters on the time domain observations.

Fault plane solutions for both events are held fixed, and the uncertainties associated with them have small effects on the calculated waveform. A point source time history is assumed to be a step for simplicity. The severe attenuation for this region, proposed by Bird and Toksoz [1977], was applied to generate the theoretical seismogram marked with 'low Q' in Figure 7a. The one marked 'normal Q' was generated with the observed attenuation for Rayleigh waves in an average earth structure (summarized by Kovach [1978]). Although the amplitude for the normal Q synthetic seismogram is 35% larger than that of the low Q
Fig. 5c. Theoretical Rayleigh wave group velocity dispersion curves (top) for three models (bottom) with a crustal thickness of 85 km for different $\beta_m$. Model E4 is not referred to. Layout same as in Figure 5a.

The synthetic seismogram, the waveforms are quite similar in the bandwidth of interest here, 60 to 90 s (Figure 7a). For both synthetic seismograms the same velocity model S7 (70-km-thick crust, $\beta_m = 4.7$ km/s, Figure 5) and a source depth of 5 km were used.

Although the observed waveform of the II-SHL path is also shown in Figure 7a, the purpose of this figure is to illustrate the effects of a very low Q. Part of the mismatch between the synthetic and observed signals arises from the end effects due to a finite filter length. The coda of the S wave and the 20-s period Rayleigh wave are the most noticeable sources of noise (Figure 2a). The calculated Rayleigh waves are delayed in time with respect to the observed ones in Figure 7a. The maximum mismatch in group arrival time corresponds to 2.7% difference in group velocity, which agrees with the dispersion curves in Figure 5 and is within the observational errors.

Figure 7b shows three synthetic seismograms with their peak-to-peak maximum amplitude normalized to a common size, assuming model S7 with different source depths of 5, 10, and 30 km (the actual amplitude ratio is about 3:2:1, respectively). The difference in waveforms for sources at 5 and 10 km is small. There is a slight difference in waveform for a source at 30-km depth. This can be attributed to the interference of the short-period (<40 s) Rayleigh waves, which have a phase shift of $\pi$ with respect to the longer periods and much weaker amplitudes than for a depth of 10 km [Tsai and Aki, 1970].

Synthetic seismograms were generated for three different velocity models: $\beta_m = 4.4$ and 4.7.
km/s for a crustal thickness of 70 km (S4 and S7) and $\beta_m = 4.7$ km/s for a crustal thickness of 55 km (F7) (Figure 7c). A 5-ka source depth and 'average-earth' Q structure were used. With a constant $\beta_m (4.7$ km/s), models with 55- and 70-km-thick crusts (model F7 and S7) gave quite different synthetic seismograms. We consider model F7 to be inadequate to explain the observation but model S7 to be satisfactory in matching the observed waveform of the I-NDI path (Figure 7c).

**Synthesis of A PL Phase**

We observed one very clear, long-period PL phase that crossed Tibet (Figure 2a). We tested various plausible structures by synthesizing the observed waveform and found that the structure
Fig. 6b. Theoretical Rayleigh wave group velocity dispersion curves (top) for three models (bottom). Model SS3 in Figure 6a is plotted for comparison. The addition of a relatively high velocity layer (β = 3.9 km/s) makes model SS5 (=model S7, Figure 5) fit the data at both long-period and short-period ends. Models S7 (SS5) and SE have nearly identical group and phase velocity dispersion curves. The slight increase in β in SE has the same effect as the introduction of a layer at the base of the crust with β = 3.9 km/s in SS5.
Fig. 7. (a) Two synthetic seismograms for different Q structures together with low-pass-filtered vertical component Rayleigh wave of the II-SHL path observation. Both synthetics employ model S7 (Figure 3) with source depth 8 km and an arbitrary seismic moment. All the seismograms have been normalized to have the same maximum peak-to-peak amplitude, emphasizing their similarity in waveforms, and for each the portion earlier than 590 s have been truncated. (b) Three low-pass-filtered synthetic seismograms of the vertical component of Rayleigh waves for model S7 with different source depths of 5, 10, and 30 km. Amplitude scales have been normalized, as in Figure 7a. (c) Low-pass-filtered vertical component of the Rayleigh wave signal from the I-NDI path together with three synthetic seismograms for different velocity structures: models F7, S4, and S7 (Figure 5). Amplitude scales have been normalized, as in Figure 7a. Note the particularly poor fit of model F7.

obtained from the surface wave dispersion curves were compatible with the observed wave form (see Appendix A on microfiche). Unfortunately, however, the resolving power of PL proved to be inadequate for this analysis to put additional constraints on the structure.

Refraction Profiles

It is well-known in classical seismology that the Pn and Sn velocities can be estimated from the travel time versus distance plot using close earthquakes [e.g., Bullen, 1963, p. 194]. This involves using many stations for each event. However, since to our knowledge, Lhasa (LHA) has been the only permanent seismograph station in Tibet, the usual method must be modified to use travel times from many earthquakes at a single station.

In practice, different earthquakes are reported with different focal depths. Moreover, the focal depths can easily be in error, causing a corresponding error in the origin time. In either case, it is necessary to normalize the focal depth to a common depth in order to avoid introducing artificially large residuals due to different focal depths and therefore different path lengths. This can be done by adjusting the origin time with a depth correction calculated from equation (3) of Bullen [1963, p. 194], assuming an approximate velocity structure. The differences between the assumed and actual origin time is assumed to be absorbed by this depth correction completely. Any incomplete trade-off between the focal depth and the origin time, together with a small error induced by an error in the assumed velocity structure, will cause a possible systematic error in the measured travel time. Because this error is presumably independent of distance in the normalized travel times, it should introduce only random errors in the evaluation of the velocity and therefore only affect the variance associated with the velocity determination, not the estimate of the velocity.

Pn and Sn arrival times were taken from the

1Appendices are available with entire article on microfiche. Order from the American Geophysical Union, 2000 Florida Avenue, N.W., Washington, D.C. 20009. Document J81-004; $1.00. Payment must accompany order.
Fig. 8. (a) Plots of $P_n$ travel time residuals normalized for a focal depth of 33 km versus distance to LHA for earthquakes north of the Indus Tsang-Po suture. The distance range is about 3°-17°. The $P_n$ velocity (8.15±0.04 km/s) and travel time residuals were obtained by a least squares fitting of the 61 observations, assuming a Moho depth of 70 km. Most of the residuals fall within ±1 s of the mean. The dashed line shows the slope of a curve to which the data would be parallel if the $P_n$ velocity were 7.9 km/s. Travel time residuals can be viewed as reduced travel times in conventional refraction analysis. (b) Plots of $P_n$ travel time residuals without depth normalization versus distance to LHA for earthquakes north of the Indus Tsang-Po suture. Distance range is 3°-10°. (c) The same data set as in Figure 8b but with a normalization for a focal depth of 33 km. Note the reduction of residuals compared with Figure 8b. (d) Plots of $P_n$ travel time residuals normalized for a focal depth of 33 km versus distance to LHA for earthquakes south of the Indus Tsang-Po suture. Distances range from 3° to 10°. (e) Plots of $S_n$ travel time residuals normalized for focal depths of 33 km versus distance to LHA for earthquakes north of the Indus Tsang-Po suture. Distances range 3° to 10°.

Lhasa station reports available between 1960 and 1965 and again between 1971 and 1973 [Institute of Geophysics, 1966, 1974]. We selected well-located events for which more than 35 readings from seismographic stations were used in the location. We used the epicenters, origin times, and depths given in the Bulletins of the International Seismological Summary (ISS) and the International Seismological Center (ISC).
<table>
<thead>
<tr>
<th>Epicentral Distance Range, deg</th>
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<td>all P_n</td>
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<tr>
<td></td>
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<td>19</td>
<td>x/(8.24±0.09) + (12.0±5.4)</td>
</tr>
<tr>
<td>3 - 10</td>
<td>sharp P_n</td>
<td>39</td>
<td>x/(8.20±0.04) + (10.6±2.6)</td>
</tr>
<tr>
<td></td>
<td>sharp S_n</td>
<td>23</td>
<td>x/(8.38±0.21) + (14.5±15.0)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>x/(8.18±0.13) + (11.2±8.8)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>x/(7.99±0.12) + (9.7±6.9)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>x/(8.12±0.06) + (10.0±3.6)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>x/(4.77±0.09) + (25.4±11.4)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>x/(4.77±0.08) + (23.5±9.5)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Regions South of the Suture</td>
</tr>
<tr>
<td>3 - 17</td>
<td>all P_n</td>
<td>24</td>
<td>x/(8.20±0.11) + (10.4±5.2)</td>
</tr>
<tr>
<td></td>
<td>(mostly</td>
<td></td>
<td>x/(8.11±0.07) + (9.1±3.6)</td>
</tr>
<tr>
<td></td>
<td>sharp)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3 - 10</td>
<td>all P_n</td>
<td>21</td>
<td>x/(8.29±0.15) + (11.0±5.8)</td>
</tr>
<tr>
<td></td>
<td>(mostly</td>
<td></td>
<td>x/(8.13±0.10) + (9.3±4.0)</td>
</tr>
<tr>
<td></td>
<td>sharp)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>sharp S_n</td>
<td>11</td>
<td>x/(4.94±0.11) + (28.3±8.8)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>x/(4.75±0.10) + (23.2±8.6)</td>
</tr>
</tbody>
</table>

Results after the depth correction are given on the second line. All velocities are calculated for a Moho at 70-km depth.

Estimated velocity should be increased by 0.5%. The P_n velocity is determined to be 8.152±0.04 km/s, compared with 8.27±0.07 km/s if no focal depth correction is applied (Table 2). If the apparent velocity were in fact 7.9 km/s, then the data should be more nearly parallel to the dashed line in Figure 8a. Such a low P_n velocity will not fit the data well unless there is some special distance dependent systematic error in locations, depths, or origin times. If we limit ourselves to sharp P_n arrivals (excluding those reported as ef or eP_n in the station reports), we obtain a P_n velocity of 8.20±0.04 km/s from 42 arrivals. Without the focal depth correction, we obtain 8.24±0.09 km/s (Table 2).

If we consider only the 39 sharp arrivals between 3° to 10° (Figures 8b and 8c), the P_n velocity is estimated to be 8.12±0.06 km/s for the depth corrected case (Figure 8c). This is the only case in all the profiles discussed in this section for which the P_n velocity determined without a depth correction, 8.0±0.1 km/s (Figure 8b) is the smaller. A total of 19 emergent arrivals between 3° to 17° gave an estimate of the velocity of 8.18±0.13 km/s. Thus the entire data set is very consistent (Table 2).

P_n across the Himalaya. The data discussed above includes only events north of the Indus Tsang-Po suture because of the possibility of a different structure beneath the Himalaya. However, recordings at Lhasa for 24 events south of the suture, for nearly all of which sharp readings were reported, yields a P_n velocity of 8.11±0.07 km/s. The epicentral distances range from 3° to 17°. Among these, 21 events occurred within 10° of Lhasa (Figure 8d). Despite the scatter, the same estimate for P_n velocity is obtained. Thus no significant difference in the P_n velocity is observed for regions north and south of the suture zone. S_n across the Tibet and the Himalaya. There are also 23 sharp S_n arrivals reported at Lhasa from events north of the suture in the distance range 3° to 10°. (Only three such arrivals were reported for distances greater than 10°.) After the depth normalization, the S_n velocity is estimated to be 4.8±0.1 km/s (Figure 8e). There is a much larger scatter in the S_n residuals than the P_n residuals probably because of the lower S_n velocity and because of difficulties in identifying the arrival on the seismograms. A low S_n velocity of 4.5 km/s would not fit the data well (Figure 8e). The 11 sharp arrivals for events south of the suture, reported at Lhasa within 3° to 10°, gave the same S_n velocity (Table 2).

Intercept times and crustal thicknesses. Despite the large uncertainties, the intercept times for P_n all suggest the depth of the interface (the Moho) to be 65±30 km (one standard deviation). S_n intercept times have even larger uncertainties, but the depth of the interface calculated from them fall within the range (Table 2) of 80±50 km. The epicenters of all the events used in this refraction study show a broad azimuthal coverage and distribution of epicentral distances with respect to Lhasa (Figure 9). Nevertheless, because purely reversed profiles cannot be made, it is remotely conceivable that a dipping Moho or a systematic horizontal variation in the velocity structure centered around Lhasa might render the real mantle velocities different from the measured apparent velocity. For models with a crustal layer with Vp ranging from 5.9 to
6.5 km/s overlying a mantle half-space with \( v_p \) ranging from 7.9 to 8.4 km/s, a uniform dip of about 0.8° of the Moho could add an uncertainty of ±0.1 km/s to the estimated \( P_n \) velocity. Such a dip would correspond to a difference in crustal thicknesses of 15 km over 10° distance. Therefore, if Lhasa were to overlie a maximum depth to the Moho beneath Tibet, such that the crustal thickness decreased by 15 km, 10° to both the north and west, then the \( P_n \) velocity could be overestimated by 0.1 km/s. We think that because the average elevation of Tibet is so uniform, a drastic variation in crustal thickness is unlikely, and it would be fortuitous if Lhasa were to overlie the deepest crust or the center of a systematic horizontal variation of the velocity structure.

Summary of refraction results. High values of the \( P_n \) and \( S_n \) velocity beneath Tibet are obtained from refraction profiles using earthquakes in Tibet and arrival times at Lhasa. Assuming a Moho at 70 km depth, the \( P_n \) velocity is estimated to be 8.12 km/s with a minimum formal standard deviation of 0.06 km/s covering the region north of the suture from 5° to 10° from Lhasa. An uncertainty of 0.1 km/s probably exists for all \( P_n \) velocity estimates reported here. No significant difference in \( P_n \) velocity was found for the regions north and south of the suture. With a somewhat larger uncertainty, the velocity of \( S_n \) is similarly determined to be 4.8±0.1 km/s. Both \( P_n \) and \( S_n \) velocities are about the same as those of the shields and stable platforms. All the intercept times have large uncertainties, and they indicate a depth of the Moho of 65±30 km.

Telesismic S-P Travel Time Residuals

To our knowledge, the high velocities for the uppermost mantle under Tibet discussed in the last section have not been proposed before. The vertical extent of the high-velocity zone is of great importance in the interpretation of the tectonics. The refraction profiles require a layer beneath the Moho with a thickness of only about 10 km. The surface wave dispersion curves are not very sensitive to the velocity structure below about 100 km. The telesismic travel time measurements discussed in this section are intended to serve as a crude estimate of the vertically averaged velocity structure to a greater depth (~300 km).

Telesismic P and S wave arrival times were measured from the records of the WSSN stations for nine earthquakes in Tibet and eight earthquakes in the Himalaya (Figure 10, Table 3). Arrival times were measured only for S wave signals that are clear on the long-period records (Figure 11). Uncertainties are about 1 s. To avoid inclusion of converted phases (e.g., Sp), we tried to measure only the transverse components of the S waves. S wave polarization angles were checked with known P wave fault plane solutions [Molnar and Tappin, 1978; Molnar et al., 1973, 1977; Ni and York, 1978]. We considered only the distance range between 30° and 80° to avoid triplication in the S wave travel time curve and possible misidentification of SKS for S.

Both P and S wave travel time residuals, with respect to the Jeffreys and Bullen [1940] (J-B) travel time tables, were examined as a function of epicentral distance, using locations and origin times given by the ISC. Since the hypocenters and origin times reported by ISC were calculated primarily using the J-B P wave travel times, the P wave residuals were automatically minimized. Because late (or early) S waves are likely to be associated with late (or early) P waves, the S-P residuals measured here probably
Fig. 10. Map of Tibet and the Himalaya showing the locations of 17 earthquakes (circles) used in teleseismic travel time measurements. Roman numerals inside the circles are the S-P travel time residuals with respect to the J-B tables. All the values are rounded to the smaller nearest integer and are given at 1-s intervals. Source depths have been normalized to 5 km for earthquakes in Tibet and to 33 km for those in the Himalaya.

give lower bounds to the S wave residuals.

For nine earthquakes in Tibet, P wave residuals scatter around the mean as expected from the hypocenter location procedure, whereas S wave travel times show large delays (Figure 11). The measured P and S wave travel time residuals are given in Appendix B on microfiche. The S-P travel time residuals have correspondingly large delays (Table 3, Figures 12a and 13). Applying station corrections [Sengupta, 1975; Sengupta and Julian, 1976] to these data does not significantly reduce the scatter but increases the S-P travel time residuals by about 1 s (Table 3, Figures 12a and 13).

With the same procedure, we studied eight earthquakes in the Himalaya (Figure 12b) and observed smaller S-P travel time residuals than for earthquakes in Tibet (Table 3, Figures 12b and 13). The rays from earthquakes in Tibet and the Himalaya to a given station more than 30° from the epicenters will have nearly identical paths except near the source region. Thus we infer that the large S-P travel time delays for earthquakes in Tibet are due to the medium properties of the region beneath Tibet and not to systematic errors in the travel time tables or to the medium along the rest of the ray paths.

In general, there is a strong trade-off between the focal depth and the origin time for events located teleseismically. Erroneous depths could also, at least partially, be responsible for the large S-P travel time delays for earthquakes in Tibet if these earthquakes actually occurred at a shallower depth than was obtained in the location procedure. The S-P travel time residuals were calculated for each event assuming three possible depths: their depths given by the ISC, 5 km, and 33 km (Figure 13). To make an extreme comparison, assume all earthquakes in Tibet occurred at 5 km depth, but all earthquakes in the Himalaya occurred at 33 km depth. On the average, there is still a difference of approximately 2 s between the S-P residuals from earthquakes in these two regions (Figure 13, Table 3).

Thus the average S-P travel times are late for events in Tibet not only with respect to the J-B tables which represent an averaged earth model but also with respect to those for events in the Himalaya. Because low P wave velocities probably cannot be detected through routine location procedures using the P wave arrival times alone, a delay of 1 s in P wave travel times from the average earth could easily exist. This would lead to another second in delay in the S wave travel time.

The measured S-P travel time residuals, rounded to the smaller nearest integer, are given in 1-s intervals in Figure 10 together with the locations of the 17 earthquakes. For this figure, the focal depths for Tibetan events are normalized to 5 km and for Himalayan events, at 33 km. Although our data points certainly do not allow a meaningful contouring of these residuals in Tibet, the two earthquakes in the center of Tibet have especially large S-P travel time residuals (>5 s), and the distribution of S-P travel time residuals in the Himalaya is relatively uniform. The immediate inference from the late S-P travel times in Tibet is that the relatively high Pn and Sn velocities obtained from the refraction profiles discussed in the last section are not likely to extend to a great depth beneath Tibet.

Seismic Wave Velocity Structure

Observed group and phase velocity dispersion curves of Rayleigh waves provide constraints on crustal thickness and the uppermost mantle shear
wave velocity. For possible thicknesses of 55 km to 70 km, or event 85 km, the corresponding uppermost mantle shear wave velocities of 4.4 km/s or less to 4.7-4.8 km/s, or even as large as 4.9 km/s, are required, respectively. We consider an $S_n$ velocity of 4.9 km/s to be unlikely and therefore conclude that the crust probably is not as thick as 85 km.

The paths for the observations of surface waves and the $P_b$ phase all cross the Himalaya. This apparent lateral inhomogeneity introduces an additional uncertainty that is difficult to quantify. The two clear long-period Rayleigh wave signals studied here cross the Himalaya at very different angles (Figure 1), and the 3% difference in the group velocities could be a consequence of lateral heterogeneity.

Nonetheless, we cannot confidently attribute this difference solely to real structural differences because of the uncertainties in the group velocity determinations.

Since both the $P_n$ and $S_n$ velocities were estimated to be relatively high from the refraction profiles, $P_n$ velocity $8.12\pm0.06$ km/s, $S_n$ velocity $4.8\pm0.1$ km/s, the crustal thickness is apparently about 70 km. It is difficult to attach a formal uncertainty to this thickness, but from the comparison between the observed and theoretical dispersion curves, we estimate that the crustal thickness is between 65 km and 80 km (Fig. 5).

Travel time measurements for earthquakes in Tibet to stations at distances of 30° to 80° indicate late $S$ waves and long delays between $P$ and $S$. Since the Tibetan earthquakes have late S-P travel times compared with earthquakes in the Himalaya, the average velocity beneath Tibet must be lower than that of the Himalaya. The relatively high-velocity zone beneath the Maho of Tibet, estimated from the refraction profiles, therefore probably is not thick.

The S-P travel time residuals can be used to estimate differences in average upper mantle velocities beneath Tibet and stable platforms or shields with some additional assumptions. Suppose all such differences were confined to the upper 240 km, as inferred for shields and ocean basins [Dziewonski, 1971; Oka], 1977]. Also
assume that the measured S-P travel time residuals are estimates of S wave travel time residuals. Let us compare the Tibetan models with model S2 of Dziwonski [1971] for the shield regions but modified to include 70-km-thick crust. As an extreme, let us assume that the earthquakes in Tibet occurred at a depth of only 5 km. Then if one increases the velocities of model S2 (without a 70-km-thick crust) by about 1% to correct for the effect of velocity dispersion due to attenuation [Hart et al., 1977], the vertical one way S wave travel time between the depths of 0 and 240 km for model S2 is 53.8 s. This is about 1.1 s faster than that for the Jeffreys-Bullen model. Therefore the observed average S-P travel time delay of 3.6 s with respect to the J-B tables for earthquakes in Tibet corresponds to an average difference of 4.7 s in S wave travel time from the unmodified model S2. Of this 4.7 s, 2 s can be attributed to the effect of a thick crust. The resulting S wave travel time difference of 2.7 s corresponds to a reduction of the shear wave velocity of 6% between 70 to 240 km from that of model S2. If we assume that models S7, S8 (70 km thick crust, $v_s = 4.7-4.8$ km/s, Figures 5 and 6), or P4 (55 km thick crust, $v_p = 4.4$ km/s, Figure 5) were appropriate for Tibet in the depth range from 0 to 100 km instead of S2, then the average shear wave velocity beneath Tibet must still be about 4 to 5% lower than that of the model S2 for shields between the depths of 100 and 240 km. Insofar as Dziwonski's [1971] model S2 is appropriate for the shields, this would correspond to the average shear wave velocity of about 4.3 km/s between 100- and 240-km depth beneath Tibet. Because of S-P residuals are lower bounds to the S wave delays and because we assumed extreme for the depths of the events in Tibet, the average shear wave velocity beneath Tibet could be still lower.

The effect of velocity dispersion due to attenuation has been neglected in the above discussions of crustal thickness. It is difficult to estimate the contribution from such velocity dispersion without sufficient data on both attenuation and phase velocity [e.g., Lee and Solomon, 1976]. Qualitatively, if strong attenuation exists, the estimate of shear wave velocities deduced from surface waves should be increased by a small amount (≈1%) above those considered here before comparing them with body wave observations [Hart et al., 1977]. Thus there might be a small systematic overestimate of a few kilometers of the crustal thickness, but this overestimate is probably much less than the uncertainty quoted above.

### Tectonic Implications

#### The Thick Crust

The apparently very thick crust (~70 km) beneath Tibet implies that the crust has been thickened approximately by a factor of 2 since the collision between India and Eurasia. This is consistent with the idea of horizontal compression and shortening [Deeke and Burke, 1973]. The hypothesis of underthrusting the Indian plate beneath the Eurasian plate can be argued as still possible if somehow either the crust of Tibet were detached from its underlying mantle while the Indian plate was subducted along the base of the crust or if the Indian crust plunged into the mantle and then migrated, probably as a melt, up to the base of Tibet's crust. In other words, the continental crust of the order of 1000-km length would have been consumed and recycled back near the surface in a time span of about 40 m.y. The current seismic data cannot discriminate between these two extremes, but the relatively low average velocity of the crust implies that the crust is relatively hot. If cold Indian lithosphere were thrust under Tibet, it would have to have been heated since then.
Fig. 12a. S-P travel time residuals (circles) with respect to the J-B tables versus distance for earthquakes in Tibet. The crosses are the results after station corrections have been added. Dashed lines show the zero-residual level for a focal depth normalized to 5 km. The averaged S-P delays are given in Table 3 with estimated standard errors.
Fig. 12b. S-P travel time residuals with respect to the J=8 tables versus distance for earthquakes in the Himalaya. Symbols are the same as in Figure 12a.

Uppermost Mantle Seismic Velocities

Both $P_n$ and $S_n$ velocities indicate a high-velocity uppermost mantle under Tibet. The late S-P travel time residuals for earthquakes in Tibet indicate that these high velocities do not reflect the properties of the deeper part of the upper mantle. Although the interpretation of the travel time residuals are highly nonunique, this high-velocity layer extending no deeper than about 100 km under Tibet is consistent with the observations.

Implications for the temperature in the mantle. The apparently high uppermost mantle velocities under Tibet are not directly comparable to those under the stable platforms and shields because the extremely thick crust in Tibet will create about 10 to 12 kbar or additional lithostatic pressure at the base of the thickened Tibetan crust. Assuming that the pressure and temperature dependence of seismic wave velocities for olivine are representative of the mantle [e.g., Anderson et al., 1968; Birch, 1969], 10 kbar of pressure causes an increase in
$V_p$ of about 0.1 km/s and a temperature increase of about 250°C causes a similar decrease. $V_p$ has very similar pressure-temperature dependence to that of the $V_s$, for most olivine group minerals. Exceptions are the forsterite samples measured by Anderson et al. [1968], which have been questioned by Birch [1969]. Thus, if the $P_n$ and $S_n$ velocities were the same beneath stable platforms and Tibet and if the temperature and pressure dependencies of the seismic velocities were those appropriate for olivine, then the temperature of the Moho beneath Tibet would be about 250°C hotter than that of the platform. This estimate of relative temperature would have an uncertainty as large as 120°C, even if the resolution of the difference in $P_n$ velocity for Tibet and shields were only ±0.05 km/s. Thus this suggested difference in temperature is only a crude estimate, not a precise measurement.

The temperature under platforms and shields are difficult to estimate. From heat flow measurements, Sclater et al. [1980] estimated a lower bound on the temperature for the Canadian shield to be less than 400°C at the base of the crust. On the other hand, if we take the measured reduced heat flow as an upper bound of the mantle heat flow for Eastern U.S. (0.80±0.05 HFU [Lachenbruch and Sass, 1976]) and the Indian shield (0.93 HFU [Rao et al., 1976]), the mantle heat flow alone gives 470°C and 540°C at the base of the crust, respectively, without considering the effect of radioactivity and assuming a crustal thickness of 35 km and thermal conductivity of $6 \times 10^{-2}$ cal/cm°C s. Uncertainties in conductivity and heat flow measurements alone will probably add 20% uncertainties to those estimates. In any case, if the temperature at the base of the crust beneath platforms and shields were about 500°C, then at the Moho beneath Tibet it could be about 750°C. Although the uncertainty in this number is very large, our point is that the high $P_n$ and $S_n$ velocities do not require an unusually cold upper mantle.

Implications for the strength of the mantle. At first glance, the relatively high $P_n$ and $S_n$ velocities are likely to be taken as evidence for a cold and therefore strong upper mantle. Because crustal shortening requires strong deformation of the underlying mantle, the inference of high strength might be taken as evidence contradictory to crustal shortening. In this section we contend that the current uppermost mantle is not very strong.

The temperature at the Moho strongly affects the mechanical behavior of the lower crust and
upper mantle. One can estimate the mechanical strength of the uppermost mantle beneath Tibet assuming that the flow laws for steady state creep of olivine are representative of the uppermost mantle rheology [Goetze and Evans, 1979; Kirby, 1977; Tappanier and Francheteau, 1978]. Note that the uncertainties in the estimate of activation energy and temperature will dominate the uncertainty of the estimate of the differential stress or strain rate in the flow law. Assuming a strain rate corresponding to 10% strain for the past 40 m.y., since the collision and using values given by Goetze [1978] for olivine, the calculated differential stress (σ1-σ3) is 1.1 kbar at 750°C or 430 bars at 800°C. If the temperature were as low as 700°C, the calculated differential stress would be larger than 2 kbar. Post [1977] obtained a quite different olivine flow law for Mt. Burnett dunite. Using his 'dry' flow law under the same assumptions, differential stress is estimated to be 3.7 kbar at 800°C while the 'wet' flow law requires only 30 bars of differential stress under these conditions. If Coble creep (synthetic recrystallization), instead of dislocation creep, however, is the predominant deformation mechanism, the calculated differential stress is only 210 bars at 800°C and 430 bars at 700°C at the given strain rate [Goetze, 1978]. Thus the lower bound of the differential stress required to deform the uppermost mantle under Tibet at the given strain rate through steady state ductile flow of olivine is only about 0.4 kbar. Although the estimate of temperature beneath Tibet is too uncertain to prove that the uppermost mantle is not strong, the apparently high Pn and Sn velocities beneath Tibet do not require high mechanical strength for the uppermost mantle.

Implications on volcanism in Tibet. It is interesting to note that at temperatures of about 800°C and 20-kbar pressure, incipient melting of crustal material (but not the uppermost mantle) is possible in the presence of a small amount of water. Figure 14 shows solidi for some likely crustal and upper mantle materials based on data from experimental petrology [Lambert and Wyllie, 1970; Merrill et al., 1970; Merrill and Wyllie, 1975]. None of the data have error bars, and therefore they can only be used qualitatively. Nevertheless, the solidus for water saturated peridotite is over 1000°C despite the pressure [Merrill et al., 1970]. With only 0.1 wt % of water, partial melting of gabbroic material is possible at 20-25 kbar pressure and between 700°C to 800°C [Lambert and Wyllie, 1970]. The solidus for peridotite with 0.1 wt % water at this pressure range is over 1250°C. Thus for Tibet there could be a geotherm such that a partially molten crust overlies a subsolidus uppermost mantle. The inference of a high-velocity mantle lid on the low velocity zone, then, is not incompatible with the widespread volcanism over Tibet unless the volcanic activity is predominantly of uppermost mantle origin or unless the lower crust of Tibet consists of only anhydrous minerals.

The geochemical evidence gathered so far is not sufficient to reach any conclusion on whether the lower crust of Tibet is hydrous or not. Simply based on the fact that no tholeiitic basalt has been reported [Hennig, 1915, Deng, 1978], it is believed that the volcanic activity involved mostly melting of crustal materials [e.g., Dewey and Burke, 1973]. However, the unusually low average crustal velocity for Tibet is also consistent with high temperatures and melting.

A Discussion of the Thermal Evolution of the Plateau

The presence of widespread volcanic activity and diffuse seismicity, with prominent normal faulting in Tibet [Molnar and Tappanier, 1978] all indicate that Tibet is tecstrically active at present. Yet the geothermal gradient in the crust should have been decreased by crustal thickening. Thus the source of heat deserves some consideration. There are three apparent heat sources that contribute to Tibet's crustal geotherm: strain heating from the thickening of the crust, radioactive heat production in the crust, and the mantle heat flux from below the crust.

Since the crust of Tibet seems to be twice as thick as normal continental crust, the average strain rate in the crust is likely to be of the order of 1/40 m.y. Assuming an averaged differential stress of 1 kbar throughout Tibet, the stress heating rate per unit volume (stress x strain rate) is only about 8 x 10^{-7} erg/cm²/s or 0.2 x 10^{-13} cal/cm²/s (0.2 HGU). This is considerably smaller than the presumed radioactive heat production rate of about 1 HGU for lower crustal minerals [e.g., Tilling et al., 1970] and can be neglected.

We investigate the effects of the mantle heat flux and radioactive heating in the crust by considering a simple one-dimensional heat conduction problem assuming that a constant heat flux (Qh) is held at the base of the crust. The initial conditions are such that thickening of the crust is instantaneous and the original geotherm is displaced accordingly. To parameterize the crustal heat source, we let
Tibet define a heat flow province such that the empirical relationship

\[ Q_b = Q_h + D \alpha \]  \hspace{1cm} (1)

between surface heat flow \( Q_b \) and surface heat production rate \( A \alpha \) is valid [Lachenbruch, 1970]. Finally, we neglect the effects of erosion. This assumption is probably justified by the facts that the Cretaceous limestone is still widespread in southern Tibet [Hennig, 1915; Norin, 1946] and that the present-day drainage system in Tibet is internal.

We separate the temperature \( T \) in the crust into the contributions from the mantle heat flux \( T_m \) and from radioactive heat production \( T_R \), and three possible distributions of radioactive heat-producing elements in the crust are considered separately. The detailed derivations are given in Appendix C on microfiche and by Chen [1979]. We will only summarize the most important results here. Throughout this discussion, the crustal thickness is taken to be thickened by a factor of 2 to a value of \( 70 \) km. We also assumed that the conductivity \( k = 6 \times 10^{-3} \text{ cal/cm} \cdot \text{°C} \cdot \text{sec} \) and the thermal diffusivity \( \kappa = 0.01 \text{ cm}^2/\text{sec} \), and we shall concentrate our discussions only on the temperature at the base of the crust where it will be highest in the crust.

First, consider only \( T_m \) with two starting temperatures \( T_m = 300^\circ \text{C} \) and \( 600^\circ \text{C} \), corresponding to mantle heat flow \( Q_h = 0.86 \text{ μcal/cm}^2 \cdot \text{sec} \) (HFU) and \( 1.03 \text{ HFU} \), respectively, at steady state. \( T_m \) will reach \( 800^\circ \text{C} \) in about \( 45 \) and \( 14 \) m.y., respectively.

The contribution from radioactive heat production strongly depends on the distribution of those elements, parameterized by the constant \( D \), in equation (1). The steady state temperature due to radioactive heat production \( T_R \) is approximately proportional to \( D \). Also, if the differentiation of the radioactive heat-producing elements is much faster than heat conduction so that \( D \) does not change during crustal thickening, then \( T_R \) contributes little to the crustal geotherm. This occurs because in most models and probably in the earth also, the source of heat is far from the base of the crust. Otherwise, \( T_R = 4T_R \) and after \( 40 \) m.y., \( T_R = 2T_R \), where \( T_R \) is the contribution to the initial geotherm due to radioactive heat. For instance, at \( 70 \) km depth, for a uniform distribution of crustal heat source with \( Q_h = 0.75 \text{ HFU} \), \( D = 25 \) km and \( \alpha = 0.2 \mu \text{gU (T} = 240\text{°C)} \), in \( 45 \) m.y., \( T = 800^\circ \text{C} \). The steady state final heat flow at the surface would be \( 1.25 \text{ HFU} \) for this case. Since heat flow measurements are not available for Tibet, these hypothetical values are only meaningful as plausibility arguments, and they should not be taken as fact. Nonetheless, the above calculations suggest that it is possible that the lower crust of Tibet could reach \( 700^\circ - 800^\circ \text{C} \) in \( 40 \) m.y., simply from the recovery of the geotherm provided that the heat flux from the mantle is greater than about \( 0.9 \) HFU regardless of the heat production in the crust. With a reasonably warm starting temperature of \( 490^\circ \text{C} \) at the Moho and if there is sufficient radiogenic heat production in the crust to produce about \( 0.5 \) HFU after the crustal thickening, then for a mantle heat flux of only \( 0.75 \) HFU the calculated temperature at the Moho could reach \( 800^\circ \text{C} \) after \( 40 \) m.y. Thus radioactive heat production could be important for the recovery of crustal geotherm in \( 40 \) m.y., especially if radioactive heating also makes a significant contribution to the original geotherm [Bird et al., 1975]. Similar conclusions on the importance of radioactive heating are reached by England [1979] for the thickened Archaean crust.

Crustal thickening beneath Tibet by horizontal shortening [Dewey and Burke, 1973] seems to be compatible with both the seismic observations and these thermal calculations. Yet the behavior of the lithosphere as a whole during the continental collision is not clear. In fact, the assumptions in these calculations are directly related to this question. If we assume that at \( 70-\text{km} \) depth the mantle heat flux is about \( 0.9 \) HFU and that the temperature at the Moho is about \( 800^\circ \text{C} \), then for a coefficient of thermal conductivity \( k \) of \( 7.5 \times 10^{-3} \text{ cal/cm} \cdot \text{°C} \cdot \text{sec} \) in the mantle, the \( 1300^\circ \text{C} \) isotherm will be reached at about \( 112 \) km. If indeed the lithosphere were maintained at an approximately constant thickness so that a heat flux of about \( 0.9 \) HFU is supplied at the base of the crustal layer, then the estimated temperature at the base of the Tibetan crust could be explained simply in terms of the recovery of the geotherm with little contribution by radioactive activity. The volcanic activity, at least in part, could be explained by the depression of the lower crust to depths where the temperature is eventually high enough for it to melt. No anomalous heat flow from the asthenosphere would be required.

Presumably small-scale convection maintains an essentially constant heat flux at the base of the plate. Parsons and McKenzie [1978] describe this in terms of the stability of a thermal boundary layer above the viscous asthenosphere and with a thickness \( 6 \) between two isotherms. If a critical Rayleigh number \( R_a \) is exceeded, the boundary layer becomes unstable so that small-scale convection occurs and provides convective heat transport. If the whole lithosphere with the thermal boundary layer thicken by horizontal shortening, \( 6 \) would increase and \( R_a \) would increase as \( 6 \). Therefore crustal shortening and thickening would be a very effective way of enhancing the instability of the boundary layer and causing small-scale convection. Effectively, there would be a detachment of the bottom of the thickening lithosphere that might tend to keep the lithosphere from continuing to thicken while the crust does. A more quantitative treatment of this problem is given elsewhere [Houseman et al., 1981].

Conclusions

1. The crust beneath Tibet is about 65 to 80 km thick with an average shear wave velocity of less than 3.5 km/s from surface wave dispersion, refraction profiles, and \( P \) phase synthesis.

2. \( P \) and \( S \) velocities were estimated to be 8.12±0.06 km/s and 4.8±0.1 km/s, respectively.

3. The averaged vertical shear wave velocity between 70- to 240-km depth is about 6% lower than that of the shield model 52 [Dziewonski, 1971], and this average velocity beneath Tibet is
also lower than that below the Himalaya by about the same
amount.

4. From the $P_n$ and $S_n$ velocities and the
pressure-temperature dependence of the seismic
velocities for olivine, the temperatures at the
base of the Tibetan crust could be about 250°C
hotter than that beneath shields or stable
platforms. Since the temperature at the Moho
beneath platforms is about 500°C, the base of the
crust beneath Tibet could be near 750°-800°C.

5. At this temperature and depending on the
deformation mechanism, the uppermost mantle is
likely to have a maximum mechanical strength of
less than about 1 kbar (differential stress).
This value decreases rapidly with depth in the
uppermost mantle.

6. Partial melting of the lower crustal
materials but not the uppermost mantle is likely
with these estimated temperatures. The low
crustal velocities are consistent with a hot or
partially molten lower crust.

7. Simple one-dimensional heat conduction
calculations suggest that no anomalous mantle
heat source is required to explain the volcanic
and tectonic activity, provided that the
thickness of the lithosphere does not increase
drastically during crustal shortening such that a
heat flux of about 0.75 to 0.9 HFU (depending on
the amount and distribution of crustal heat
production) is available at the base of the crust.

Acknowledgments. We are grateful to A.
Dainty, who did much work in implementing a
double-couple source to the generalized ray
program STPSYN-1 by D. V. Helmberger and R. A.
Wiggins, M. Bouchon, A. Dziwnowski, and
H. Kanamori kindly made their programs available
to us. J. H. K. N. provided several fault plane solu-
tions in Tibet prior to their publication. K.
Anderson, T. A. Chou, W. Y. Chung, W. Ellsworth,
N. Patton, and C. Stewart helped us in
calculations. We also benefited from

discussions with K. Burke, P. England, B.
Parsons, and P. Tapponnier. This research was
supported by the National Science Foundation
grant 77-23017 EAR and by NASA grant NCCS-8 (NSG
5260).

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(Received July 29, 1979;
revised September 15, 1980;
accepted December 11, 1980.)