

# Lithospheric temperature estimates from seismic attenuation across range fronts in southern and central Eurasia

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## ABSTRACT

**Seismic attenuation ( $1/Q$ ) changes abruptly between the stable Eurasian craton and adjacent mountain belts associated with the Alpine-Himalayan collision. New attenuation measurements on local body and coda waves provide quantitative constraints on this contrast, near the Caucasus, Kopet Dagh, and Tien Shan mountains. The  $1/Q$  estimates from all range fronts are similar, the mountains being two to three times more attenuative than adjacent stable platforms. These  $1/Q$  variations are a reasonable proxy for temperature variations, and reveal crustal heating on the order of 100 °C but probably not melting beneath the mountain ranges. Attenuation ( $1/Q$ ) is high wherever high topography is seen, implying that active mountains are typically hot. Hence, processes that generate or transport heat dominate advective cooling due to lithospheric thickening, even over the 10–20 m.y. lifetimes of these young mountain belts.**

## INTRODUCTION

In continental settings, plate convergence is accommodated by thickening of the lithosphere. Although thickening should result in downward advection of isotherms, the crust in many continental collisional belts appears to be hot. Given sufficient time, temperatures in thickened crust will rise directly through diffusive reheating, but other mechanisms may be required to explain the extreme heating seen in some cases (e.g., Houseman et al., 1981; Kincaid and Silver, 1996; Huerta et al., 1998). Most of these theories of crustal reheating have been tested against observations in only a few remarkable settings such as Tibet, where crustal melting may be widespread (Nelson et al., 1996). It is unclear whether such high temperatures are common in other active mountain belts, where convergence rates and crustal thickening may differ.

One approach to estimating in situ temperature has been to utilize the qualitative disappearance of seismic shear waves to indicate the presence of high temperatures or melt (e.g., Barazangi and Ni, 1982). Seismic waves attenuate in amplitude ( $A$ ) with both increasing traveltime ( $t$ ) and frequency ( $f$ ), so that at a given  $t$  amplitudes decay with frequency  $A(f) \sim \exp(-\pi ft/Q)$ , where  $1/Q$  describes the rate of attenuation in a given medium (Aki and Richards, 1980). Many physical phenomena affect  $1/Q$ , but temperature probably plays a dominant role below the upper crust. Compared to seismic wave velocities, the signal is large:  $1/Q$  varies by a factor of 2–10 between different regions in Eurasia (e.g., Mitchell et al., 1997; Sarker and Abers, 1998a). Here we document quantitative and systematic constraints on  $1/Q$  variations beneath several mountain belts in Eurasia, utilizing new high-quality seismic data from this region. By sampling both the cratons and adjacent mountain belts where basement geology is grossly similar we are able to isolate effects caused by mountain building.

## TECTONIC SETTING

Eurasia assembled in a series of amalgamation events during the Proterozoic and Paleozoic Eras (e.g., Şengör and Natal'in, 1996). Later collisions led to the Alpine-Himalayan orogenic complex, producing several synchronous collisional mountain belts of several differing tectonic styles. This study examines the northernmost of these ranges, which typically formed in the past 10–20 m.y.

The most recent Neogene convergence between the Eurasian and Arabian plates generated the Greater Caucasus and the Kopet Dagh mountain ranges at the northern margin of the collision. Convergence and uplift of these mountain belts began in the middle to late Miocene, and intensified during the middle Pliocene (e.g., Philip et al., 1989). The mountain belts are cored by the Precambrian and Paleozoic igneous and metamorphic rocks, and expose Mesozoic and younger rocks over the structurally highest regions. Mesozoic sedimentary rocks cover the southern edge of the Eurasian platform immediately to the north of both ranges, overlying Hercynian (Ordovician-Devonian) basement. Total shortening in the Greater Caucasus is estimated to be  $200 \pm 50$  km (Dotduyev, 1986), mostly since the middle Miocene. Modern convergence across the region has been measured, and is 10 mm/yr, including an unknown fraction of convergence across the Lesser Caucasus (Reilinger et al., 1997). Shortening in the Kopet Dagh is less-well known, although most of the mountain building also appears to be Neogene-Quaternary in age (Stöcklin, 1968).

Within the Tien Shan, some structures within the Eurasian Paleozoic basement have been reactivated as thrust faults (e.g., Burtman et al., 1996). Paleozoic basement in the Tien Shan is exposed over much of the structurally highest regions, which are separated by intervening late Cenozoic basins (Sadybakasov, 1990). The basement varies from the Precambrian Siberian craton in the north to a variety of accreted Paleozoic and Mesozoic terranes in the core of the mountain belt. The current shortening,  $\sim 20$  mm/yr (Abdrakhmatov et al., 1996), is believed to be accommodated by east-west-trending basement-involved faults, many of which are well south of the current range front. Assuming no change in rate, the Cenozoic shortening of  $203 \pm 50$  km in the western Tien Shan (Avouac et al., 1993) could have taken place within the past  $\sim 10$  m.y. Hence, all three mountain belts studied here formed in the past 10–20 m.y.

## DATA AND METHODS

We analyze data from two digital networks in the Greater Caucasus and Kyrgyz Tien Shan, and one broadband array in the Kopet Dagh (Fig. 1; Abers, 1994; Pavlis et al., 1994; Vernon, 1994). Attenuation is measured from seismograms corresponding to crustal paths, from spectral decay of body waves. These data are sufficient to constrain  $1/Q$  for five distinct tectonic regions (Fig. 1): the Russian platform just north of the Greater Caucasus, the western Greater Caucasus, the Kopet Dagh mountain range, the Kazakh platform, and the adjacent Kyrgyz Tien Shan.

We measure attenuation values from individual spectra of body and coda waves at frequencies where signal to noise ratios are high, typically 1–15 Hz (details in Sarker and Abers, 1998a, 1998b). Each spectrum is fit to a three-parameter model of the form:

$$H(f) = I(f)S(f)\exp(-\pi ft^*), \quad (1)$$

where  $H(f)$  is the observed amplitude spectrum,  $I(f)$  is the instrument response to ground displacement,  $t^*$  is the frequency-independent attenuation operator, and  $S(f)$  is a  $f^2$  source model parameterized by seismic moment and corner frequency, following Hough et al. (1988). Tests with other source models gave slightly different absolute values, but indistinguishable  $Q$  and temperature variations (Sarker, 1998). The  $S(f)$  term implicitly includes geometrical spreading and other frequency-independent effects; these do not change shapes of calculated spectra.

A least-squares inversion fits the model to observed log-amplitude spectra of both body and coda waves, to determine  $t^*$  and the two source

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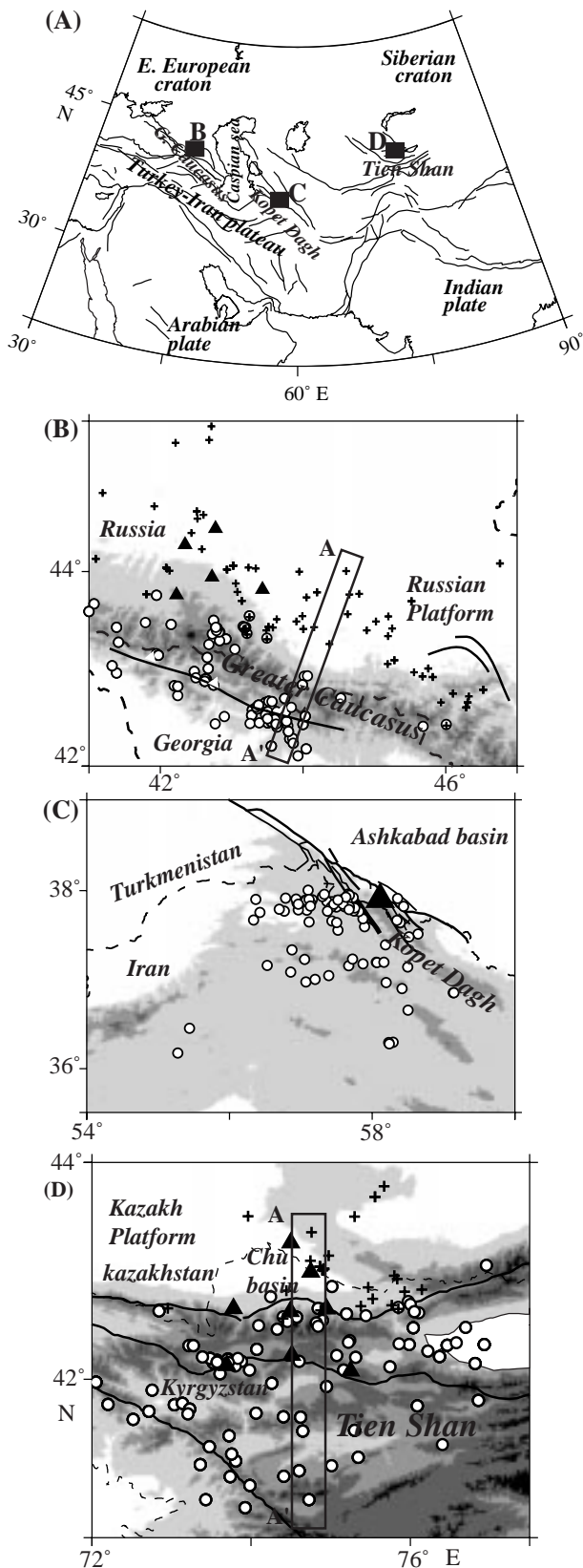


Figure 1. A: Tectonic sketch map of Eurasia are from Jackson (1992). Solid squares indicate study areas. Network stations (solid triangles) and events (circles and plus signs) used in this study are shown for (B) Greater Caucasus, (C) Kopet Dagh, and (D) northern Tien Shan. Ray paths for events are assigned to either shield (plus) or mountain (circle) region, according to where they spend substantial time. For all maps, shading becomes progressively darker with increasing elevation, white being elevations below 700 m and black being elevations higher than 4 km. Boxes indicate locations of cross sections of Figure 3.

parameters. We then determine path-averaged  $Q$  values ( $\bar{Q}$ ) for each tectonic region (ray paths are assigned to either a shield or mountain region, according to where they spend the most time), by fitting the observed increase in  $t^*$  with traveltimes ( $t$ ) to

$$t^*(t) = t_0^* + \frac{t}{Q} \quad (2)$$

(Hough et al., 1988), where  $t_0^*$  represents a near-receiver contribution, and  $t$  is calculated from local velocity models (Fig. 2). Linear error propagation on this regression provides uncertainties. We also estimate  $Q$  from coda waves using the same method to test for scattering effects.

Several sources of uncertainties in  $Q$  estimates have been tested for all three data sets. These include: (1) trade off between  $t^*$  and source corner frequencies; (2) source-depth effects; (3)  $t^*$  estimates with variable window lengths; and (4)  $Q$  variations as a function of frequency. These tests (Sarker and Abers, 1998a, 1998b) show that the added uncertainties in  $Q$  values in all cases are well within 1  $\sigma$  standard error, and never exceed 10%.

The estimates probably reflect attenuation below the upper crust. About 70% of events used in this study fall within  $1^\circ$ – $4^\circ$  of the stations. Ray tracing shows that the  $Pg$ - $Pn$  cross-over is at a distance of about 100 km, indicating that the bulk of the ray paths used in this study lie in the lower crust or in the uppermost mantle (Sarker, 1998). Because  $\bar{Q}$  is effectively determined from differences in attenuation at different  $t$  (Fig. 2), the upper-crustal legs of these paths are similar; the main difference is the extent to which each path samples greater depths. We also sample a narrow time window around the first body-wave arrivals, so that near-surface scattered waves, which have much longer traveltimes, are minimized. Hence, at these distances  $\bar{Q}$  should reflect attenuation below the upper crust. By contrast,  $t_0^*$  correlates well with near-surface geology and to velocities in the uppermost crust in the Kyrgyz Tien Shan (e.g., Ghose et al., 1998), and probably represents upper crustal attenuation (discussed more fully in Sarker and Abers, 1998a; Sarker, 1998).

## RESULTS

Estimates of attenuation parameters for  $S$  and  $P$  waves,  $\bar{Q}_S$  and  $\bar{Q}_P$ , are shown in Figure 2. Details of estimates were discussed previously for the

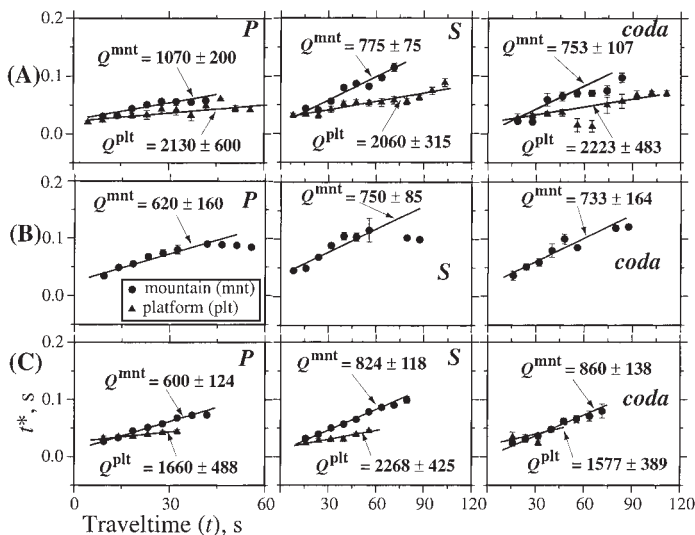


Figure 2. Frequency-independent attenuation operator ( $t^*$ ) and travel-time ( $t$ );  $t^*(t)$  is calculated from spectral fall-off of body and coda waves for frequency-independent parameterization, for (A) Caucasus network, (B) Geyokcha array, and (C) Kyrgyz network. Measurements of individual  $t^*$  estimates are averaged over roughly 5 to 10 s traveltimes window, as appropriate. Also shown are 1  $\sigma$  standard deviations of mean. Linear regressions of  $t^*(t)$  give path-averaged attenuation ( $Q$ ) for each phase ( $P$ ,  $S$ , and coda): mnt is mountain, plt is platform.

Caucasus and Kopet Dag (Sarker and Abers, 1998a, 1998b); the Kyrgyz estimates were made in a similar way although the data set is somewhat larger.

For all techniques, uncertainties are estimated to be 10%–25%, sufficiently small to distinguish  $\bar{Q}$  in mountain belts from the platforms. Specifically, we observe  $\bar{Q}_S$  of 2000–2300 in the platform areas and 700–900 in the mountain ranges; there is remarkable similarity between networks. These values are higher than some previous coda  $Q$  results (e.g., Rautian et al., 1979; Kopnichev, 1985; Mitchell et al., 1997), probably because  $Q$  varies with frequency (those studies consider predominantly 1 Hz signals, whereas our estimates are dominated by attenuation at 8–12 Hz; Sarker and Abers, 1998b). The previous studies still indicate similar relative variations between shields and mountains.

All mountain paths in central Eurasia are two to three times more attenuating than those of adjacent platforms (Fig. 2). For both sets of paths,  $\bar{Q}_S$  is roughly equal to  $\bar{Q}_P$ . Hence, all the mountain ranges from the Greater Caucasus to western Tien Shan constitute a region of high attenuation, separated by sharp boundaries from the low-attenuation platforms to the north. Tomographic inversions on  $t^*$  for  $1/Q$  from the Greater Caucasus and the western Tien Shan also reveal a sharp lateral transition between cratonic  $1/Q$  values and mountainous  $1/Q$  values (Fig. 3).

### POSSIBLE CAUSES OF $Q$

Seismic attenuation could be sensitive to different physical processes at different depths and tectonic environments. For example, in near-solidus conditions shear attenuation dominates compressional attenuation and one should observe  $Q_P/Q_S \sim 9/4$  (e.g., Anderson, 1989). By contrast, in the upper lithosphere it is often observed that  $Q_P/Q_S \sim 1$  (e.g., Frankel, 1982), consistent with lower temperature thermo-elastic mechanisms (Anderson, 1989), but also with elastic scattering caused by near-surface small-scale heterogeneity (Richards and Menke, 1983). Lateral heterogeneity of velocity structure may cause scattering, focusing and/or defocusing, and mode conversion, any of which may severely affect the amplitude of seismic waves. However, as discussed in detail elsewhere (Sarker and Abers, 1998a, 1998b; Sarker, 1998), the observations presented here are not consistent with most common scattering or focusing mechanisms. One important observation is that  $Q_{\text{coda}}$  is similar to  $\bar{Q}_S$ . Hence, little of the high-frequency energy absent from  $S$  waves enriches the late coda, as would be predicted if scattering caused the apparent attenuation (Richards and Menke, 1983). Another observation is that where

best resolved, in the Tien Shan, attenuation actually increases with depth (Fig. 3), the opposite of what is expected for crack-induced attenuation.

Besides ray tracing, the main evidence for a lower crustal, thermal origin for  $\bar{Q}$  probably lies in the geographic pattern of attenuation we observe (Fig. 2). The  $\bar{Q}$  values do not correlate with station location and subsurface geology (although  $t_0^*$  does), hence near-surface effects are unimportant. Similarly, the estimates come from sources of a variety of distances and azimuths, so near-source effects should average out. Although there is some correlation of  $\bar{Q}$  with topographic roughness, it is not exact, especially in the Tien Shan (Fig. 3). By comparison, slower phases such as  $Lg$  show attenuation that is well correlated with topography (Zhang and Lay, 1994). These comparisons and several others are extensively documented elsewhere (Sarker and Abers, 1998a, 1998b; Sarker, 1998).

By contrast, several independent lines of evidence support elevated temperatures beneath the mountain ranges, where we observe high attenuation. Active volcanoes in the Greater Caucasus attest to local melting (Philip et al., 1989). Although published heat flow data are scant in this region, high temperatures in excess of  $116 \text{ mW m}^{-2}$  are observed in the core of the Greater Caucasus relative to a mean temperature of about  $45 \text{ mW m}^{-2}$  in the adjacent Russian platform (e.g., Lubimova, 1975). These observations, together with  $Sn$  blockage in the Caucasus and Kopet Dag (Rodgers et al., 1997) and low  $Pn$  velocities (Hearn and Ni, 1994), are evidence of high sub-crustal temperatures. Strong lateral variations in mantle anisotropy in the western Tien Shan have been used as evidence for vigorous small-scale convection in the upper mantle (Makeyeva et al., 1992), also suggestive of rapid heat transfer. Although many of these arguments are weak on their own, combined with the evidence for a consistent attenuation signal they suggest that the mountains are significantly warmer than adjacent shields.

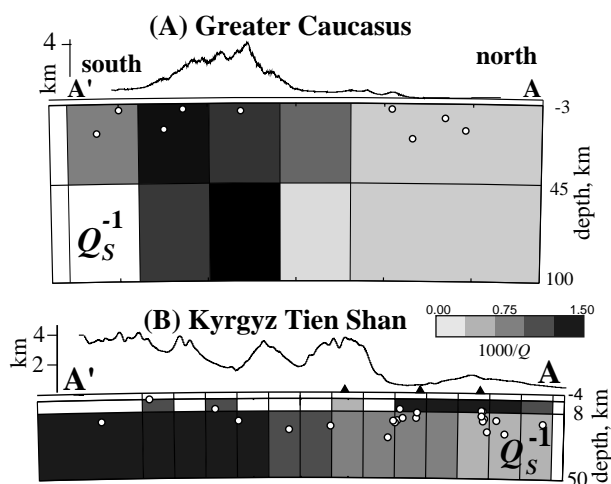
### $Q$ AND TEMPERATURE

Although low, the  $Q$  values in the mountains do not require melt and are consistent with small subsolidus temperature increases. Laboratory experiments done at high temperature and seismic frequencies (e.g., Kampfmann and Berckhmer, 1985) show that  $Q_S$  and  $Q_P$  may drop by a factor of 5–10 as temperature increases close to solidus. Hence, the factor of two to three attenuation contrasts seen along Eurasian range fronts can be easily explained by temperature variations well below the solidus and without the presence of partial melts.

The factor of two to three attenuation contrasts between platforms and mountains indicates an elevated temperature no more than  $\sim 200 \text{ }^\circ\text{C}$  and probably  $100 \text{ }^\circ\text{C}$  or less. The calibrations of Kampfmann and Berckhmer (1985) suggest temperatures of  $515\text{--}540 \text{ }^\circ\text{C}$  for the shield paths, and temperature increases of  $50\text{--}100 \text{ }^\circ\text{C}$  associated with the mountains (see Sarker and Abers, 1998a, for calculation details). These estimates were made from  $Q_S$  measurements of 2000–2300 in shields and 700–900 in mountains (Fig. 2), assuming that our results reflect  $Q_S$  at 8–12 Hz frequencies (Sarker and Abers, 1998b). Kampfmann and Berckhmer (1985) suggested a  $\pm 50 \text{ }^\circ\text{C}$  uncertainty on absolute measurements, and the high-pressure dunite experiments of Jackson et al. (1992) revealed a larger temperature dependence of  $Q$ . Both sets of experiments suggest that temperatures need not exceed  $600 \text{ }^\circ\text{C}$  to produce  $Q_S$  of 700–900 at 8–12 Hz. Hence, although these estimates are averages over some region of the crust, they do not indicate that melting conditions have been reached, even for wet felsic compositions.

### IMPLICATIONS

Our results show high  $Q$  values beneath the Eurasian cratons that are adjacent to the mountain belts of Alpine-Himalayan collision (Fig. 2). High  $Q$  values in cratonic regions are expected because they have been stable for a long time. By contrast, all the mountain belts from the Greater Caucasus to western Tien Shan constitute a region of high attenuation. Some process is heating these mountain ranges. Several mechanisms have been proposed: given sufficient time, convective instability within the mantle may elevate temperatures following lithospheric thickening (e.g., Houseman et al., 1981); viscous dissipation at high stress levels may do so over wide areas (Kincaid



**Figure 3.** Results of cross section of attenuation inversion ( $1/Q$ ) for (A) Greater Caucasus and (B) western Tien Shan (for details, see Sarker and Abers, 1998a; Sarker, 1998). Locations of cross sections are shown in Figure 1. Actual topography is shown at top of each cross section. Inversion results show high attenuation beneath mountains (dark), separated by sharp boundary from low-attenuation platforms to north (light). Triangles are stations and circles are events, projected from 20 km swath along profile.

and Silver, 1996); or rapid erosion and underplating may substantially elevate temperatures (e.g., Huerta et al., 1998). Very low upper mantle seismic velocities in the Tien Shan and Caucasus (Roecker et al., 1993; Hearn and Ni, 1994) coincide with the attenuation anomalies seen here, so high temperatures may extend too deep to be attributable to shear heating on faults.

In any case, the consistent crustal temperature anomalies inferred in this study suggest that a common mechanism is responsible for heating mountain belts having a wide range of geologic histories. This mechanism must operate over relatively short time scales, because all the observed mountain belts formed in the past 10–20 m.y. Better constraints on  $Q$ -temperature relations, as well as better imaging of variations of  $Q$  with depth and comparison with metamorphic assemblages, may allow some of the possibilities to be tested.

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