Mantle earthquakes in the Himalayan collision zone

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ABSTRACT

These earthquakes are known to occur beneath southern Tibet at depths up to ~95 km. Whether earthquakes occur within the lower crust thickened in the Himalayan collision or in the mantle is a matter of current debate. Here we compare vertical travel paths expressed as delay times between S and P arrivals for local events to delay times of P-to-S conversions from the Moho in receiver functions. The method removes most of the uncertainty introduced in standard analysis from using velocity models for depth location and migration. We show that deep seismicity in southern Tibet is unequivocally located beneath the Moho in the mantle. Deep seismicity in continental lithosphere occurs under normally ductile conditions and has therefore garnered interest in whether its occurrence is due to particularly cold temperatures or whether other factors are causing embrittlement of ductile material. Eclavitzation in the subducting Indian crust has been proposed as a cause for the deep seismicity in this area. Our observation of seismicity in the mantle, falling below rather than within the crustal layer with proposed eclavitzation, requires revisiting this concept and favors other embrittlement mechanisms that operate within mantle material.

INTRODUCTION

The Himalaya-Tibet continental collision zone has featured prominently in a debate on whether strength in the lithospheric resides in two layers in the brittle crust and uppermost mantle separated by a weak lower crust, or in a single layer largely limited to the crust (Chen et al., 1981, 2012; Chen and Molnar, 1983; Zhu and Helmberger, 1996; Henry et al., 1997; Catlin and Avoouac, 2000; Jackson, 2002; Chen and Yang, 2004; Beaumont et al., 2004; Monsalve et al., 2006; de la Torre et al., 2007; Bendick and Flesch, 2007; Liang et al., 2008; Priestley et al., 2008; Craig et al., 2012). The distribution of seismicity with depth in strained regions serves as a proxy for the lithospheric strength profile because the occurrence of earthquakes implies sufficient strength to sustain brittle failure, due to either cold conditions (e.g., Chen and Molnar, 1983; Sloan and Jackson, 2012; Blanchette et al., 2018) or local embrittlement processes (e.g., Incel et al., 2017, 2019; Prieto et al., 2017). The task is therefore to map where exactly local earthquakes occur with respect to layers within the crustal and uppermost mantle column. A difficulty is posed by uncertainty in estimated depths for both local seismicity and the Moho. Crustal earthquake depths are more difficult to constrain than their epicenters because the stations used to locate them are on the surface and there is a strong tradeoff between location depth and P and S velocity structure above the events. Intralithospheric interfaces such as the Moho can be seen with teleseismic converted waves (receiver functions), but determining their depth suffers from the same tradeoff with velocity structure above the interface as in estimating earthquake depths. Comparing hypocentral depths with structural depths is hampered by the uncertainties in both parameters. The comparison is accurate only if the same velocity models are used for depth determination for both data sets. More typically, results are compared between different studies, and a lack of consistency between the velocity models used exacerbates the uncertainties. We propose a simple workaround to the depth determination uncertainty by comparing seismicity and structure directly as delay times between shear and compressional waves (“S minus P”, or S-P), rather than converting both to depth first.

DATA AND METHODS

We used data from the CE 2001–2003 HIMNT (Himalayan Nepal Tibet) seismic broadband experiment (Fig. 1). Of the seismicity located using the network (Monsalve et al., 2006), deep events are seen in Nepal just south of the Lesser Himalaya (Fig. 1, cluster C) and under Tibet (Fig. 1, clusters A and B). Figure 2B shows the same set of events on a previous structural depth profile from receiver functions on a line crossing the collision zone in a N18°E orientation perpendicular to the local range strike (Schulte-Pelkum et al., 2005). The Moho is visible as a contrast in isotropic velocity, and the Main Himalayan thrust as a contrast in anisotropy (Fig. 2B; Schulte-Pelkum et al., 2005). Even in this best-case scenario, with relocations and structural imaging done with the same stations and the same velocity models, the deep seismicity appears diffuse in depth and is difficult to clearly assign to crust or mantle.

Local earthquake depths are usually calculated from the travel-time difference between P and S arrivals from the event at each station. In receiver function studies, delays between teleseismic P and converted S are used to determine depths to interfaces under the station. In both cases, the S-P delay time is mapped to depth by assuming P and S velocity models between the event or converting interface and the station. Teleseismic arrivals and local events close to the station have similar steep incidence angles (Fig. 3C). Direct comparison of such S-P delay times between the two data sets therefore avoids the introduction of bias from using velocity models for depth migration and from unaccounted-for lateral variations, because the ray paths sample similar volumes. The two comparison data sets are constructed as follows.

Local Seismicity

For the local seismicity used in the S-P delay comparison, we used a previously relocated and published local event set of ~500 earthquakes (Monsalve et al., 2006; Fig. 1), each located with at least 10 P and S arrivals picked with the HIMNT network; pick statistics are shown.
Figure 1. Map of study area, Himalaya-Tibet collision zone. Inset map shows regional geographical context, with gray shading showing topography for orientation. Blue circles in inset are 12 events from the compilation of Chen and Yang (2004) with depths of 80 km or more. Magenta open circles in inset map are our deep events 1–6 (Table DR1 [see footnote 1]). In large map, background color is elevation shaded by topography. Small circles are best-relocated local seismicity from 2001 to 2003 (Monsalve et al., 2006, 2008, 2009). Large circles are the three deep events from the compilation of Chen and Yang (2004, their events T8, T9, and T12; displayed here with event year and body wave magnitude) that fall within the map area. Events are color coded by depth below sea level. Small circles have thin white outline for events in the initial catalog (Monsalve et al., 2009; white circles in Fig. 2B) and thicker black outline for events used in delay-time analysis after selection for station distance, location error, and pick uncertainty (blue circles in Fig. 2B). Black triangles are HIMNT (Himalayan NepalTibet) stations used in this study. Red line marks the profile shown in Figures 2 and 3. White squares show locations of Kathmandu, Mount Everest, and Kanchenjunga. Black labeled ellipses roughly outline clusters of deep seismicity A–C discussed in the text; while the three larger events from Chen and Yang (2004) are located near them, the clusters’ distribution in space, time, and magnitude do not fit aftershock characteristics (see the Data Repository). Dashed rectangle shows map area of Figure 4A, and pale orange squares show receiver function piercing points at 75 km depth for stations in Figure 4.

Structure from Receiver Functions

We used a previously published teleseismic Ps receiver function set from the HIMNT network (Schulte-Pelkm et al., 2005) to determine S-P times for the comparison to local seismicity. We additionally processed receiver functions to obtain a new higher-frequency set (Fig. 4) using automated event and receiver function quality-control criteria detailed by Schulte-Pelkm and Mahan (2014). Arrival times for the Moho conversion were picked from the original receiver function set for each station after corrections for slowness, with errors based on the width of the arrival and azimuthal variations seen at the station (Fig. 2C). A comparison with the second higher-frequency receiver function set shows close agreement in picked Moho times (examples in Fig. 4B). A group of stations in southern Tibet shows a positive amplitude arrival preceding the Moho peak (Fig. 4B), which marks the top of a layer with elevated lower-crustal velocities seen previously under this as well as other networks along the Himalaya (Yuan et al., 1997; Kind et al., 2002; Wittlinger et al., 2009; Schulte-Pelkm et al., 2005; Nábělek et al., 2009; Zhang et al., 2014). As in these studies, we picked the later arrival as the Moho at these stations because the resulting Moho is contiguous with that determined at neighboring stations. We also picked Main Himalayan thrust arrival times on the peak amplitude of the first azimuthal harmonic at each station (Schulte-Pelkm and Mahan, 2014).

Distance Correction

The events selected as above as well as the teleseismic receiver functions have steep ray paths under the stations and therefore sample similar structure. The S-P delay time is a function of the incidence angle. We applied a moveout correction to vertical incidence to the receiver function waveforms’ time axis prior to picking the interface S-P times (Schulte-Pelkm et al., 2005), and applied a correction to the local event S-P time to mimic vertical incidence. The correction requires assumption of average crustal velocities between the event or interface in question and the station:

\[
\Delta t = \sqrt{\Delta t^2 - d^2 \left( \frac{1}{V_p} - \frac{1}{V_s} \right)^2}.
\]

(where \(\Delta t\) is the vertical time difference, \(\Delta t^2\) is the measured time difference, \(d\) is the epicentral distance, and \(V_p\) and \(V_s\) are the P and S wave speeds, respectively; so our S-P time comparison is not completely velocity free. However, Figure 3 demonstrates that the error introduced by a possible bias in velocity model is of second order because of the small epicentral distances involved, whereas a velocity model bias leads to a first-order error in depth determination.

After correcting all data to vertical incidence, we plotted the vertical S-P delay times from local seismicity and from teleseismic receiver functions on the same range-perpendicular profile (Fig. 2C). The profile is the same as the distance-depth profile in Figure 2B, but the vertical axis is now S-P delay time instead of depth. This representation introduces a distortion of the depth scale but minimizes the effect of the unknown velocity model (Fig. 3) and shows the relative depth relationship between the Moho and earthquake hypocenters. To test for any influence of lateral variations in Moho depth, we compare individual deep events to nearby receiver-function Moho piercing points in Figure 4.
are not the deepest in the depth profile (Fig. 2B). An additional event just south of station RBSH plots within RBBS’s Moho error bar on the S-P profile (Fig. 2C, near −10 km profile distance). The magnitudes of these six deep events range from M, 2.4 to 2.9 (Table DR1).

The larger error bar on the Moho time picked at station RBSH is due to azimuthal variations of the Moho pick in the receiver functions from that station. To exclude the influence of lateral variations in Moho depths or in velocities along ray paths, we compared deep-event S-P times to those from receiver functions with the closest piercing points. Figure 4A shows the map locations of the five sub-Moho and one near-Moho events in Figure 2C, as well as receiver function piercing points at 75 km depth for nearby stations. Figure 4B compares the S-P delay times of those events to receiver functions from nearby piercing points (larger dots in Fig. 4A). Receiver function stacks for those piercing points are shown for the same frequency band as used in the cross section in Figure 2B as well as for a higher-frequency band to obtain sharper Moho resolution. The shallower event south of RBBS (event 6) is below even shallower Moho picks for nearby piercing points. The considerable difference in Moho pick times between receiver functions from southern compared to northern back-azimuths at RBBS may represent actual Moho topography, but may also be due to lateral variations in velocity structure above the Moho.

The six deep events’ delay times are 0.3–1.1 s larger than the Moho delay times (Fig. 4B). Using uppermost mantle velocities from the standard AK135 model (Kennett et al., 1995) as well as those for Tibet from Monsalve et al. (2006), this delay time range corresponds to event depths of 3–11 km below the Moho. None of the deep events in southern Tibet that pass the error and distance criteria plot above the Moho. In an S-P profile with more events including those with larger location errors and pick uncertainties (Fig. DR3), only one event plots above the error bar of the Moho in southern Tibet despite more scatter overall.

Shallower seismicity is concentrated in the southern Tibetan upper crust, with one event just above the Main Himalayan thrust (Fig. 2C, profile distances 0–150 km). No events are seen in the Indian crust (below the Main Himalayan thrust and above the Moho) under southern Tibet.
from ~50 km profile distance to the north. This bimodal depth distribution in seismicity and the concentration of deep seismicity in cluster A persists from the 2001–2003 recording period of the HIMNT network (Monsalve et al., 2006; Huang et al., 2009; Carpenter, 2010) may also color-coded to show which station has an S-to-P delay (S-P) pick fulfilling distance and error criteria (event 1 has picks at two stations). Moho picks and errors in Figure 2C are based on receiver functions from all piercing points (small dots). Triangles show Moho peak in receiver functions for lower (red) and higher (black) frequency band. Question marks are arrivals that may stem from top of high-velocity lower crustal layer identified in other studies cited in text.

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