The crustal structure from the Altai Mountains to the Altyn Tagh fault, northwest China

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[1] We present a new crustal section across northwest China based on a seismic refraction profile and geologic mapping. The 1100-km-long section crosses the southern margin of the Chinese Altai Mountains, Junggar Accretional Belt and eastern Junggar basin, easternmost Tianshan Mountains, and easternmost Tarim basin. The crustal velocity structure and Poisson’s ratio (\(\sigma\)), which provide a constraint on crustal composition, were determined from \(P\) and \(S\) wave data. Despite the complex geology, the crustal thickness along the entire profile is nearly uniform at 50 km. The thickest crust (56 km) occurs at the northern end of the profile beneath the Altai Mountains and the thinnest (46 km) crust is beneath the Junggar basin. Beneath surficial sediments, the crust is underlain by a 15–30 km thick high velocity (6.9–7.0 km/s; indicative of an intermediate crustal composition. The entire 1100-km-long profile is divided into 110-km-long segments, and the crustal velocity varies significantly along the profile. The northern half of the profile below the Altai Mountains and Junggar Accretional Belt has a higher Poisson’s ratio of \(\sigma = 0.26–0.27\) to a depth of 30 km, which suggests a quartz-rich, granitic upper crustal composition. The northern half of the profile is the thickest crust (56 km) and occurs at the northern end of the profile beneath the Altai Mountains and the thinnest (46 km) crust is beneath the Junggar basin. Beneath surficial sediments, the crust is found to have three layers with \(P\) wave velocities (\(V_p\)) of 6.0–6.3, 6.3–6.6, and 6.9–7.0 km/s, respectively. The southern half of the profile, including the eastern Tianshan Mountains and eastern margin of the Tarim basin, shows low \(P\) wave velocities and \(\sigma = 0.25\) to a depth of 30 km, which suggests a quartz-rich, granitic upper crustal composition. The northern half of the profile below the Altai Mountains and Junggar Accretional Belt has a higher Poisson’s ratio of \(\sigma = 0.26–0.27\) to a depth of 30 km, indicative of an intermediate crustal composition. The entire 1100-km-long profile is underlain by a 15–30 km thick high velocity (6.9–7.0 km/s; \(\sigma = 0.26–0.28\)) lower-crustal layer that we interpret to have a bulk composition of mafic granulite. At the southern end of the profile, a 5-km-thick midcrustal low-velocity layer (\(V_p = 5.9\) km/s, \(\sigma = 0.25\)) underlies the Tianshan and the region to the south, and may be indicative of a near-horizontal detachment interface. \(P_n\) velocities are \(~7.7–7.8\) km/s between the Tianshan and the Junggar basin, and \(~7.9–8.0\) km/s below the Altai Mountains and eastern margin of the Tarim basin. We interpret the consistent three-layer stratification of the crust to indicate that the crust has undergone partial melting and differentiation after Paleozoic terrane accretion. The thickness (50 km) of the crust appears to be related to compression resulting from the Indo-Asian collision.

INDEX TERMS: 7205 Seismology: Continental crust (1242); 7218 Seismology: Lithosphere and upper mantle; 8110 Tectonophysics: Continental tectonics—general (0905); KEYWORDS: crustal structure, China, seismic refraction, compression, Indo-Asian collision


1. Introduction

[2] Northwest China consists of a 2000-km-wide orogenic belt located between the Siberian craton and the northern edge of the Tibetan plateau. The Indo-Asian collision is the latest in a series of tectonic events that have built this impressive orogenic belt. This region offers excellent field exposures that document a wide range of geological processes that have formed and modified the Earth’s crust.

Northwest China is considered a “type locality” for studies of crustal accretion, deformation, and stabilization [Yang and Yang, 1981; Zhang et al., 1984; Ren et al., 1987; Yuan et al., 1992; Avouac and Tapponnier, 1993; Dong, 1993; Berzin et al., 1994; Rowly, 1996; Yin and Nie, 1996].

[3] As part of a geoscience transect that extended from the Altai Mountains to Taiwan, a 1100-km-long seismic refraction survey was conducted across northwest China (Figures 1 and 2) by the former Ministry of Geology and Mineral Resources (MGMR). The three-component seismic data were acquired in 1988. The goal of this project was to gain a better understanding of the deep crustal structure beneath different tectonic provinces in the area and to provide crustal information for mineral resources development [Xu and Wang, 1991; Wang, 1992; Yuan et al., 1992].

[4] Prior information regarding the seismic properties of the crust and upper mantle in northwest China have been obtained using local and teleseismic tomography [Liu et al.,...
1989; Roecker et al., 1993; Teng et al., 1992, 1994], the analysis of seismic surface waves [Feng et al., 1980; Cotton and Avouac, 1994; Mahdi and Pavlis, 1998], seismic refraction studies in the southwestern Tarim basin [Kao et al., 2001], and modeling of gravity data [Burov et al., 1990]. The majority of prior work has focused on the region of the Tianshan where a crustal thickness of 50–55 km has been reported [Roecker et al., 1993]. Prior work has not included the crustal structure of the Altai Mountains or the eastern edge of the Tarim basin, as we present here. Cotton and Avouac [1994] report low shear wave velocities in our study area, a result that is consistent with our observations. There have been geophysical studies of the Junngar basin, but these have mainly addressed shallow depths where hydrocarbon deposits are known [e.g., Song et al., 2000] and these data have not been made available.

[5] This information, combined with recent advances in processing techniques of three-component data, have enabled us to model both the $P$ and $S$ wave velocity structure ($V_p$ and $V_s$, respectively), and to infer compositional changes within the crust and uppermost mantle using Poisson’s ratio [Castagna, 1985; Holbrook et al., 1988, 1992; Eesley, 1989; Gajewski et al., 1990; Romero et al., 1993; Zandt et al., 1994; Kosminskaya, 1995; Krilov et al., 1995; Christensen and Mooney, 1995; Christensen, 1996; Catchings and Lee, 1996; Hawman, 1996; Kern et al., 1999; Swenson et al.,...
2000]. In this paper we present the $P$ and $S$ wave velocity structure of the crust from our seismic refraction data, and describe an interpretive geological crustal section that is based on these results and geologic mapping.

2. Geological Setting

[6] Southern Eurasia, of which northwestern China is a part, was formed by the coalescence and accretion of diverse crustal fragments from a series of collisional events during Paleozoic and Mesozoic times [Yang and Yang, 1981; Zhang et al., 1984; Ren et al., 1987; Zhang, 1997; Berzin et al., 1994; Carroll et al., 1995; Rowly, 1996; Yin and Nie, 1996; Neil and Houseman, 1997; Yin et al., 1998]. Most of western China, including the Altai Mountains, Junggar basin, Tianshan Mountains (henceforth simply the Tianshan, since “shan” means “mountains”), and the eastern margin of the Tarim basin (Figure 2), where assembled prior to the mid-Cenozoic Indo-Asian collision that reshaped the region and reactivated older structures [Molnar and Tapponnier, 1975; Tao and Lu, 1981; Tapponnier et al., 1982, 2001; Windley et al., 1990; Dong, 1993; Avouac and Tapponnier, 1993; Avouac et al., 1993; Liu et al., 1994; Allen et al., 1995; Yin and Nie, 1996].

[7] The tectonic terranes comprising the study area are bounded by faults and inferred sutures (Figure 2). Some of

Figure 2. Geological map of the region investigated by the seismic transect. The transect location is indicated by a solid line; circled numbers 1 through 12 indicate shot point locations. The northernmost shot point (SP1) is located at the boundary between the Altai Mountains and the Junggar Accretional Belt. The southernmost shot point (SP12) is located adjacent to the Altn Tagh fault. The seismic transect crosses, in addition, the eastern Junggar basin, Bogda Shan arc (SP5-6), the Turpan-Hami basin (SP6-9), the Tianshan (SP9-10), and the eastern Tarim basin (SP10-12).
these faults are seismically active and have been sites of major historic earthquakes, while others are aseismic and mark the sutures between Paleozoic and Mesozoic terranes. The geologic makeup and tectonic history of each separate terrane is briefly discussed from north to south.

2.1. The South Margin of the Altai Mountains and the Junggar Fold (Accretional) Belt

The Altai Mountains (north of shot point 1, Figure 2) include a 12-km-thick early Paleozoic accretionary wedge, consisting of metamorphosed sandstone, shales, and minor limestones. Intercalated within these sedimentary sequences are mid-Paleozoic island arc volcanics and calc-alkaline intrusives [Sengor et al., 1993; Qu and He, 1993; Cunningham et al., 1996a, 1996b]. Gneissic rocks along the southern border are believed to comprise the Precambrian basement [Feng et al., 1989], but the existence of older basement in the Altai is not yet resolved. Windley et al. [1994] suggest that these gneisses may represent Precambrian to early Paleozoic accreted fragments. This suggestion is supported by Berzin et al. [1994] who show the Altai Mountains as a microcontinent with possible Precambrian basement [Bibkova et al., 1992].

The Junggar Accretional Belt (shot points 1, 2, and 3; Figure 2) is bound by the Irtysh and Kelimali faults and consists mainly of accreted Devonian and Carboniferous sediments and volcanics with a few scattered Ordovician and Silurian rocks [Feng et al., 1980], and is a part of the south margin of the Altai Block. Dismembered ophiolites and associated cherts that record the involvement of oceanic crust of the Paleo-Junggar Ocean in the accretion process are concentrated along the Irtysh, Kelimali, and Almanti faults (Figure 2). The ages of these rocks indicate continued subduction and accretion from the Middle Devonian to the Early Carboniferous. Within the Altai and Junggar Mountains, Paleozoic calc-alkaline intrusions were formed above a northward-dipping oceanic slab. The Paleozoic calc-alkaline plutons predate the final Late Carboniferous to Early Permian amalgamation along the North and central Tianshan suture.

2.2. The Junggar Basin

The sediments of the Junggar basin (shot points 4 and 5; Figure 2) have been deposited on a basement similar to that of the Junggar Accretionary Belt [Feng et al., 1980; Tao and Lu, 1981; Kwon et al., 1989; Carroll et al., 1990, 1995; Allen et al., 1991, 1995]. The basin is triangular in shape and filled with deposits thickening gradually from north to south, and the largest accumulations being found along the Bogda Shan island arc (Figure 2). The basal Upper Permian deposits of the Junggar basin are nonmarine and were deposited within the subsiding foreland basin [Kwon et al., 1989; Carroll et al., 1990]. Geological reconstructions suggest that the basement underlying the Junggar basin consists of incompletely subducted, imbricated mid-Carboniferous oceanic crust and trench-wedge volcanoclastic sediment.

2.3. The Bogda Shan Volcanic Arc

The Carboniferous Bogda Shan volcanic arc, south of the Junggar basin (shot points 5 and 6; Figure 2), was formed as a consequence of northward subduction and extension during and prior to the accretion of the Paleozoic and Precambrian Tianshan blocks. The arc appears to be centered in the Bogda Shan where thick sections of Carboniferous submarine pyroclastics and basalt-andesite flows extend north and south across the present Junggar and Turpan basins [Coleman, 1989] (Figure 2). To the west of the Bogda Shan, the Carboniferous volcanics rest unconformably on an early Paleozoic accretionary complex of trench sediment and dismembered ophiolite. Evidence for a Precambrian basement underlying these Carboniferous volcanics is not seen along the northern front of the Bogda Shan [Ren et al., 1987]. Similar Permian lacustrine deposits and overlying felsic clastics in both basins suggest that an Andean-type arc did not separate the Turpan and Junggar basins at this time [Greene et al., 1997].

2.4. The Turpan-Hami Basin

The depositional sequences in the Turpan-Hami basin (shot points 7, 8, and 9; Figure 2) are similar to those in the Junggar basin and consist of basal Carboniferous andesitic volcanics, overlain by Permian lacustrine mudstones interlayered with continental fluvial and volcanic sediments. The Permian sequence is overlain by nearly 3000 m of Triassic and Jurassic foreland-style continental deposits, rich in coals and lacustrine mudstones [Carroll et al., 1995; Greene et al., 1997]. The Carboniferous volcaniclastic sequences exposed on the southern margins of the Turpan-Hami basin are intruded by late Paleozoic calc-alkaline granites related to the Bogda Shan arc.

2.5. The Tianshan

Located south of the Kushi fault (shot point 10; Figure 2) (also called the South Tianshan fault or Nicolaev Line), the Tianshan has been tightly compressed during Mesozoic and Cenozoic time. The central Tianshan consists of Proterozoic gneiss and schist intruded by Paleozoic calc-alkaline granites [Allen et al., 1992, 1993; Windley et al., 1990]. These older rocks are overlain by passive margin carbonate and continental clastics of Middle and Upper Proterozoic age [Ren et al., 1987]. The rock formations of the South Tianshan merge with the passive margin of the Tarim Platform. A late Paleozoic subduction scenario between the Tarim and central Tianshan has been suggested by Allen et al. [1992] based on fragments of ophiolite bodies found within the South Tianshan fault zone. Further east, the central Tianshan wedges out as a result of dextral strike-slip faulting and crustal subduction of the central Tianshan block under the north-facing passive margin of the Tarim Platform (Figure 1). Finally, the Cenozoic shortening of this area is considered to result from the collision of the Indian Craton to the south [Carroll et al., 1995; Windley et al., 1994; Hendrix et al., 1994]. The Cenozoic imbrication, and thickening of the crustal section south of the Bogda Shan, was accomplished by thrusting and strike-slip faulting at the surface, submerging the central Tianshan under the Tarim passive margin (Figure 2). Present-day strike-slip motion along the Altyn Tagh (Figure 2) and thrusting in the area [Zhou and Graham, 1996] has, together with convergence, produced an unusually thick (50 km) continental crust beneath the seismic transect at this location.
2.6. The Tarim Basin

The Tarim basin (shot points 11 and 12; Figure 2) is filled with a thick (∼6 km along our profile) foreland sequence of Mesozoic and Cenozoic continental sediments. These are underlain by Paleozoic shallow marine sediments, and Late Proterozoic rocks which floor the entire basin [Jia et al., 1991; Allen et al., 1991; Sobel, 1999]. The Altyn Tagh fault system has exposed possible Archean basement on the southern boundary of this craton. These inferred older rocks are intruded by early Paleozoic calc-alkaline granite (Figure 2).

2.7. The Altyn Tagh Fault and the Qaidam Depression

The Altyn Tagh fault zone (shot point 12; Figure 2) truncates the E-W trending Qaidam units, juxtaposing them against the mildly deformed Tarim Platform sequences [Zhou and Graham, 1996; Ritts and Biffi, 2000]. The Qaidam Depression is filled with Mesozoic to Cenozoic basinial sequences of lacustrine and fluvial fan deposits [Carroll et al., 1990; Ritts, 1995]. The basement consists of a late Paleozoic fold belt containing active continental-margin sediments and volcanics. The entire sequence is strongly folded and is related to the E-W trending Kunlun suture which shows evidence of Paleozoic convergence with its high-pressure/low-temperature metamorphic rocks [Chen et al., 1999].

3. Seismic Data

3.1. Acquisition and Processing

Seismic energy was provided by 12 chemical explosive shots fired in boreholes. The charge size ranged from 1500 to 4000 kg, sufficient to provide clear first arrivals to a maximum distance of 300 km. The distance between shot points ranged from 63 to 125 km, and the interval between portable seismographs was between 2 and 4 km. The profile was recorded along existing roads and provided four nearly straight profile segments (Figure 2).

To assist in phase correlation, record sections with reduction velocities of 6.0 and 3.46 km/s were used for P and S waves, respectively. The timescale used for S wave record sections was compressed by a factor of 0.58 in order...
to match the $P$ wave arrival times. To avoid the small time
shift introduced by digital filters, unfiltered $P$ wave data
were used for phase correlation and travel time picking.
In order to improve the signal-to-noise ratio for phase corre-
lation, the $S$ wave data were filtered with a 0–8 Hz band
pass.

3.2. Correlation of Phases

Standard nomenclature has been used to identify first
and secondary arrival phases. The upper crustal refraction
commonly referred to as $P_g$ and $S_g$ actually corresponds to
two travel time segments (Figures 3 and 4). The first,
usually a short travel time branch that propagates very near
the shot point, is a diving wave within the sedimentary
layer, whereas the second and much longer travel time
branch is the refraction (or diving wave) from the top of
basement. True first-arrival refractions from the middle or
lower crust were rarely observed. Much more apparent were
wide-angle reflections, that are labeled $P_1P$ and $S_1S$, $P_2P$
and $S_2S$, and $P_1P$ and $S_1S$, corresponding to reflections from
the top of the middle crust, lower crust, and low-velocity
layer (where present) for $P$ and $S$ waves, respectively. The
phase $P_mP$ ($S_mS$) is the reflection from the Moho, and $P_n$
($S_n$) is the refraction from the Moho.

[18] In general, $P_g$ is a very clear phase that contains
detailed information about the upper crustal velocity struc-
ture. The $S_g$ phase is not always as clear as $P_g$ because of a
lower signal-to-noise ratio. The first segment of $P_g$ ($S_g$)
corresponds to waves propagating in the sediments or
weathered layer, and is the first arrival within a distance
of 50 km of the shot points. $P$ wave velocities observed
within the three basins that are crossed (Figure 2) indicate a
higher velocity gradient than is measured within the crys-
talline crust. The apparent velocities for $P_g$ and $S_g$ from shot
points SP4 (Figure 4) and SP7 (Figure 5), which are located
in the Junggar and Turpan basins, respectively, are relatively
low (<5.00 km/s for $P$ wave and <2.60 km/s for $S$ wave)
corresponding to thick sediments. Where sediments are
absent, the velocities of $P_g$ and $S_g$ are about 6.0 and
3.46 km/s, respectively (Figures 3 and 6).

[20] The phase that arrives after $P_g$ ($S_g$) is the wide-angle
reflection $P_1P$ ($S_1S$) from the top of the middle crust (6.3–
6.6 km/s). At certain distances, $P_1P$ seems to be a first
arrival because the $P_g$ energy is too weak to be detected.

Figure 4. Record sections of shot point SP4. (a) Trace-normalized band-pass filtered (0–8 Hz) $S$ wave
record section (transverse component) with a reduction velocity of 3.46 km/s and a factor of 0.58 in
timescale with respect to $P$ wave record section; (b) trace-normalized $P$ wave record section with a
reduction velocity of 6.00 km/s. The wide-angle reflection from the Moho ($P_mP$) and the refracted arrival
from below the Moho ($P_n$) are very clear on the $P$ wave record section.
The phase $P_2 P (S_2 S)$ corresponds to a reflection from the top of the lower crust (6.8–7.0 km/s), and is visible on many of the record sections. $P_2 P (S_2 S)$ intersects $P_m P (S_m S)$ (the Moho reflection) at a distance of 200–250 km from the shot. Despite the dominance of the $P_m P (S_m S)$ phase, the $P_2 P (S_2 S)$ phase can often be clearly identified.

To the south of shot point SP8 (Figure 2), another phase, referred to as $PLP (SLS)$ (Figure 6), can be identified just above the signal-to-noise level in the record sections between $P_1 P (S_1 S)$ and $P_2 P (S_2 S)$. Despite its weak energy, this phase is visible in the trace-normalized record section when compared to the very strong $P_m P (S_m S)$ phase (Figure 6). $P_1 P (S_1 S)$ nearly parallels $P_1 P (S_1 S)$, but has a lower average velocity. It is the low-velocity zone (LVZ) that gives rise to this phase that delays the Moho reflections in the related regions (Figures 5 and 6). No obvious time delay appears for the phases before $P_2 P (S_2 S)$, and therefore the phase $P_1 P (S_1 S)$ is interpreted to be a reflection from the top of a low-velocity layer beneath the southern section of the profile (Tianshan, Turpan-Hami basin, and eastern Tarim basin).

The Moho reflection $P_m P (S_m S)$ is the dominant phase on all record sections at distances greater than 100–150 km. Due to the existence of a complex crustal structure, especially the midcrustal low-velocity layer below the southern part of the profile, the phase $P_m P (S_m S)$ sometimes exists in two segments with a clear time delay separating them. Examples of this phenomenon can be found in the record sections from shot points SP7 (Figure 5) and SP10 (Figure 6). In the record section of SP7, the arrivals of $P_m P (S_m S)$ reflections arrive later at a distance greater than 170 km. Likewise for SP10, the travel times of $P_m P$ and $S_m S$ are delayed at distances less than about 200 km from the shot in the northern segment. These results indicate that the northern limit of the deep crustal low-velocity layer must be located somewhere between SP8 and SP9 along the profile near the northern edge of the Tianshan (Figure 2).

$P_n$ was observed as the first arrival in several record sections at distances greater than 200 km (Figures 4–6). The apparent velocity of $P_n$ ranges from 7.6 to 7.9 km/s. The lowest values occur beneath the Junggar and Turpan-Hami basins, but $P_n$ has a high apparent velocity beneath the Altai Mountains, Junggar Accretional Belt, Tianshan Accretional Belt, and eastern margin of the Tarim platform. $S_n$ is also visible in some $S$ wave sections. Forward
modeling of $P_n$ observations was used to determine the velocity structure of the uppermost mantle.

4. Modeling the Data

[24] Using the abovementioned phase correlations, the first arrivals of $P_g$ were used to invert for the upper crustal velocity structure using a finite difference tomographic method [Kelly et al., 1976; Hole, 1992; Ohminato and Chouet, 1997]. The reflection phases $P_1P$, $P_2P$, and $P_{m}P$ were used to determine the approximate velocity structure of the middle to lower crust using a combination of the $x^2-\tau^2$ method [Giese et al., 1976] and the reflectivity method [Fuchs and Müller, 1971]. After establishing an initial crustal $P$ wave velocity structure, the final $P$ wave model was determined using 2-D forward raytracing [Cerveny et al., 1977; Cerveny and Psenick, 1984]. Travel times and amplitudes of the various phases on the record sections were fitted by adjusting the velocities and depths of the boundaries with the raytracing method (Figures 7 and 8), thus establishing the final $P$ wave velocity model. The ray method that we have used to model the data is biased toward relatively flat-laying structures with discrete seismic discontinuities. We admit the possibility of alternate seismic models with gradational velocity boundaries rather than first-order discontinuities. However, the main conclusions of this paper would not change significantly if such models were obtained.

[25] In recent years, there has been an increased interest in modeling both the $P$ and $S$ wave velocity structure of the crust because this combination provides better constraints on composition [e.g., Castagna, 1985; Holbrook et al., 1988, 1992; Eesley, 1989; Gajewski et al., 1990; Kosminskaya, 1995; Rudnick and Fountain, 1995; Zandt and Ammon, 1995; Krilov et al., 1995; Hauksson and Haae, 1997; Musacchio et al., 1997; Eberhart-Phillips and Michael, 1998; Menke et al., 1998; Stoerzel and Simithson, 1998; Catchings, 1999]. We have modeled the $S$ wave data by assuming the depths to all boundaries are those obtained from the $P$ waves. We were able to fit the travel times for the crystalline upper crust by replacing the $P$ wave velocities with $S$ wave velocities assuming a Poisson’s ratio of 0.25 in the initial $S$ wave model. Two-dimensional forward modeling was then used to fit the $S$ wave reflections and create a complete $S$ wave model for the crust.

[26] Uncertainties in the final velocity model primarily depend on the correct identification of the various phases and the density of rays, the shot point interval, the receiver...
density, and the degree of lateral variations in the thickness of surficial sediments. Our result has shown that, depending on the uniformity of the structure and the density of the rays, the ability to resolve velocity and depths to the interfaces are 2 and 5%, respectively (Table 1). Thus a velocity of 7.0 km/s has an uncertainty of ±0.14 km/s and a layer depth of 30 km has an uncertainty of ±1.5 km. The $V_p/V_s$ ratio has an uncertainty of 4%. Figure 9 shows a three-dimensional perspective of the final structural model along the seismic refraction profile. $P$ wave velocity and Poisson’s ratio are indicated in this figure.

### 4.1. Crustal Seismic Velocity Structure

The crustal model (Figures 9 and 10, and Table 1) shows many significant features. As expected, the Junggar, Turpan-Hami, and Tarim sedimentary basins have low velocities and steep velocity gradients with $V_p = 3.6–5.8$ km/s and $V_s = 2.3–3.3$ km/s. The Turpan-Hami basin has the greatest thickness (10 km) of sediments. Beneath the eastern Tarim basin, the velocity increases from 4.4 km/s near the surface ($V_s = 2.7$ km/s) to 5.8 km/s ($V_s = 3.3$ km/s) at a depth of ~6 km. The strong velocity contrast between the basin fill and the crystalline basement clearly outlines the 2-D configuration of the sedimentary basins. Along most of the profile, the velocity of the crystalline upper crust is 6.0–6.1 km/s at depths 8–14 km. The Bogda Shan, however, is underlain by a higher-velocity block ($V_p = 6.30$ km/s, $V_s = 3.7$ km/s) at the same depth range.

Figure 7. Observed and synthetic record sections of $P$ wave and ray coverage for shot SP7. The wide-angle reflection from the Moho is very clear, as is the refracted arrival from below the Moho ($P_n$).
The LVZ causes clearly observed time delays of the Moho reflections recorded south of Tianshan (Figures 5–8).

The entire seismic profile is underlain by a remarkably uniform high-velocity (6.9–7.0 km/s) lower crust whose top is at a depth of 25–30 km. The lower crust has an average thickness of about 20 km. The maximum lower crustal thickness (30 km) is found to the north, beneath the Altai Mountains, where the Moho is at a maximum depth of 55 km (Figure 9). The thickening of the lower crust is accommodated by both deepening of the Moho and thinning (isostatic uplift) of the upper crust. The thinnest (15 km) portion of the lower crust underlies the Junggar basin, where the Moho depth is 46 km. South of the Junggar basin, the Moho depth increases to about 50 km and then deepens abruptly (to about 68 km) to the south of the Altyn Tagh fault [Wang et al., 1997].

The seismic velocity of the uppermost mantle (Pn) is less than or equal to the global average (8.09 km/s) in all parts of the study area. The lowest values (7.7–7.8 km/s) occur beneath the Junggar and Turpan-Hami basins. Beneath the Altai Mountains, Junggar Accretional Belt, Tianshan Accretional Belt, and eastern margin of the Tarim platform, Pn velocity is 8.0 km/s.

4.2. Poisson’s Ratio

Poisson’s ratio shows the largest lateral variation within the uppermost crust (Figure 9). Poisson’s ratio increases with depth in the upper crust, especially in the Mesozoic and Cenozoic basins, such as the Junggar, Turpan-Hami, and Dunhuang basins (Figure 2). We note that our measured velocities can also be affected by crustal temperatures. However, our seismic profile passes through a region of generally low surface heat flow [40–45 mW/m²;
For temperatures more than 10% below the solidus, thermal effects act approximately equally for P wave and S wave velocities [Christensen, 1996]. Therefore we do not expect that temperature will significantly alter the measured Poisson’s ratio in the upper and middle crust. The observed $P_n$ velocity along the central portion of our profile is a low 7.7 km/s, nearly 0.4 km/s lower than the global average $P_n$ velocity (8.09 km/s). If this low $P_n$ velocity is due to a warm uppermost mantle, and not seismic anisotropy, we would expect that thermal effects would reduce Poisson’s ratio in the uppermost mantle and lower crust. While we have not estimated Poisson’s ratio in the uppermost mantle, we do observe a lower Poisson’s ratio (0.26) in the central portion of the lower crust. Error estimates for Poisson’s ratios are presented in Table 1.

The Mesozoic and Cenozoic sedimentary basins have low values of Poisson’s ratio (0.19–0.22), with the lowest value in the Turpan-Hami basin. Poisson’s ratio in the upper crust is about 0.25 in most areas, except beneath the Bogda Shan, where the value is about 0.26.

On the basis of Poisson’s ratio, the middle crust along the transect can be divided at the Bogda Shan into a northern and southern segment. In the northern segment, including the Bogda Shan, the middle crust shows a
Poisson’s ratio as high as 0.27. However, the middle crust to the south of the Bogda Shan has a more typical Poisson’s ratio of around 0.25. Poisson’s ratio in the lower crust is 0.26–0.28, with the highest value (0.28) beneath the Altai block, and the lowest value (0.26) beneath the Turpan-Hami basin. As discussed below, the profile segment with higher values of Poisson’s ratio correlates with Nd, Sr, and Pb isotopic anomalies determined on the granitic plutons [Hopson et al., 2003] indicating crust sources of oceanic (mafic) origin beneath the Junggar, Bogda Shan, and Turpan-Hami blocks. Isotopic ratios indicate that more felsic continental crust is present beneath the central and South Tianshan [Hopson et al., 2003].

4.3. Geological Interpretation

We have constructed a geologic cross section from the surface to the Moho along the seismic profile (Figure 10). The cross section uses information on surface geology from Tao and Lu [1981], Coleman [1989], Yuan et al. [1992], Teng et al. [1994], and Allen and Vincent [1999]. We describe this 1100-km-long cross section from north (Altai Mountains) to south (Altyn Tagh fault).

[35] The greatest total crustal thickness (55 km) and lower-crustal thickness (30 km) are found north of the Kelimali fault in the Altai Mountains and Junggar Accretional Belt where the high Poisson’s ratio (0.28) in the lower crust is consistent with a mafic lower-crustal composition. The lower crust is less than 15 km thick beneath the Junggar basin. There has been much discussion concerning the origin of the deep basement of the Junggar basin, which may be composed of either Precambrian crystalline rocks or accreted Paleozoic metasedimentary and/or imbricated mafic oceanic rocks [Yang and Yang, 1981; Tao and Lu, 1981].

[36] In the area of the Bogda Shan Arc and Turpan-Hami basin the crustal structure inferred from geology indicates a complex evolution, and the composition and origin of the
deep crust is controversial. We correlate the high velocities and lower Poisson’s ratio (0.26) with the exposed rocks that include Paleozoic ophiolitic, accretionary and arc rocks of oceanic affinity. Isotopic variations of Nd, Sr, and Pb in exposed Paleozoic granitoid plutons of this region show distinctive isotopic signatures indicative of source rocks with an oceanic affinity \cite{Hopson et al., 2003}. The crust, with these quartz-rich, postcollisional granites and abundant calc-alkaline intrusives, therefore shows evidence of homogenization and melting as a result of a late Paleozoic thermal event.

\cite{Wang et al.: Crustal Structure of the Altai Mountains} Beneath the Tianshan and eastern margin of the Tarim Craton, the relatively low Poisson’s ratio (0.25) and low \textit{P} wave velocity (6.3 km/s) of the middle crust, together with an underlying LVZ (\textasciitilde 5.9 km/s), imply the presence of abundant felsic rocks. Hence it appears that the crust in the southern part of the profile was either invaded by granitic plutons during a late Paleozoic thermal event, or that the crust has been doubled by thrusting, the LVZ being meta-sedimentary rocks at the top of the deeper crustal section. The lower crust (6.95 km/s) has similar values of seismic velocity and Poisson’s ratio as exists to the north, and may consist of mafic granulite facies rocks. It is remarkable that the Moho is flat from the Tianshan to the Tarim basin. Thus neither the late Cenozoic uplift of the Tianshan nor the Tarim Craton, the relatively low Poisson’s ratio (0.25) and lower seismic velocities of 6.0–6.3 km/s, indicative of higher densities due to the uplift of midcrustal rocks. (2) The seismic velocities within the middle-crustal layer are 6.5–6.6 km/s from the Altai Mountains to the Turpan-Hami basin, and then decrease to 6.3 km/s beneath the Tianshan and the Tarim basin (Figure 9). Thus the northern boundary of the Tianshan marks a fundamental change in crustal structure, with a more dense and mafic crust to the north and a lower-density, silicic crust to the south. (3) Beneath the Tianshan and continuing to the south, a 5-km-thick LVZ (5.9 km/s) in the middle crust is found at 25–30 km depth. This LVZ may serve as a detachment surface upon which the upper crust of the Tianshan and Tarim basin are thrust, or may be a zone of silicic igneous intrusions. (4) The high-velocity (6.9–7.0 km/s) lower-crustal layer has an average thickness of 20 km along the entire profile. It reaches a maximum thickness of 30 km beneath the southern margin of the Altai Mountains and a minimum thickness of 15 km beneath the Junggar basin. Both the higher seismic velocities throughout the crust and isotopic ratios (Nd, Sr, Pb) from the granitic plutons \cite{Hopson et al., 2003} indicate the presence of oceanic crust and mantle north of the Tianshan, i.e., beneath the Junggar, Bogda Shan, and the Turpan-Hami blocks (Figure 10), and felsic continental crust beneath the central and South Tianshan. (5) The seismic velocity of the uppermost mantle \textit{(Pn)} is a low 7.7 km/s, except beneath the Altai Mountains to the north and the Tarim basin to the south. This \textit{Pn} velocity is \textasciitilde 0.4 km/s lower than the global average (8.09 km/s) and indicated either significant seismic anisotropy or elevated temperatures in the uppermost mantle. Surface heat flow is low (40–45 mW/m²) which indicates that seismic anisotropy is the more likely explanation.

\cite{Christensen and Mooney, 1995} The three-layer stratification of the crust into well-defined upper, middle, and lower crust on this profile is similar to stable continental crust elsewhere (i.e., Precambrian platforms and shields). However, the lower-crustal layer (6.9–7.0 km/s) is remarkably thick, with an average thickness of 20 km (amounting to some 40% of the total crustal thickness). The middle crust (6.3–6.6 km/s) has an average thickness of 17 km, but the upper (crystalline) crust has an average thickness of only 9 km. Sedimentary accumulations range from 0 to 10 km in thickness. The average thickness of the entire crust along the profile is 50 km, and the middle and lower crust comprise 74% of the crust.

5. Discussion and Conclusion
5.1. Crustal Structure

\cite{Wang et al.: Crustal Structure of the Altai Mountains} The depth of each of the sedimentary basins along the transect is well determined from first arrivals in the seismic data. We find that the Junggar, Turpan-Hami, and Tarim basins each have 5–10 km of sedimentary fill (Figure 10). However, we do not find any crustal thinning (i.e., Moho uplifts) beneath the sedimentary basins, as is often found elsewhere. This significant observation implies that the sedimentary basins are not the result of local crustal extension, but rather result from a combination of compressional and especially horizontal (strike-slip) deformation.

\cite{Wang et al.: Crustal Structure of the Altai Mountains} The crystalline crust consists of an upper, middle, and lower crust with seismic velocities of 6.0–6.3, 6.3–6.6, and 6.9–7.0 km/s, respectively. Several important features are evident along the transect. (1) The Bogda Shan arc shows the highest upper crustal velocities (6.3 km/s), indicative of higher densities due to the uplift of midcrustal rocks. (2) The seismic velocities within the middle-crustal layer are 6.5–6.6 km/s from the Altai Mountains to the Turpan-Hami basin, and then decrease to 6.3 km/s beneath the Tianshan and the Tarim basin (Figure 9). Thus the northern boundary of the Tianshan marks a fundamental change in crustal structure, with a more dense and mafic crust to the north and a lower-density, silicic crust to the south. (3) Beneath the Tianshan and continuing to the south, a 5-km-thick LVZ (5.9 km/s) in the middle crust is found at 25–30 km depth. This LVZ may serve as a detachment surface upon which the upper crust of the Tianshan and Tarim basin are thrust, or may be a zone of silicic igneous intrusions. (4) The high-velocity (6.9–7.0 km/s) lower-crustal layer has an average thickness of 20 km along the entire profile. It reaches a maximum thickness of 30 km beneath the southern margin of the Altai Mountains and a minimum thickness of 15 km beneath the Junggar basin. Both the higher seismic velocities throughout the crust and isotopic ratios (Nd, Sr, Pb) from the granitic plutons \cite{Hopson et al., 2003} indicate the presence of oceanic crust and mantle north of the Tianshan, i.e., beneath the Junggar, Bogda Shan, and the Turpan-Hami blocks (Figure 10), and felsic continental crust beneath the central and South Tianshan. (5) The seismic velocity of the uppermost mantle \textit{(Pn)} is a low 7.7 km/s, except beneath the Altai Mountains to the north and the Tarim basin to the south. This \textit{Pn} velocity is \textasciitilde 0.4 km/s lower than the global average (8.09 km/s) and indicated either significant seismic anisotropy or elevated temperatures in the uppermost mantle. Surface heat flow is low (40–45 mW/m²) which indicates that seismic anisotropy is the more likely explanation.

\cite{Wang et al.: Crustal Structure of the Altai Mountains} The three-layer stratification of the crust into well-defined upper, middle, and lower crust on this profile is similar to stable continental crust elsewhere (i.e., Precambrian platforms and shields). However, the lower-crustal layer (6.9–7.0 km/s) is remarkably thick, with an average thickness of 20 km (amounting to some 40% of the total crustal thickness). The middle crust (6.3–6.6 km/s) has an average thickness of 17 km, but the upper (crystalline) crust has an average thickness of only 9 km. Sedimentary accumulations range from 0 to 10 km in thickness. The average thickness of the entire crust along the profile is 50 km, and the middle and lower crust comprise 74% of the crust.

5.2. Crustal Thickening

\cite{Wang et al.: Crustal Structure of the Altai Mountains} The crustal thickness (\textasciitilde 50 km) found in this study is 11 km thicker than the global average for continental crust \cite{Christensen and Mooney, 1995}. Crustal thickening is likely to be the result of compressional shortening associated with the Tibetan orogeny and/or magmatic additions. In order to thicken a 39-km-thick crust to a net thickness of 50 km, 30% shortening is required. However, the fact that the upper crust (6.0–6.3 km/s) has only
approximately half the thickness of the middle and lower crust suggests that isostatic uplift has caused erosion of about 10 km of the (shortened) upper crust. If we tentatively accept that 10 km of crust has been removed, then the shortening along the profile amounts to 50%. A major uncertainty in this calculation is the initial crustal thickness.

[45] Alternatively, the crust may have been thickened by magmatic additions. If magmatism was mainly of intermediate to mafic composition, thickening will have occurred mainly in the middle and lower crust. Silicic intrusions may have thickened the upper crust, which was subsequently thinned by erosion processes. The present geophysical data do not allow us to discern which process, compression or magmatic addition, has dominated the mechanism of crustal thickening.

5.3. Crustal Composition

[46] The interpretation of crustal composition from seismic velocity measurements has recently been reviewed by Holbrook et al. [1992], Rudnick and Fountain [1995], and Christensen and Mooney [1995]. We begin our discussion with the measured \( V_p \) structure, and then proceed to the \( V_p/V_s \) ratio, or equivalently, Poisson’s ratio.

[47] The average \( P \) wave velocity of the crystalline crust varies along the seismic transect, with important boundaries coinciding with: (1) the southern Junggar Accretional Belt and (2) the northern Tianshan. Beneath the Altai Mountains and the Junggar Accretional Belt (SP1 and SP2; Figure 10) the average crustal velocity is 6.7 km/s. This value, which is significantly higher than the global average (6.45 km/s), is due to two factors: (1) the moderately high seismic velocity (6.5–6.6 km/s) of the middle crust and (2) the unusually large thickness (30 km) of the high-velocity (7.0 km/s) lower crust (Figure 10). Between the Junggar basin and the northern flank of the Tianshan (SP3 through SP8; Figure 10) the average crustal velocity is 6.5 km/s, a value that is essentially equal to the global average. From the Tianshan to the Altyn Tagh fault, the average crustal velocity is 6.3 km/s, a value that is 0.15 km/s lower than the global average and is significantly lower (0.2–0.4 km/s) than the portion of the transect north of the Tianshan. Thus three crustal types may be clearly distinguished along this transect on the basis of average crustal velocities of 6.3, 6.5, and 6.7 km/s. The origin of these crustal types may be attributed to: (1) diverse accreted terranes (magmatic arcs, oceanic crust, continental fragments) that form the crust and (2) different amounts of crustal shortening and postaccretional magmatic additions that have modified the crust.

[48] Christensen and Mooney [1995] present statistical averages of seismic velocity as a function of depth for continental crust, and relate seismic velocity to rock types. Upper crustal velocities in the study area are consistent with geologically observed metasedimentary and felsic intrusive rocks. Measured Poisson’s ratio for these rocks (0.25) are consistent with this interpretation. As noted above, seismic velocities within the middle crust are lower beneath the Tianshan and eastern Tarim basin (\( V_p = 6.3 \) km/s) than under the terranes to the north (\( V_p = 6.5–6.6 \) km/s), and Poisson’s ratio increases slightly from south (0.25) to north (0.26–0.27). These observations indicate that the bulk composition of middle crust beneath the Tianshan is close to that of a tonalite or granodiorite, and becomes more mafic to the north (i.e., equivalent to a diorite). For example, an increase in the abundance of amphibolite in the middle crust from 5% beneath the Tianshan to about 35% beneath the terranes to the north would satisfy the observations [cf., Christensen and Mooney, 1995]. The seismic velocity (6.9–7.0 km/s) and Poisson’s ratio (0.26–0.28) of the lower crust are remarkably uniform along the entire 1100 km transect. The seismic measurements are consistent with a mafic composition (mafic granulite and/or mafic garnet-granulite), or anorhosite (less likely). Hopson et al. [2003] suggested that the lower crust consists largely of mafic migmatite in which a large mafic (amphibolite and granulite) and ultramafic (meta peridotite) component is responsible for the 6.9–7.0 km/s \( P \) wave velocity, whereas a subordinate granitoid (i.e., metadiorite, metatonalite, and leucotroandhjemite) component is responsible for a somewhat lower-than-expected Poisson’s ratio. However, these compositions could provide a Poisson’s ratio as high as 0.29–0.31, which is 0.01–0.05 higher than the measured values. Granulite grade metapelite provides the best fit to the seismic velocity data, but it seems unlikely, based on heat flow constraints and volumetric considerations, that the lower continental crust is composed of a thick (20 km) layer of metapelite over a distance of 1100 km. Thus we favor mafic granulite composition.

[49] Our crustal cross section (Figure 10), which is based on surface geology and \( P \) and \( S \) wave seismic velocities, reveals a 50-km-thick crust that is remarkable for its relatively uniform three-layer stratification. The crustal structure (layer velocities, thickness, and Poisson’s ratio) of the central portion of the profile (i.e., from the Turpan-Hami basin to the Junggar basin) is similar to that of Precambrian platforms and shields [Rudnick and Fountain, 1995; Christensen and Mooney, 1995]. We therefore conclude that the accretionary process combined with compression and ~50% crustal shortening, with some magmatic additions to the base of the crust, is a valid description of the formation and evolution of stable continental crust. The outstanding major question in this study area is the composition and configuration of the underlying mantle lithosphere. Such information can be obtained from geochemical studies and seismological investigations that probe to sub-Moho depths.

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