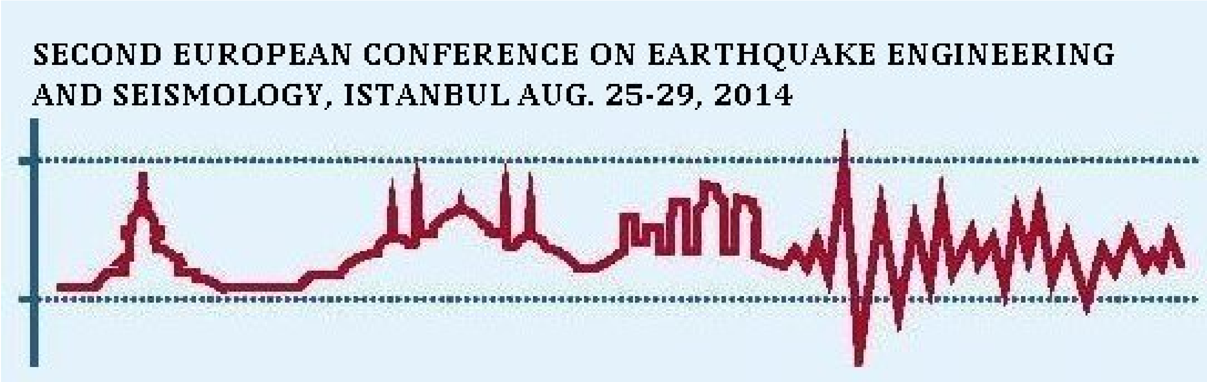

Q_P AND Q_S ATTENUATION MODELS OF THE SOUTHERN AEGEAN SUBDUCTION AREA

Chrisanthi VENTOUZI[1], Constantinos PAPAZACHOS[2], Christos PAPAIOANNOU[3], Panagiotis HATZIDIMITROU[4] and the EGELADOS working group

The anelastic attenuation (Q-factor) has been considered one the main controlling factors affecting seismic wave propagation. Studies of the Q-factor also provide significant information on the earth’s structure, since several geodynamics processes (e.g. subduction, temperature, volcanism, etc.), (Stachnik et al., 2004; Pozgay et al., 2009) are reflected in the 3D distribution of intrinsic attenuation, which controls the spatial distribution of earthquake ground motions. In order to examine the attenuation structure of the broader Southern Aegean subduction area, more than 300 intermediate-depth earthquakes which occurred between 2002 and 2007 were employed. The events were recorded by two local networks which operated in the area for the aforementioned time period. The first network is CYCNET (Bohnhoff et al., 2004), which operated from 2002 to 2005 in the broader region of Cyclades islands (central southern Aegean). The second network was the large scale local network of EGELADOS seismic monitoring project, which consisted of 65 land stations and 24 OBS recorders, which operated from 2005 until 2007. All data from the local networks were enriched by data provided by the permanent networks of National Observatory of Athens and GEOFON.

Using the recorded waveforms we calculated a frequency-independent path attenuation operator t^* for each waveform for both P and S waves. Two independent approaches were adopted for the t^* determination. At first, an automated method was used, where t^* was automatically calculated from the slope of the acceleration spectrum above the corner frequency, f_c, assuming an ω^2 source model. Calculations of t^* were carried out in the frequency band of 0.2 to 25 Hz, using only spectra with a signal-to-noise ratio larger than 3 for a frequency range of at least 4Hz for P-waves and 1Hz for S-waves. In the second approach, t^* computations were performed manually after a visual inspection by the user for the selection of the linearly-decaying part of the spectrum, in order to obtain an optimal spectral fitting. Estimated travel times as a function of epicentral distance were generated, using data provided from the CYCNET experiment for different groups of focal depths, in order to use them when original travel time information was not available.

The values of t^*, as estimated from both techniques, do not show any significant dependence with hypocentral distance which is clearly a result of the strong spatial and depth variations of the anelastic attenuation, that superimposes the distance effect. In order to further investigate this observation, we looked at the spatial variation of t^* values for different hypocentral-depth groups. The derived results reveal that along-arc stations show significantly lower values of t^*, while back-arc stations exhibited much larger values. The observed t^* along-arc/back-arc differences becomes more pronounced as the depth of the earthquakes increases, pointing out the effect of the high-attenuation (low-Q) mantle wedge beneath the volcanic arc.

To invert for the 3-D Q-structure of Southern Aegean area, a 3-D Cartesian coordinate system was created and a 3D grid of nodes was generated. Each block was assumed to have constant t^* value

[1] Phd Candidate, Aristotle Univeristy of Thessaloniki,Thessaloniki, xrusven@geo.auth.gr
[2] Professor of Geophysics, Aristotle Univeristy of Thessaloniki,Thessaloniki, kpapaza@geo.auth.gr
[3] Research Director, E.P.P.O.-ITSAK, Thessaloniki, chpapai@itsak.gr
[4] Professor of Seismology, Aristotle Univeristy of Thessaloniki,Thessaloniki, chdimitr@geo.auth.gr

and for each point in space the corresponding value was calculated by three-dimensional cubic interpolation. All the foci of the intermediate depth earthquakes considered are distributed within this grid of nodes. Using this grid of nodes we have determined a set of the attenuation equations, which concern the spectral slopes calculated with both approaches, leading to the generation of a linear system of equations. This system's unknowns are the attenuation operators. A 3D ray tracing technique proposed my Moser et al. (1992) and modified by Papazachos and Nolet (1997) was adopted in order to trace the seismic rays for all events. Figure 1 shows the logarithm of the density of seismic rays for the study area considered for the three-dimensional model, for both P- and S-wave spectral slopes calculated by the automatic approach. A better coverage of seismic rays is observed for the depths of 50-80 km, while for larger depths the rays are mainly localized in the eastern part of the region, where the generation of intermediate depth earthquakes with larger depths (>100 km) is more frequent.

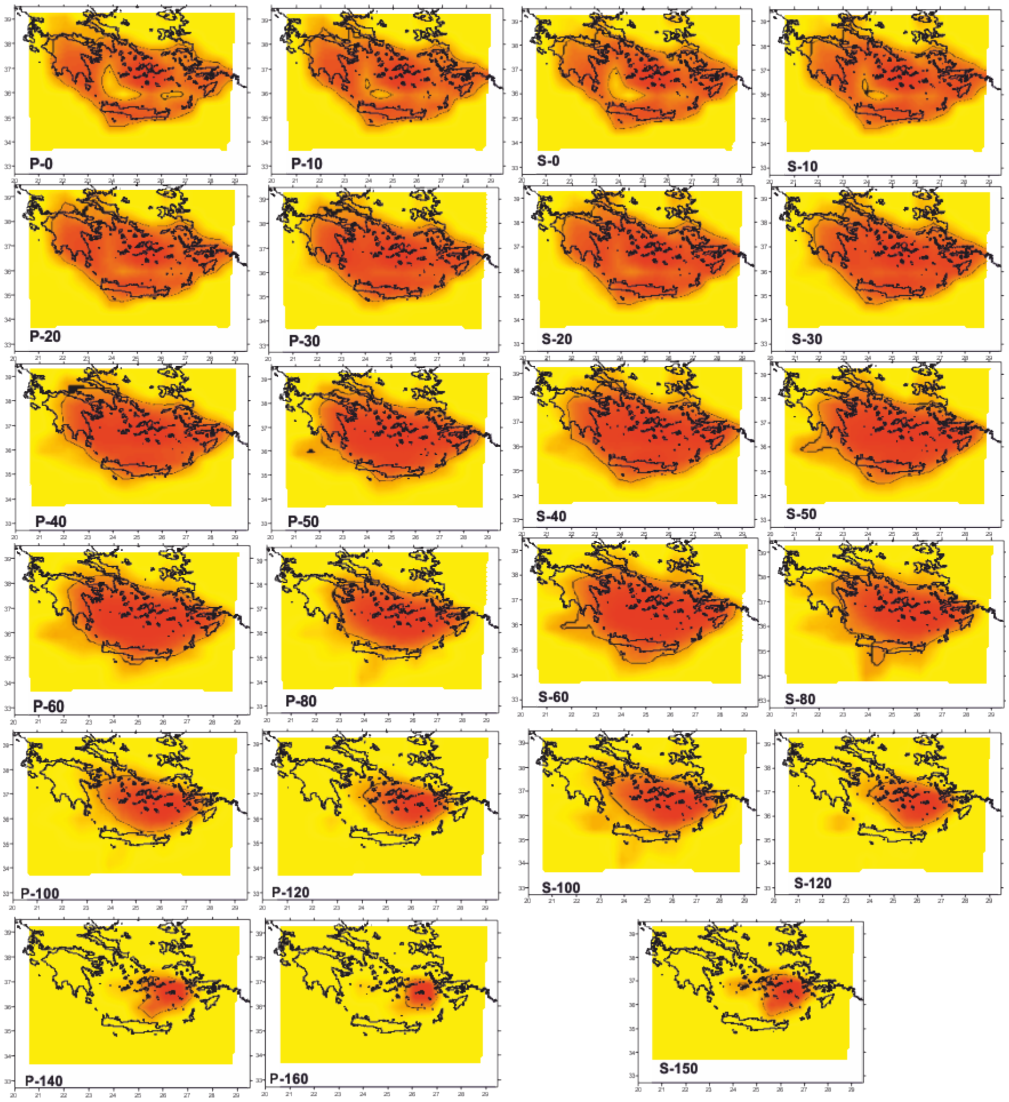

Figure 1. Logarithm of the density of seismic rays calculated by the automatic approach for P and S waves

The system of the linear equations was solved by applying an appropriate code which employs the LSQR method of Paige and Saunders (1982). An initial inversion was performed for the *t** values estimated by the two approaches (automatic and manual) using a one-dimensional background velocity model for the determination of the seismic ray paths (1-D ray tracing). Figure 2 shows the preliminary results from the inversion process for the P-waves attenuation factor ($1000/Q_pV_p$) using attenuation times estimated by the automatic technique. In general, the highest anelastic attenuation is observed for the depth of 60km, with the high-attenuation area covering almost the entire volcanic-arc of the southern Aegean subduction zone. Additional inversion models will be obtained, using different initial velocity models, in order to determine the final three-dimensional attenuation structure model of the Southern Aegean area.

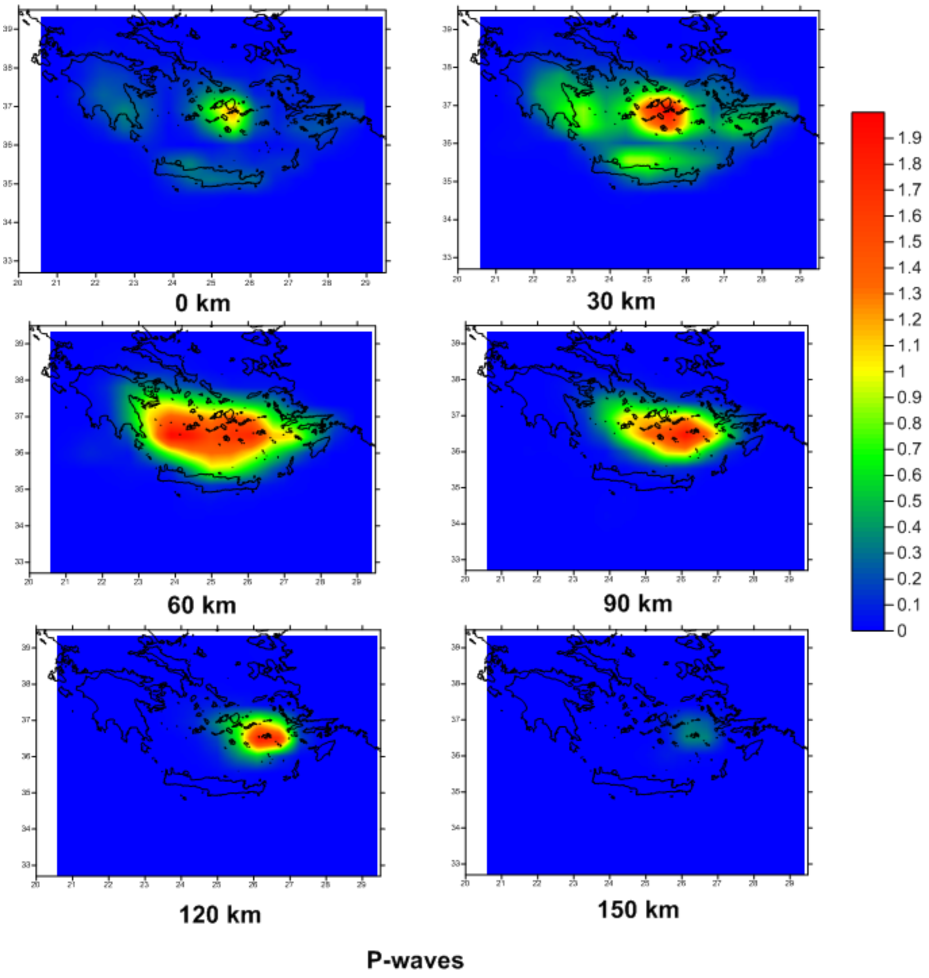

Figure 2. Spatial distribution of the P-waves attenuation factor ($1000/Q_pV_p$) for the depth range 0-150 km using a 1D-background velocity model.

This work has been partly supported by the 3D-SEGMENTS project #1337 funded by EC European Social Fund and the Operational Programme "Education and Lifelong Learning" of the ARISTEIA-I call of the Greek Secretariat of Research and Technology.

REFERENCES

Bohnhoff M, Riche M, Meier T, Endrun B, Harjes H-P, Stavrakakis G, (2004) A temporary seismic network of the Cyclades (Aegean Sea, Greece), *Seismological Research Letters*, 75/3, pp. 352-357

Moser T J, Nolet G and Snieder G (1992) Ray bending revisited, *Bull. Seismol. Soc. Am.*, 82, pp. 259-289

Paige C C and Saunders M A (1982) LSQR: an algorithm for sparse linear equations and sparse least squares, *ACM Trans. Math. Softw.*, 8, pp. 43-71

Papazachos C B and Nolet G (1997) Non-linear arrival time tomography, *Annali di Geofisica*, 40, pp. 85-97,.

Pozgay S H, D A Wiens, J A Conder, H Shiobara, Sugioka H (2009) Seismic attenuation tomography of the Mariana subduction system: Implications for thermal structure, volatile distribution, and slow spreading dynamics, *Geochem. Geophys. Geosyst.*, 10, *Q04X05*.

Stachnik J, Abers G and Christensen D, (2004) Seismic attenuation and mantle wedge temperatures in the Alaska subduction zone, *Journal of Geophysical Research* 109(B10): doi: 10.1029/2004JB003018. issn: 0148-0227.