

t^* —an unsuitable parameter to characterize anelastic attenuation in the Eastern Carpathians

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ABSTRACT

Investigation of teleseismic P -wave recordings at a temporary network in the Eastern Carpathians, equipped with predominantly short-period sensors, is compared with synthetic modelling of anelastic attenuation of teleseismic waves in the upper mantle. Using the t^* approach, we examine variations of amplitude decrease over frequency for teleseismic recordings in the frequency band 0.5–1.5 Hz. The results reveal a consistent pattern of increased t^* values in the centre of the network, in the Vrancea region at the bend of the Carpathian Arc, although the magnitude of the observed variation in t^* is much higher than expected. Synthetic t^* parameter computations for the same event-receiver configurations reproduce the observed pattern in terms of relative variations. However, the amplitude of the synthetic t^* values explains only 10–20 per cent of the observed variation in t^* . t^* is not a direct measure for anelastic attenuation but rather for a combination of anelastic and other attenuating effects such as scattering and amplitude fluctuation related to velocity inhomogeneities. If regional amplitude variations are solely attributed to anelastic attenuation, all other effects are mapped into Q . We discuss the role of anelastic attenuation and other effects in the case of the Eastern Carpathians and conclude that t^* is an unsuitable parameter to characterize anelastic attenuation in the Eastern Carpathians.

Key words: attenuation, Eastern Carpathians, P waves, upper mantle.

1 INTRODUCTION

Amplitude variations within a regional seismic network can be used to study attenuation characteristics of the Earth by estimating the quality factor Q . To determine the Q parameter, the logarithmic decrement of the wave amplitude over frequency is often used as a measure in global as well as regional studies (e.g. Teng 1968; Bock & Clements 1982; Roth *et al.* 1999; Schurr *et al.* 2003). Since the quality factor Q is more sensitive to temperature variations than elastic velocity (see Romanowicz 1998 and references therein) it is often applied as an input parameter to invert for anelastic parameters of the Earth. However, the amplitude of a seismic wave is affected along its travel path not only by anelastic attenuation but also by frequency-dependent scattering, focusing effects from heterogeneities or dispersion. Additionally, the amplitude of a body wave recorded at the surface can be significantly influenced by receiver-side structural variations in the crust and shallow lithosphere. Observed variations in amplitude over a certain frequency band may thus include any of the described effects with an *a priori* unknown weight.

In this study, we estimate the t^* parameter from short-period recordings of teleseismic earthquakes at a temporary network in the Eastern Carpathians and show that large variations of this parameter cannot be explained by a presumed distribution of anelastic Q_p in the upper mantle alone.

The tectonic setting in the study region results from the complex interplay of Cenozoic convergence related to Alpine collision. After continent–continent collision in the Northern Carpathians in Mid-Miocene, the westward-directed subduction led to the detachment of the descending lithospheric slab in the northern Carpathians with the detachment process migrating southeastwards. In the Vrancea region at the southeastern bend of the Carpathian Arc, with its confined volume of intermediate-depth seismicity, a high-velocity body surrounding the volume of seismic activity (Weidle *et al.* 2005) is believed to be the last detaching or delaminating segment of this detachment process. For a more detailed description of the tectonics in the region, see for example, Wortel & Spakman (2000), Sperner *et al.* (2001) or Russo *et al.* (2005) for a summary. The latter also indicates the still ongoing debates on the origin of seismicity in Romania.

2 THE t^* APPROACH

A common way to describe the attenuation properties of the Earth is the representation of the quality factor Q , given as a measure for the loss of energy of a seismic wave as it travels through the Earth, through the t^* parameter. t^* represents the mean attenuation along the entire travel path and is given by an integral over the ray path, similar as the traveltime t of the seismic wave. Thus it is numerically

possible to invert for 3-D variations of the integrand (involving Q) through the Earth, given an adequate data coverage (e.g. Roth *et al.* 1999; Schurr *et al.* 2003). Such studies claim to resolve anelastic properties in the subsurface, due to the fact that the quality factor Q is more sensitive to temperature variations than elastic velocity (see Romanowicz 1998 and references therein). For many studies on global and regional scale the t^* parameter was observationally determined to be somewhat constant for periods longer than 1 s and is expected to be ~ 1 s for P and ~ 4 s for S waves over teleseismic distances (e.g. Lay & Wallace 1995).

In theory, the ground displacement induced by a seismic wave is (in the case of an effective point source which is applicable in teleseismic studies) given by the moment tensor and the derivative of the Green function (Aki & Richards 1980) in terms of frequency as

$$U_n(x, f) = M_{pq} G_{np,q}, \quad (1)$$

where $U_n(x, f)$, M_{pq} , $G_{np,q}$ are the Fourier transforms of the n th component of ground displacement $U_n(x, t)$, of the moment tensor M_{pq} and of the derivative in q -direction of the Green function $G_{np,q}$, respectively (applying the summation convention over p and q); x is the space coordinate and f the frequency.

The Green's function encompasses all conceivable wave propagation effects, such as body waves with reflected, refracted and converted phases, surface waves, scattering of both wavetypes, attenuation, effects related to free surface and near surface elastic properties (site effects), etc. In order to extract a specific effect, for instance, the body wave attenuation of mantle and crust, simplifying assumptions are made about the Green function and selected portions of the wavefield are considered, such as the first few seconds of the body wave arrivals. A common approach for the Green function is to factorize it in the frequency domain

$$A(f) \sim A_0(f) A_{\text{Geo}} A_{\text{Site}}(f) A_Q(f), \quad (2)$$

where $A_0(f)$ is the source spectrum, A_{Geo} the geometrical spreading factor, $A_{\text{Site}}(f)$ describe site effects and $A_Q(f)$ the attenuation term.

Eq. (2) thus describes the variations of P -wave amplitude as the wave travels from source to receiver. Some of the factors can be easily estimated and removed from the data. The source displacement spectrum $A_0(f)$ can be approximated by an ω^{-n} decrease above the corner frequency ω_c with proposed values for n ranging between 1.5 and 3 (e.g. Bock & Clements (1982) and references therein). Since we consider only event-wise variations within the network the source effect on the measurements will be not considered although we are aware that it may induce an (event-wise) constant shift in t^* determinations.

Geometrical spreading (A_{Geo}) is independent of frequency and can be removed by correction of the recorded amplitudes with an empirical distance- and depth-dependent correction factor (Nolet *et al.* 1998).

The role of site effects has been intensively studied in earthquake engineering literature. Site effects play a considerable role in the amplification of horizontal components of S waves of ground motion. For very low frequencies (< 0.2 Hz) it may be admissible to ignore those effects within an array of recorders. At 1 Hz, however, dependent on the near-surface geology, they should be taken into account if variations of amplitudes within an array are to be studied. According to the basic theory of site effects (see, for instance, Kramer 1996), the amplification of vertically propagating S waves is modelled, so that by definition site effects on vertical components

are ignored. Consequently, the variations of amplitudes of vertical components are not studied. It is believed that they are much smaller as compared to shear waves since they are often used as reference values, for example, in the H/V method for quantifying (shear wave) site effects. Thus they are generally ignored ($A_{\text{Site}} = 1$) in teleseismic studies when results based on amplitudes are discussed.

The only remaining term in eq. (2) then is the attenuation term A_Q which describes the amplitude behaviour of the wave as it travels through the Earth. In the case of a wave travelling through a homogeneous, attenuating medium, the amplitude decays as

$$A(x, f) = A_0 e^{-\pi f x / v Q}, \quad (3)$$

where A_0 is the source amplitude, v the seismic velocity and the quality factor Q , defined as fractional loss of energy per oscillation cycle

$$1/Q = -\Delta E / 2\pi E.$$

Just like seismic velocity, Q varies within the Earth and $A(x, f)$ is thus implicitly path-dependent. Replacing x/vQ and accounting for variations in v and Q along the propagating path of the recorded wave arrival, the t^* operator is given as

$$t^* = \int_S \frac{ds}{v[x(s)]Q[x(s)]} \quad (4)$$

yielding for the amplitude spectra

$$A(f) \sim \exp(-\pi t^* f). \quad (5)$$

The parameter t^* is thus the integrated attenuation along the ray path S involving the seismic velocity v and the quality factor Q at location x . Beside the path-dependence in v and Q , the parameter is also a function of frequency because Q is not constant over the entire band of seismic frequencies and an attenuating medium needs to be dispersive for wave propagation to remain causal (Aki & Richards 1980; Lay & Wallace 1995). However, as Q may not be frequency independent in the investigated frequency band, we expect (as discussed in Section 6) rather a constant offset in all measurements from that effect than a change in the pattern of relative variations. To interpret the data recorded at the regional network, only variations (not absolute values) of the parameter t^* within the network are of interest. We refrain from presenting results derived with a spectral-ratio approach (Teng 1968) to calculate differential t^* (δt^*) values contributing to the fact that our main findings persist while the quality of measurements is significantly worse for spectral-ratio determinations than for the t^* values derived with the procedure described in Section 3. For a convenient presentation of the variations in t^* within the recording network we will in the following present, for each event separately, the difference of each individual t^* measurement to the mean t^* and denote it as Δt^* .

3 DATA AND PROCESSING

We investigated 16 teleseismic events recorded at the temporary CALIXTO network (Lorenz *et al.* 1999) in the Eastern Carpathians, Romania. From May to November 1999, 110 mobile stations were deployed in the greater Vrancea region. The majority of the stations were equipped with short-period Mark geophones (1 s period) but also 27 broad band stations with periods of 20–120 s were used. The acquired dataset was complemented with data from permanent instalments in the region, such as the Romanian Telemetry Network and a regional Strong Motion Network. The main processing of the

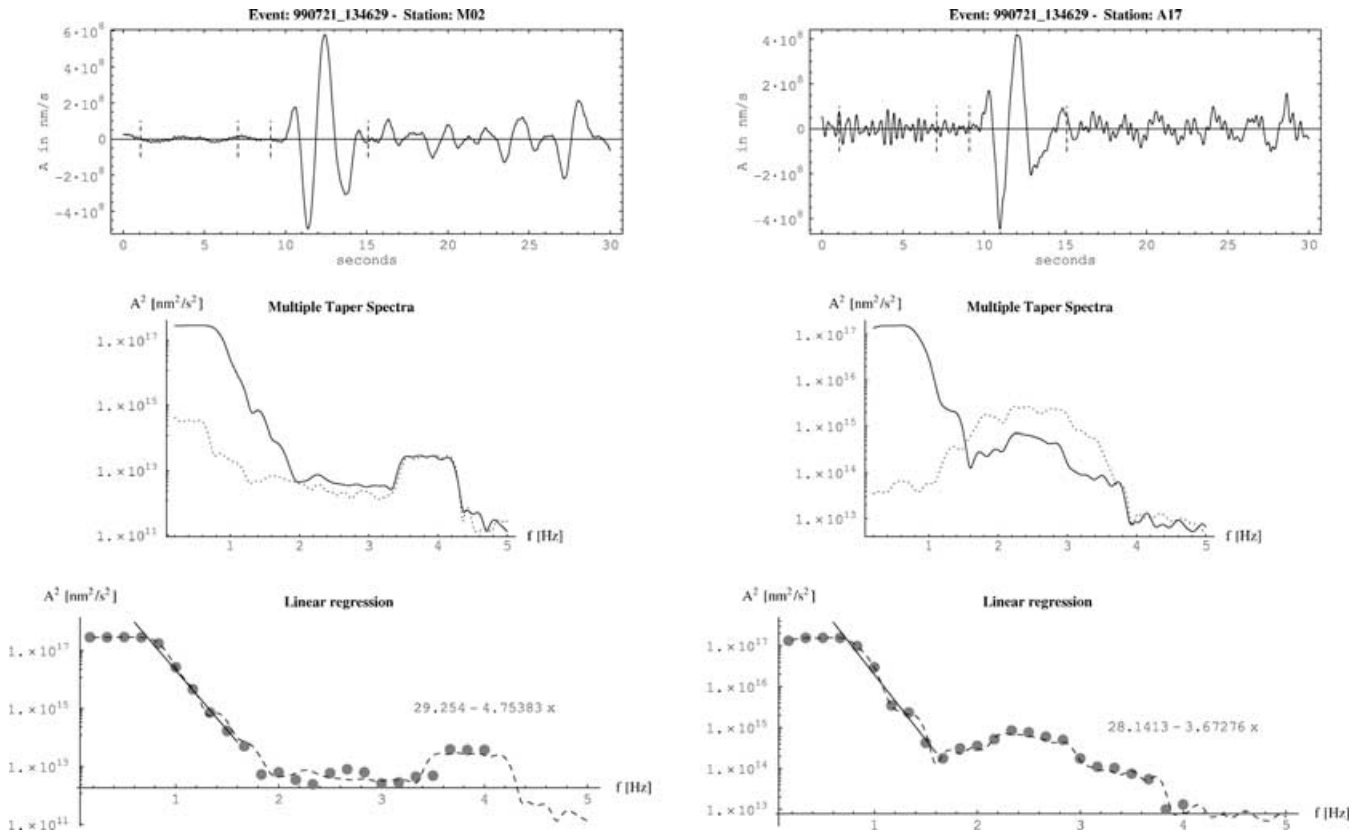


Figure 1. Two examples of recordings of the 1999 July 21 Sumatra earthquake (see Table 1). Station M02 is in the central part of the network, in the mountains whereas A17 is in the South Romanian lowlands in the southwestern part of the station network (see Fig. 2). Both are good examples for good (M02) and common (A17) signal-to-noise ratios. In the top plot the seismogram is displayed together with the noise and data windows indicated by vertical dashed lines. The centre graphs display the spectra calculated from the data (solid) and noise (dashed) window. It clearly shows dominating noise above 1.5–2 Hz in both recordings, independent of the noise level. The bottom graph displays the smoothed data spectra (dashed line), the independent frequency determinations (grey dots) and the linear regression through six points between 0.5 and 1.5 Hz. Note the different slopes.

raw data included deconvolution of the instrument response functions, standardization to a common data format (50 Hz sampling rate), and cutting of event windows adjusted to the NEIC event catalogue.

For this study, we searched a subset of all recorded events used earlier in traveltime tomography studies (Weidle *et al.* 2005; Martin *et al.* 2006) for events with an apparently convenient signal-to-noise ratio in a 5 s to 5 Hz band, constrained by the short-periodicity of the majority of the recording instruments. This was done by visual inspection of bandpass filtered seismograms, thus fully qualitatively as a first sorting process; at a later stage the remaining individual recordings were inspected more quantitatively. To ensure that we only use data with a clear and distinct P -onset, free of reflections like pP or PcP , all events with a traveltime difference of $T_{pP} - T_p < 8$ s were removed, as were events near the shadow zone and $\Delta > 100^\circ$.

Applying the absolute arrival time picks from Weidle *et al.* (2005), all traces (per event) were aligned on the P onset and time windows of the P wave onset from the seismograms with a length of 6 s extracted using a multiple-taper approach (Sudhaus 2005, adaptive multiple tapers with 3π -prolate spheroidal sequences). To assess the noise in the P -wave spectra, a 6 s pre-onset time window is extracted together with the P -wave window starting at the P wave onset and noise- and data-spectra are calculated, respectively. All spectra were manually revised and recordings with a signal-to-noise ratio < 2 below 2 Hz omitted. Due to our manual inspection procedure we

refrained from subtracting the noise from the data spectra, but determined the usable frequency band for t^* measurements manually as 0.5–1.5 Hz, since many of the remaining data reveal a significant increase in noise level above 1.5–2 Hz (Fig. 1). No correction for source spectra was applied, as we found it difficult to constrain a corner frequency consistent with all data (per event) that would have allowed, for example, a f^2 correction without deteriorating many of the spectra.

After calculating the spectra the dataset was again manually cleaned from defect and too noisy traces, before calculating t^* values through linear regression of the data spectra. The correlation coefficient of the linear regression is in general well above 0.8.

4 OBSERVATION OF t^* VARIABILITY ACROSS THE EASTERN CARPATHIANS

From the processing routine described above, 16 events remained which were subdivided into four source regions with different back-azimuth or distance, namely Hindukush, South-Asian, Japan-Kurile and Aleutian-Polar events (Table 1, Fig. 2). Through this subdivision, the reliability of the measurements can be validated, as events from the same source region should yield a consistent pattern of t^* observations. For convenient display of the variations in t^* , we subtracted the mean t^* per event from each measurement and give the

Table 1. Source parameters for events used in the attenuation study.

	Date (yr/month/d)	Lon (°E)	Lat (°N)	Depth (km)	M_w	Δ (°)	BAZ (°)
Hindukush	99/06/10	71.19	36.17	107.8	5.2	34.84	89.27
	99/06/21	70.71	36.34	218.5	5.4★	34.42	89.39
	99/06/26	69.44	29.99	25.1	5.6	36.96	99.40
	99/06/29	71.38	36.62	185.9	5.7	34.75	88.59
	99/07/12	69.44	29.97	20.8	5.7	36.96	99.46
South-Asia	99/06/02	123.47	00.09	158.6	5.7	75.85	42.08
	99/07/21	97.09	4.55	177.1	5.6	73.44	100.55
	99/07/22	91.92	21.60	1.1	4.4	58.02	91.10
	99/09/22	121.14	23.80	33.2	6.4	76.75	69.72
Japan-Kurile	99/07/07	155.48	49.27	43.4	6.1	75.83	31.54
	99/08/13	149.12	43.88	33.5	5.3	77.65	38.49
	99/08/20	135.43	34.06	64.8	5.6	78.21	53.28
	99/09/09	154.23	47.52	34.7	5.5	76.79	33.24
Aleutian-Polar	99/06/11	83.63	85.57	14.7	5.0	42.44	5.54
	99/09/28	168.176	54.64	22.3	6.1	75.19	21.75
	99/10/13	-161.205	54.64	23.9	6.4	80.01	4.38

Note. Distance (Δ) and backazimuth (BAZ) are given with respect to the centre of CALIXTO network at 26.25°E/45.5°N. The source locations are taken from Engdahl *et al.* (1998). ★: Magnitude is mb instead of M_w .

Events and station distribution

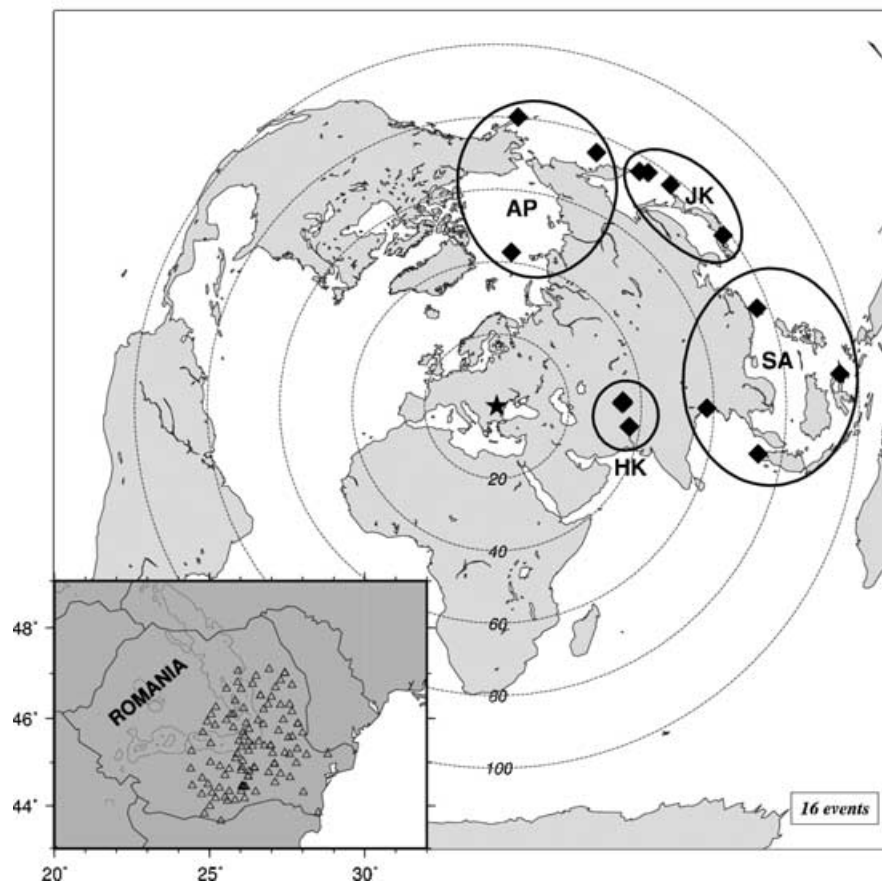


Figure 2. Location of events used in this study. Dashed lines mark epicentral distance from recording network (★). Annotated ellipses mark the four clusters of Hindukush (HK), South-Asian (SA), Japan-Kurile (JK) and Aleutian-Polar (AP) events. See also Table 1. Inset shows recording network in eastern Romania.

mean t^* as well as the mean correlation coefficient in the respective plots. All deviations from the mean $|\Delta t^*| \leq 0.15$ s are plotted as crosses of equal size, while for larger deviations $|\Delta t^*|$ is indicated through the respective symbol size.

4.1 Hindukush events (Fig. 3)

The events from the Hindukush region show almost identical patterns of t^* parameters. The variations in t^* for the Hindukush events

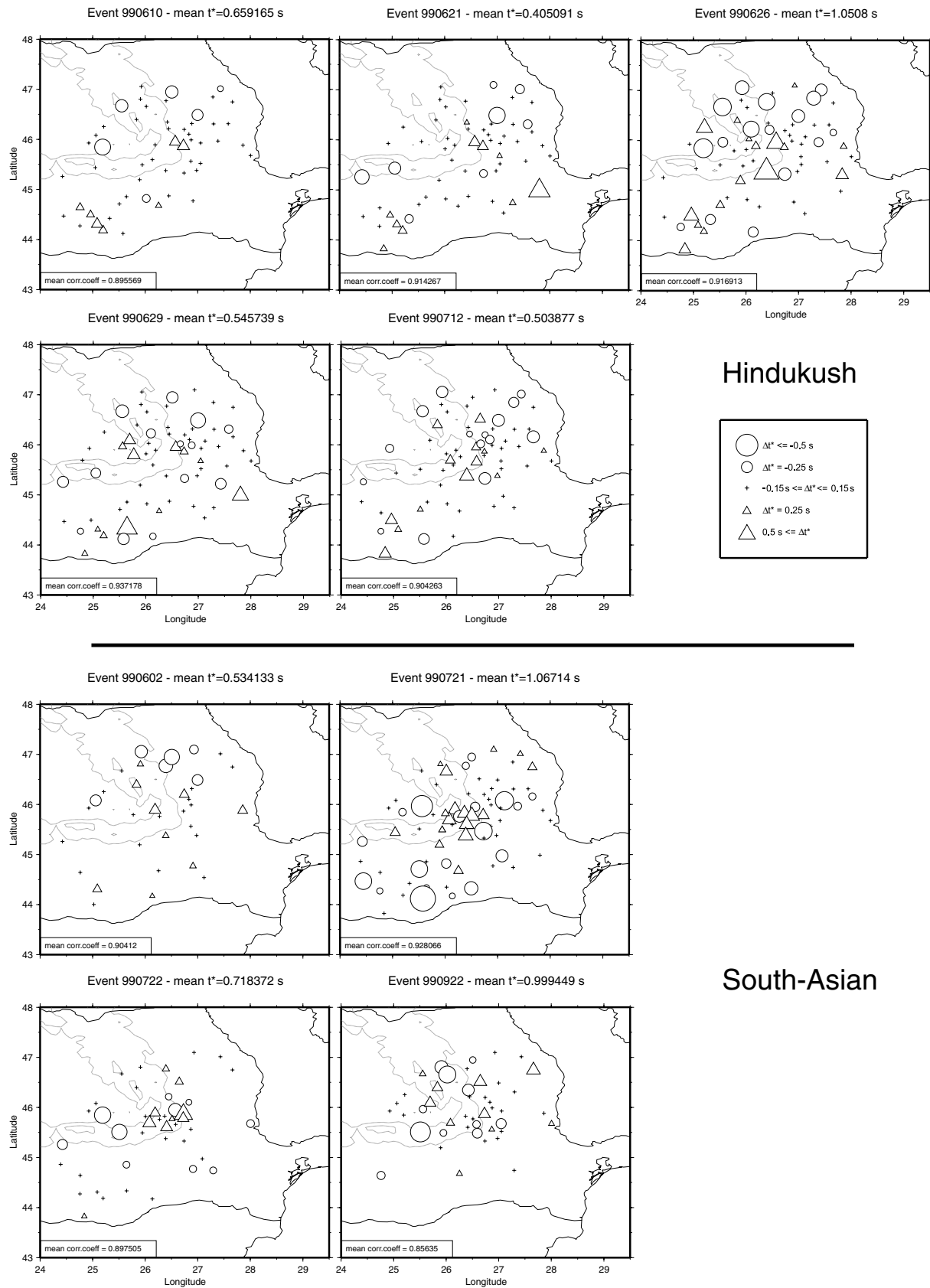


Figure 3. Distribution of t^* for the Hindukush events (top) and the South-Asian events (bottom) (see Table 1). The grey dashed lines indicate 800 m contours of the Carpathian orogen.

reveal a clear trend with lowest values in the north and northeast of the network ($t^* \geq 0$ s) and largest values observed at the network centre and southwesternmost stations ($t^* \leq 1.3$ s). It must be noted that the mean value and the spread of the June 26 event differs largely from the others although the observed pattern is consistent.

4.2 South-Asian events (Fig. 3)

The northern Sumatra event (1999 July 21) is the highest quality recording of the compared measurements. An excellent signal to noise ratio allowed for 73 stations to be used. In contrast to the Hindukush events, the t^* distribution reveals a slightly different pattern of variation. The largest t^* values are found at the centre of the network (≤ 1.25 s) while t^* decreases with distance in all directions. The lowest values are found in the southern part of the network (≤ 0.25 s). The t^* observation for the Sumatra event is supported by the t^* observations from the Burma event (1999 July 22). Although the data quality resulted in fewer usable stations, the pattern of largest t^* values in the centre of the network is confirmed. The other south Asian events in eastern Indonesia (Molucca Sea, 1999 June 2) and Taiwan (1999 September 22) also tend to reveal largest values in the centre although with less significance.

4.3 Japan-Kurile events (Fig. 4)

The events from the Japan-Kurile region again show stronger coherency in t^* . The pattern of variation can be roughly divided into two portions with moderate to higher values north of an imaginary line prolonging the southern Carpathians to the east and pronounced lower values to the south. However, the maximum t^* observations are found approximately in the centre of the network.

4.4 Aleutian-Polar events (Fig. 4)

Finally, two events from the Aleutian Arc and one from the northern Polar Sea were examined. For the Aleutian events (1999 September 28 and October 13), the t^* values reveal a pattern consistent with the previous observations of a maximum in the centre of the network and comparably larger t^* values in the north than in the south. A similar observation holds for the event in the Polar Sea (1999 June 11) with few recordings in the southern part of the network, but a slight decrease in t^* from the Vrancea Region northwards.

4.5 Summary of observations

For all events, the observed t^* parameters range between 0.2 and 1.7 s which is in general agreement with other observations of teleseismic P -waves yielding on average $t^* \sim 1$ s for 1 Hz (e.g. Anderson & Given 1982; Bock & Clements 1982; Lay & Wallace 1995; Warren & Shearer 2000). Der (1998) summarizes that a global average of t^* is between 0 and 0.5 s and values close to 1 s are rather rare, although considerable variations in t^* are observed worldwide.

Regarding the variations within the network, one property consistent through most observations is found for distances $>50^\circ$, that is that the largest t^* parameters are observed in the central part, the Vrancea region, which is located right at the bend of the Carpathian Arc. Apart from that, slightly higher t^* values are found in the northern, especially northeastern part compared to lower t^* values south of the Vrancea Region. Nevertheless, the events from the Hindukush region deviate from that pattern with a clear north-south trend of increasing t^* . As this source region is closest to the observational

network (epicentral distance $\sim 38^\circ$) this might be correlated to lower angles of incidence of the teleseismic waves and thus reflect shallower structures.

The most intriguing observation is however the comparably large range of variability in which t^* is observed in the region ($|\Delta t^*| \leq 1$ s). This means that the variation of t^* within the network is of the same order of magnitude as its absolute value of integrated attenuation along a ray path with a length of several thousands kilometres. Under the assumption that source and common paths effects are irrelevant by discussing only variations within the network, the observations are significantly larger than one would expect as a result from variations in attenuation parameters in the crust and upper mantle beneath the study region.

On the other hand, the observed patterns and amplitude variations in t^* are consistent over all recorded earthquakes thus indicating stable results. Arbitrary mapping of uncertainties or noise should not result in such a consistent pattern of t^* variations. However, the question of ‘what structural variations do the observations reflect’ remains. Using a priori information of the subsurface structure from teleseismic tomography, synthetic t^* parameters are calculated in the following to estimate what magnitude of t^* and its variation Δt^* might be expected from structural variations of anelastic attenuation in the upper mantle.

5 MODELLING OF ANELASTIC ATTENUATION

Extensive research over the last 20 years, especially recently published constraints on the crustal and upper mantle structure in the Eastern Carpathians offer the opportunity to model the t^* pattern as it may be expected from anelasticity in the upper mantle. Anelastic attenuation mechanisms are related to heat losses or thermal relaxation processes. In attenuation tomography studies (e.g. Schurr *et al.* 2003), observed attenuation is thus often interpreted as temperature variations in the upper mantle, where scattering is minor or at least hardly detectable at the surface. Based on a velocity model of the upper mantle, a temperature and subsequently a Q_P model is derived in the following and synthetic t^* parameters calculated.

A recent P -wave velocity model derived by teleseismic tomography (Martin *et al.* 2006) is used to estimate the temperature field beneath the Eastern Carpathians. The seismic tomographic model of the region consists of eight layers of different thicknesses (from 15 to 50 km), which are each subdivided laterally into $42 \text{ km} \times 42 \text{ km}$ blocks. To convert the P -wave seismic velocity anomalies beneath the region into temperature we model initially synthetic P -wave seismic velocities considering the effects of anharmonicity (composition), anelasticity and partial melting on the seismic velocities (Ismail-Zadeh *et al.* 2005). Ismail-Zadeh *et al.* 2005 tested several compositional models for the slab, shallower (above 300 km) and deeper (below 300 km) mantle in the conversion from P -wave velocities to temperature and found that the magnitude of the compositional changes is relatively small (less than about 1% in temperature).

Once the synthetic velocities are calculated for a first-guess (background) temperature, an iterative process is used to find the ‘true’ temperature, minimizing the difference between the synthetic and ‘observed’ (in seismic tomography experiments) velocity anomalies. The temperature in the shallow levels of the region is constrained from measured surface heat flux corrected for palaeoclimate changes and for the effects of sedimentation (Demetrescu *et al.* 2001).

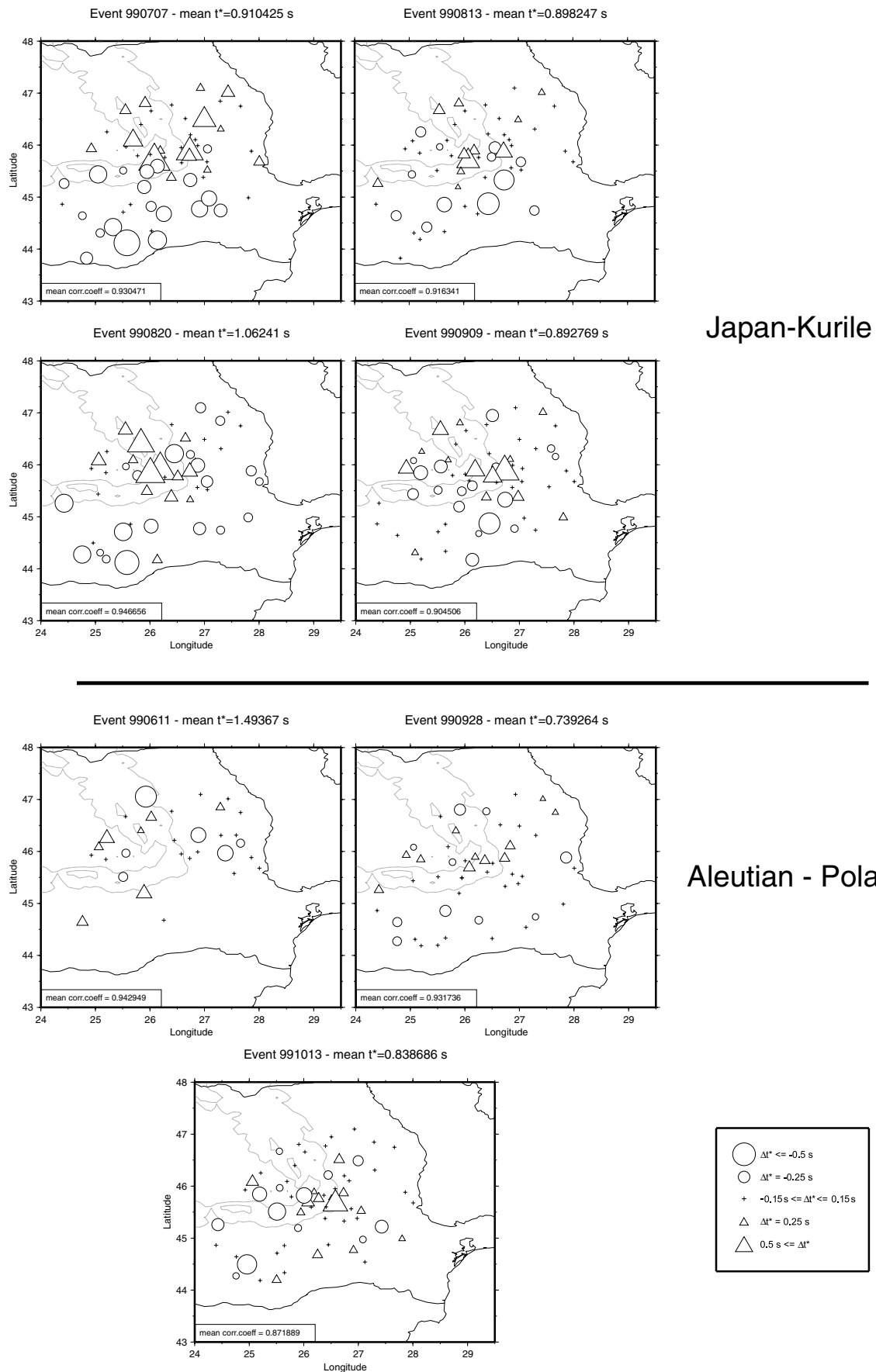


Figure 4. Distribution of t^* for the Japan-Kurile events (top) and the Aleutian-Polar events (bottom).

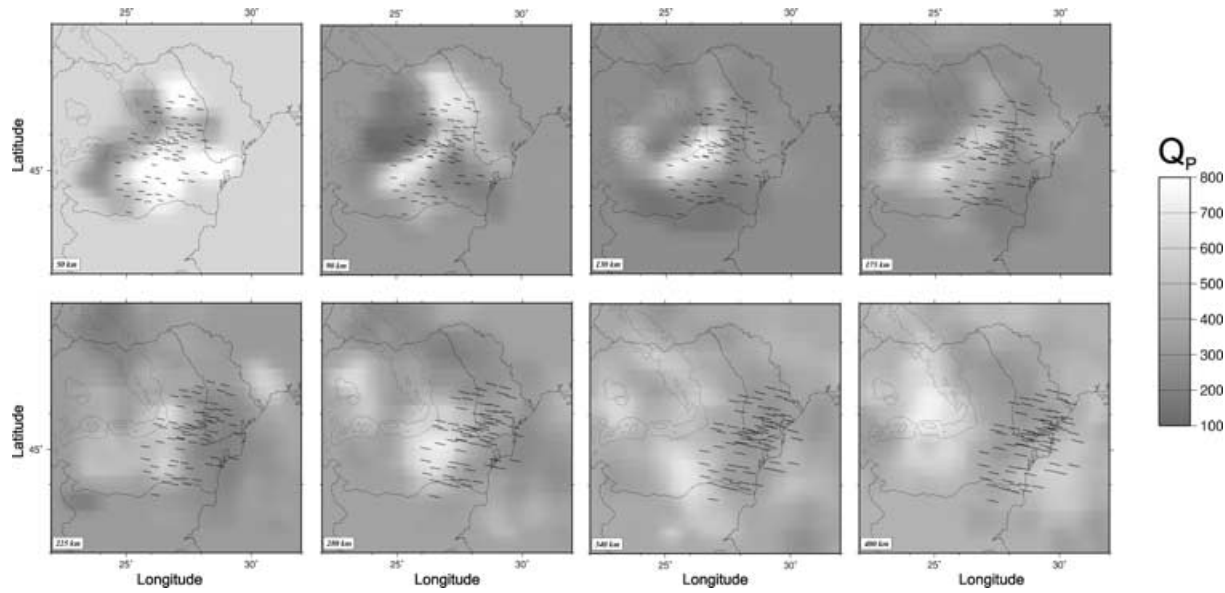


Figure 5. Q_P model for the Vrancea region superimposed by ray paths (black line segments) at corresponding depths calculated for the Sumatra earthquake (1999 July 21).

Following Karato (2004), the shear wave Q -factor is subsequently calculated as

$$Q_S = A\omega^a \exp[a(H^* + PV^*)/RT]$$

with $A = 0.148$, $a = 0.15$, the activation energy $H^* = 500 \text{ kJ mol}^{-1}$, P representing pressure, the activation volume $V^* = 2 \times 10^{-5} \text{ m}^3 \text{ mol}^{-1}$, the gas constant R and the temperature T . For the sake of simplicity, we assume that the average bulk Q -factor does not depend on temperature and frequency and use the average $Q_K = 479$ at a wave period of 1 s (Anderson & Given 1982). The P wave Q -factor is consequently calculated as

$$Q_P^{-1} = L \cdot Q_S^{-1} + (1 - L) \cdot Q_K^{-1}, \quad L = 4v_S^2/3v_P^2$$

with v_P and v_S representing synthetic P - and S -wave velocities for given compositions of the slab and mantle. The resulting model of Q_P variations in the upper mantle between $41^\circ\text{--}50^\circ \text{ N}$, $21^\circ\text{--}32^\circ \text{ E}$ and 35–440 km depth (Fig. 5) predicts high values for the Vrancea slab (<800) and relatively small values for the surrounding mantle, increasing from $\sim 100\text{--}200$ at 90 km depth to $\sim 300\text{--}500$ beneath 340 km depth. The corresponding values for shear Q_S correlate well with studies by Anderson & Given (1982), Umino & Hasegawa (1984) or Sato (1992).

Since the temperature model is derived from the P -wave velocity anomaly model by Martin *et al.* (2006), the geometries of the seismic velocity anomalies are to a large extent preserved in the temperature and Q models. Comparison of traveltime and attenuation tomography results show generally similar geometrical patterns: studies in subduction zones reveal basically pronounced low Q values in the backarc regions and high values of Q in the subducting slab correlating geometrically well with observations of negative and positive anomalies of wave speed perturbation, respectively (e.g. Roth *et al.* 1999; Schurr *et al.* 2003).

The 3-D ray tracing algorithm by Koketsu & Sekine (1998) is used to trace ray paths from the source to the receivers through a 1-D reference model (ak135 by Kennett *et al.* 1995), superimposed by the 3-D P -wave velocity model by Martin *et al.* (2006) beneath the study region. As the frame of the velocity and the synthetic Q_P model are identical, synthetic t_{an}^* values can be directly calculated

by solving the integral in eq. (4) along the calculated ray path within the model frame. Corresponding to teleseismic tomography, a plane-wave approximation is applied which in this context assumes that variations of observed attenuation within the network are generated only inside the model frame. Attenuation effects accumulated along the rest of the waves' travel paths are not considered as they are assumed to be identical for all observations within the regional network. The synthetically modelled t_{an}^* can then be compared with the observed t^* measurements.

The modelling results reveal that for rays coming from easterly directions, that is, events from the Hindukush and South Asia, the highest t_{an}^* values are projected into the centre of the network, the Vrancea region and the deep forearc basin region (Fig. 6). To explain this, one must account for the integrative accumulation of attenuation and hence the projection of lower Q_P regions onto the surface. Plotting of ray paths in the Q_P model for the Sumatra event in Fig. 5 reveals that, for example, rays ending at stations in the far west of the network (near the southern Carpathians), accumulate the lowest attenuation ($t_{\text{an}}^* \sim 0.1 \text{ s}$) as they pass from bottom to top of the model entirely through structures with $Q_P > 400$. Contrarily, rays arriving at the centre of the network have mainly travelled through regions where $Q_P < 500$, although they do not pass a specific depth with exceptional high attenuation. Among the events in the east, the Hindukush events show generally larger t_{an}^* values than the more distant events. This is easily explained by the shallower ray paths, and therefore longer ray segments within the Q_P model than the steeper in coming rays from larger distances. The overall pattern for the Hindukush events is, however, identical to those from the other easterly events.

For the events in the north and northeast, that is, Japan-Kurile and Polar-Aleutian events, the pattern of t_{an}^* parameters in Fig. 6 is slightly different. The highest t_{an}^* values are shifted to the southeast of the Vrancea region and additionally slightly higher t_{an}^* values are found at the stations in the Transylvanian Basin. Investigation of the ray paths in the Q_P model reveals that this pattern is dominantly influenced by the shallower layers of the model. Especially the stations in the Transylvanian Basin are strongly attenuated at 90 km depth, as they pass through the absolute minimum of $Q_P \sim 100$ in

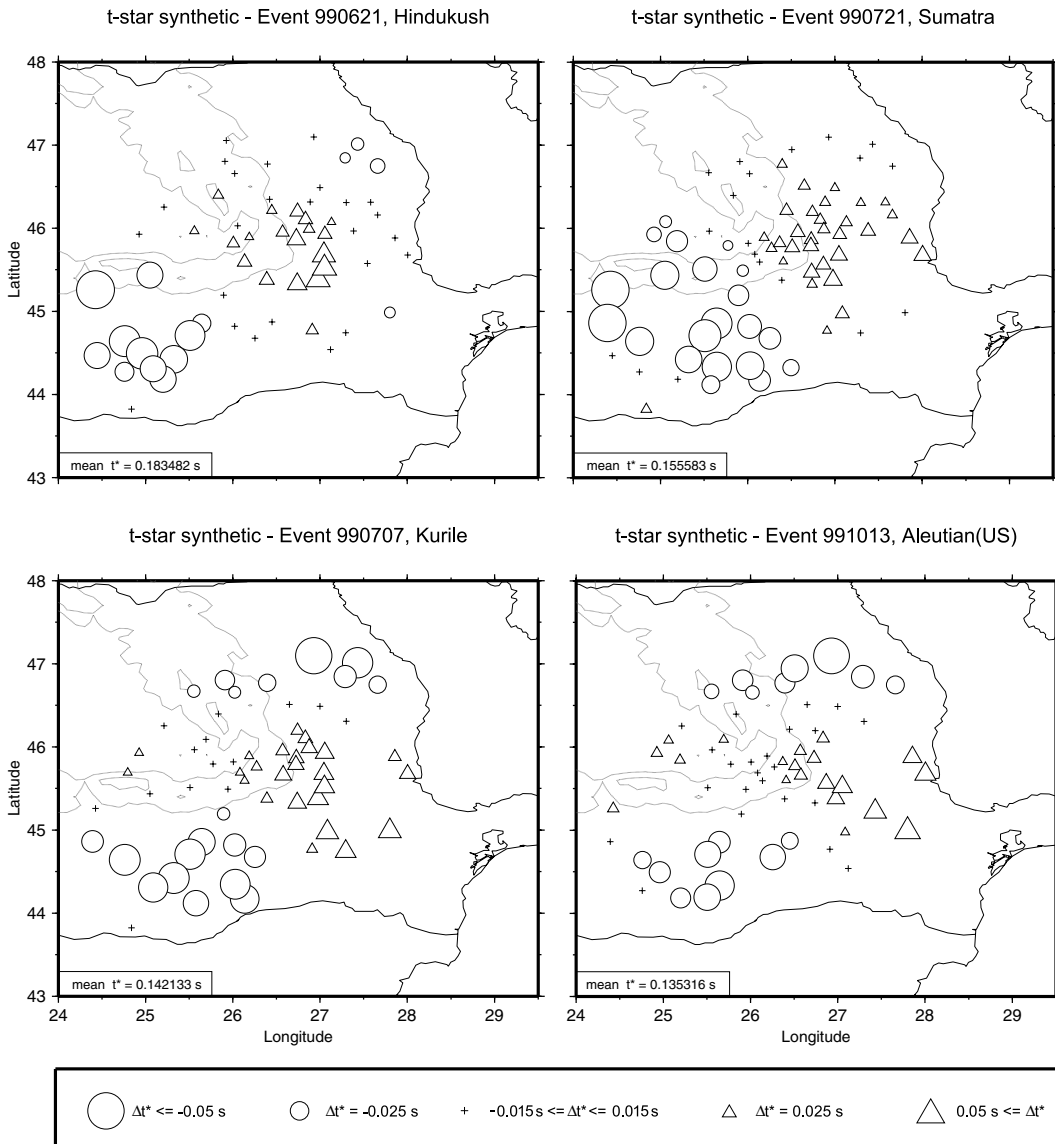


Figure 6. Synthetically modelled t^* operators. For each source region, one exemplary distribution is given. Note that the scale is a factor 10 smaller than in Figs 3 and 4.

the model. Another intriguing observation is that almost all of the stations in the southern part of the network reveal t^* values close to or even lower than 0.1 s.

Despite the somewhat consistent pattern of variations for the different events with the observation of real data t^* , the amplitudes of t^* variations $|\Delta t^*|$ are significantly less. They range between 0.08 and 0.11 s and can thus only partially explain the real data observations ($|\Delta t^*| = [0.5; 1.1]$ s).

Naturally, the Q_P -model is only an estimate considering the uncertainties in the assumed parameters. The choice of parameters, for example, the bulk Q -factor Q_K directly influences the resulting Q_P . Applying the PREM value $Q_K = 57822$ (Dziewonski & Anderson 1981) instead of $Q_K = 479$, the Q_P values in the resulting model double on average. Consequently, the synthetic t^* reduce by a factor of 2 but still, the pattern of variation is preserved. On the contrary, if we try to achieve the measured range of t^* determinations by tuning the model, the amplitudes of t^* must increase by a factor of 5–10. In such a scenario, Q_P must be reduced by the same factor leading

to a Q_P model with unrealistically low values. We thus assume that the derived Q_P model presented above is sufficiently accurate for our purposes.

6 DISCUSSION

The energy loss described by t^* was originally given as a parameter for anelastic attenuation (Ben-Menahem *et al.* 1965). However, the attenuation term A_Q (eq. 2) includes also amplitude manipulations from other effects. While intrinsic attenuation implies energy loss due to internal friction (e.g. heat losses, thermal relaxation), attenuating effects from scattering or defocusing of the wavefront affect the direct wave amplitude through redistribution of energy in space and time (Sato & Fehler 1998). This happens on all scale-lengths of heterogeneities each affecting a different frequency range of the entire spectrum which in turn yields a strong frequency dependence of attenuation.

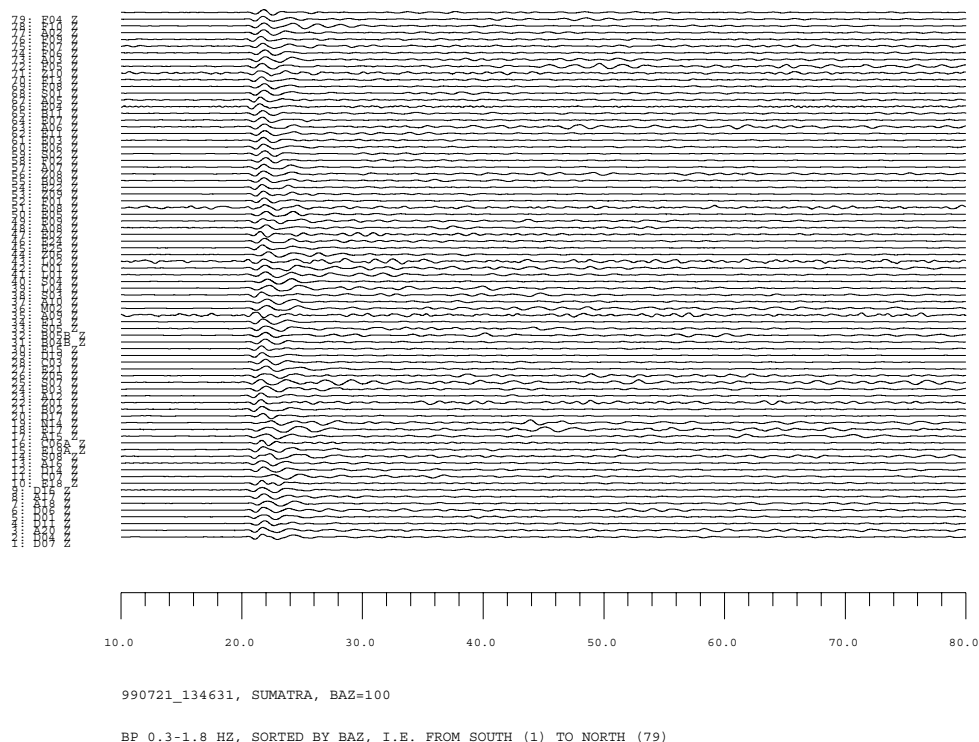


Figure 7. Seismic section through selected stations of the network from south (trace 1) to north. The traces are shifted to the manually picked P -onset. Note the increase in Coda towards the centre of the network (i.e. Carpathian Arc bend region).

The quality factor Q is thus often separated into an anelastic Q_{an} and a scattering Q_{sc} , but a quantification of the respective contributions to the observed t^* values is generally difficult. Q_{sc} is defined only for small-scale scatterers much smaller than the wavelength (λ), resulting in a limited distortion of the first-arriving wavefront. Through scattering at such small-scale velocity heterogeneities, the wavefront is broadened (in the direction of propagation) as energy of the direct arrival is partly shifted into the coda, but the wavefront itself remains coherent. With increasing scale length of the heterogeneities (with respect to λ) this ‘simple’ definition of a scattering Q is no longer necessarily valid as for longer travel distances the wavefront may become incoherent due to defocusing at long-wavelength components (Shapiro & Kneib 1993; Sato & Fehler 1998).

We present a seismogram section through selected stations of the network in Fig. 7. Although the signal-to-noise ratio is variable, it is obvious that the coda length after the P onset shows some distinct variations along that north–south profile (trace 1 is the southern-most). Especially the stations in the Carpathian Arc bend zone (traces 35–50) show a deformation of the P waveform and a clear coda over up to 15 s length. Although this is only a qualitative observation, the inference of scattering, especially variations in scattering throughout the study region is reasonable.

Defocusing effects of the seismic wavefront due to diffraction at extended velocity heterogeneities were in the past commonly neglected in attenuation studies for both local and teleseismic sources. However, Allen *et al.* (1999) showed that focusing effects can play a dominant role in explaining the pattern of observed t^* parameters.

For a dominating incoherent wavefield the recorded amplitude of the first arrival may be attenuated as well as amplified due to destructive or constructive interference of (de-)focused wave paths, respectively. To quantify the contributions of scattering and (de-)focusing to our t^* observations, waveform modelling could be

an appropriate method. Miksat (2006 and personal communication) showed by FD waveform modelling of the strong Vrancea (1986) earthquake that the topography of the innercrustal discontinuities, especially the basement topography has a major influence on the distribution of peak ground accelerations in Romania, whereas the presence of random heterogeneities in the mantle appears to have a minor effect. However, it is unclear how far both the structure and random heterogeneities affect the amplitude of a teleseismic P wave.

Another amplitude amplification effect that must be taken into account is site effects which is generally not discussed in the context of teleseismic P -wave amplitude studies (and thus the t^* approach). Compressional waves are assumed to be much less affected by site effects than shear waves and hence often treated as a reference for the amplification of shear amplitudes. However, for larger heterogeneities (or thicker layers), such effects must be theoretically considered to exist for P waves in a comparable range as they do for shear waves and there is thus no reason to omit them from the discussion, moreover as no studies on site effects of P waves are available. Especially for investigations of shorter period data with frequencies well above 0.1 Hz, site effects must be taken into account to appear significantly in the observation of t^* parameters not only by station-to-station variations but also by larger scale patterns of regional fluctuations reflecting extended, shallow geologic structures.

In our study region, site effects have been studied (e.g. Sokolov *et al.* 2004; Wirth *et al.* 2003) and significant variations reflecting the geological variety of the region have been observed, although the amplification functions show much less differences at low frequencies of 1 Hz as compared to 5 or 10 Hz.

For this study, we have to consider that significant site effects in the observation of P -wave amplitudes at the investigated frequency band would have to be independent of distance and backazimuth

of the recorded earthquakes. However, this is not observed in our results. Additional visual inspection of vertical components data from the German–Romanian K2 strong motion network indicates that the basic assumption of negligible amplification on the vertical component is acceptable.

We also assumed that Q is independent of frequency according to the narrow frequency band investigated. In theory, if Q does depend on frequency, the linear relation of t^* and logarithmic amplitude changes to an apparent $t^*_{\text{app}} = t^* + f(dt^*/df)$ (e.g. Der (1998)). However, frequency dependence of Q is generally assumed to be small, for example, Oki *et al.* (2004) give a power law with exponent 0.04 assuming that Q in PREM is strictly given only at 200 s. Thus, dt^*/df should be negligible especially when only small frequency windows are applied. Since dt^*/df may be in any case relatively constant for all spectra analysed, accounting for frequency dependence may have the same effect for all paths—changing the mean t^* rather than variations in it.

Generally it is thus clear, that investigation of amplitude variations using the t^* parameter must be done carefully with respect to the investigated frequency band. While for long-period investigations, anelastic attenuation can be reasonably expected to be dominant, one must consider that for short-period data, the amplitude of a wave arriving at teleseismic distances may be affected not only by anelastic attenuation but also considerably through velocity heterogeneities with a lateral extent of the order of a wavelength or less and thus beyond resolution of seismological imaging methods.

7 CONCLUSIONS

In the presented study, teleseismic P waveforms recorded at the temporary CALIXTO network in the Eastern Carpathians were used to determine the decay of spectral amplitude in a narrow band of 0.5–1.5 Hz. The observations from the recorded data reveal a generally consistent pattern of largest t^* in the centre of the network, right at the bend of the Carpathian Arc, where crustal and upper mantle structures reveal the most complex interrelation in the region. Also, t^* values tend to be higher north of the centre as compared to south of it. When comparing only the pattern, not the amplitude of the t^* observations, this agrees fairly well with the observation by Ivan (2003) who found higher attenuation in the Focsani basin and lower attenuation around Bucharest and in Dobrogea. Investigations of teleseismic events, accidentally recorded during a seismic refraction experiment in the Eastern Carpathians also revealed strong amplitude variations along the profile with the largest t^* parameters in the centre part of the profile (Sudhaus & Ritter (2005)).

Recently, Russo *et al.* (2005) concluded that attenuation is low north and east of Vrancea while it is high above and near Vrancea. This conclusion is in general agreement with our observation of maximum t^* in the Vrancea Region. Russo *et al.* investigated local Vrancea earthquakes and calculated shear wave attenuation Q_S (at higher frequencies, approximately 2–15 Hz, no specific values are given). Our synthetic Q model (see Fig. 5; the Q_S values form a wider scale than the displayed Q_P , approximately 50–1000, but the geometries are the same) coincides quite nicely with their findings of high Q_S values east of the Vrancea seismic region ($Q_S > 400$) and low Q_S above and west of Vrancea ($Q_S < 200$).

The consistency within the investigated data (with respect to the large scale patterns) as well as the agreement with findings from other studies indicate that the observed patterns in t^* do indeed reflect structural variations. However it is also clear, that these variations cannot be simply assigned to variations in Q_{an} . We estimated

the contribution of a purely anelastic Q_{an} by a modelling approach and found that structural variations in anelastic attenuation in the upper mantle can only partly explain the amplitude of the observed variations in t^* . The modelled variations in t^*_{an} account only for 10–20 per cent of the observed t^* variations within the CALIXTO network. Other studies (e.g. Akinci *et al.* 1995; Hock *et al.* 2000) have also shown that anelastic attenuation can be inferior to scattering attenuation at frequencies around 1 Hz.

In conclusion, the presented t^* observations lead to an ambiguous interpretation.

(1). The results appear to be questionable due to the fact that the observed variations Δt^* reveal an interval within the network of the order of absolute t^* determinations for entire ray paths. Additionally, the investigated frequency band is too narrow to exhibit lateral variations in attenuation of the teleseismic wavefront but may be equally explained by local amplification or reduction of amplitudes due to the geometry of the shallow subsurface. Interpretation of observed t^* values as variations in attenuation properties is hence not plausible.

(2). Nonetheless, the observed variations are stable over nearly all investigated back-azimuths and incidence angles. Slight deviations from the consistent pattern (e.g. for the Hindukush events) lead to the conclusion that site effects cannot be expected to be dominant. From the stable pattern one may argue that the results are reliable and do reflect structural variations on the scale of a P -wave wavelength although interpretation in terms of 3-D structure is impossible. With respect to the complicated crustal and subcrustal structure in the Vrancea region (e.g. Hauser *et al.* 2001), the inference of scattering in both mantle and crust is reasonable. Our results may thus reveal a correlation of anelastic (controlled by temperature) and elastic (reflecting the intensity of scattering in the crust and upper mantle) attenuation effects. High attenuation in the mantle is accompanied by strong scattering in the mantle and probably the crust and conversely low intrinsic attenuation is associated with little scattering. At this point we do not speculate on why this relation exists and what the implications for the tectonics and mechanics of the lithosphere/asthenosphere system are. However it may turn out that although t^* is an unsuitable parameter for measuring the intrinsic attenuation, combined with studies of tomography and thermomechanic modelling it may help to reveal properties of the deep earth's interior.

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