ATTENUATION TOMOGRAPHY OF FRIULI VENEZIA GIULIA ITALIAN REGION
S. Gentili, F. Gentile
Centro di Ricerche Sismologiche, Istituto Nazionale di Oceanografia e di Geofisica Sperimentale, Udine, Italy

Introduction. The Friuli Venezia Giulia Italian region and the western Slovenia are very peculiar from the tomographic point of view, due to the large lateral variation of velocity and Vp/Vs ratio, that is related to the high level of fracturing and, on the east, to the inhomogeneity of the medium. In this work we carry out for the first time the attenuation tomography of Friuli Venezia Giulia Italian region and western Slovenia. In particular, we analyze the 3D distribution of the frequency independent part of the S-waves quality factor, starting from a dataset of high frequency attenuation parameter k. The spectral decay parameter k was estimated using data from small-to-moderate earthquakes recorded by the NEI (northeastern Italy) seismic network managed by the National Institute of Oceanography and Experimental Geophysics (Istituto Nazionale di Oceanografia e Geofisica Sperimentale, OGS – http://www.crs.inogs.it/). The next section describes the relation between attenuation parameter and quality factor. In the following, the method adopted for k decay parameter estimate is described. Subsequently, the seismicity distribution (historical and instrumental) and details of the application of method tomographic inversion to Friuli Venezia Giulia region are shown. Finally, the obtained results are compared with both the previous k attenuation studies and the tomographic studies on Vp Vs velocities in the same area.

Attenuation parameter vs. quality factor. The attenuation of seismic waves which propagate through the Earth can strongly influence the ground motion recorded at a site thus modifying the energy content of the signal radiated from the source. The attenuation is modelled in literature either using the quality factor \( Q \), a dimensionless parameter introduced to quantify the fractional energy \( \frac{\Delta E}{E} \) loss per cycle of oscillation as \( Q = 2\pi \frac{\Delta E}{E} \) (Aki and Richards, 1980), or using the attenuation parameter \( k \) (e.g. Anderson and Hough, 1984).

The attenuation due to high frequency spectral decay of acceleration amplitude Fourier spectrum has been modelled as (Cormier 1982):

\[
A(r, f) = A_0 e^{-\alpha f}
\]  
(1)

\( A_0 \) depends on the source and the geometrical spreading and \( f \) is the frequency. In this model \( t^* \) is defined as:

\[
t^* = \int \frac{1}{Q V} dr
\]  
(2)

where \( Q \) is the quality factor and \( V \) is the average velocity of the waves. Anderson and Hough (1984) modelled high frequency acceleration spectrum as:

\[
A(r, f) = A_0 e^{-\alpha f}
\]  
(3)

Hough and Anderson (1988) modelled \( k \) as:

\[
k = \int \frac{1}{Q_I(z)V(z)} dr
\]  
(4)

where \( Q_I \) is the frequency-independent part of \( Q \), and \( Q \) is parameterized as:

\[
Q(f)^{-1} = Q_D(f)^{-1} + Q_I^{-1}
\]  
(5)

\( Q_D \) and \( Q_I \) are respectively the frequency dependent and independent part of \( Q \). \( Q_I(z) \) means that in their simplified model \( Q_I \) depends only on the depth (plane and parallel layers) but can be generalized. Hough and Anderson (1988) noted that the model they used for \( k^* \) is...
the same of Cormier (1982) with the difference that only frequency independent part of $Q$ is considered. In the works on $k$ estimation the term $Q_D$ is generally considered (Anderson and Hough, 1984, Castro et al., 1997, Franceschina et al., 2006, Bressan et al., 2007). Vice versa, when $r^*$ is estimated for attenuation tomography inversion, the frequency dependent term is usually neglected (Rietbrock 2001, Haberland and Rietbrock 2001, Olsen et al., 2003, Eberhart-Phillips et al., 2005). The opportunity to considering or not the $Q$ dependence on frequency is debated. The reason for neglecting it is the difficulty in the evaluation of this term for the whole area under study and/or the lower dependence of the quality factor on frequency for higher frequencies. The latter reason is debated. Several different authors analysed the area under study in this paper and estimated the frequency dependence of $Q$. Console and Rovelli (1981) obtained a dependence of $-f^{1.1}$ in the frequency range 0.1-10 Hz, Castro et al. (1996) obtained a dependence of $-f^{1.1}$ in the frequency range 0.4-25 Hz, Malagnini et al. (2002) obtained a dependence of $-f^{0.55}$ in the frequency range 0.5-14 Hz. Estimates of $k$ in the study area of this paper were performed by Franceschina et al. (2006) and Bressan et al. (2007) considering a term $Q_D=78 f^{0.96}$ [according with Govoni et al. (1996) in the frequency range 1-25 Hz]. In a successive paper, Gentili and Franceschina (2011) considered the relation $Q_D(f) = 251 f^{0.7}$ correspondingly to the total quality factor estimated for this area by Bianco et al. (2005) in the frequency range 0.5-16 Hz. In this work, for coherence with previous ones in the same area, we will not neglect the frequency dependence of $Q$. In particular, we will adopt Bianco et al. (2005) estimation, like Gentili and Franceschina (2011). Considering such correction does not affect too much the value of $Q$ for small values of $Q$. For example, considering the value of $Q_D$ at a frequency in the middle of our spectral window (Eberhart – Phillips et al., 2005), say 15 Hz, for $Q_i=20$ we have $Q=19.8$. For very low attenuation, the effect of considering frequency dependence is larger: e.g. with $Q_i=1000$ we have $Q=626$. For coherence with previous papers notation, we will call the frequency independent part of the quality factor for $S$ waves simply $Q_S$, instead of $Q_{IS}$.

**Attenuation parameter estimation.** Assuming a negligible dependence of the geometrical spreading on frequency, we estimated the parameter by correcting the S-wave spectra for the frequency dependent part of the quality factor. Both the N-S and the E-W horizontal components of the signal were used. The S wave window was manually selected for both traces and the resulting data were tapered by a 5% cosine taper and padded with zeros before applying the FFT. The resulting spectra were smoothed by a sliding Hann window (Oppenheim and Schafer, 1999) of 0.5 Hz half-width. The spectral band adopted for the analysis was determined by selecting the part of the spectrum where a linear decay was clearly evident. In particular, $S$ wave window was manually selected for both traces and the resulting data were tapered by a 5% cosine taper and padded with zeros before applying the FFT. The resultting spectra were smoothed by a sliding Hann window (Oppenheim and Schafer, 1999) of 0.5 Hz half-width. The spectral band adopted for the analysis was determined by selecting the part of the spectrum where a linear decay was clearly evident. In particular, $S$ wave window was manually selected for both traces.

In order to fit the amplitude spectrum, we used the least absolute residual method. This method minimizes the absolute difference of the residuals rather than the squared differences, thus decreasing the influence of outliers on results. The $k$ values estimated from the two horizontal components of each record were averaged by computing the corresponding weighted mean and standard deviation. In order to increase the robustness of the method, the weights were chosen as the inverse of the error of each single fit.

**Seismicity distribution and tectonics of the area.** The area we analysed is sited in northeastern Italy (Friuli Venezia Giulia region) and western Slovenia between the outer front of
Southern Alps and the Periadriatic lineament. It is included in a convergent margin zone between the Adria microplate and the Eurasian plate (see e.g. Castellarin et al., 2006; Burrato et al., 2008) and it is one of the most seismic active zones of Italy.

The area is sited at the tip of Adria plate. The progressive north-eastward pushing of African plate, respect to the Eurasian plate, generates an anticlockwise rotation of Adria microplate, which causes the present complex tectonic deformation (Mantovani et al., 1996). Several tectonic phases in the region inherited and reactivated the main pre-existing faults and fragmented the crust into different tectonic domains corresponding to different seismotectonic zones (Bressan et al., 2003). The eastern part of the region, in western Slovenia, is characterized by a strike-slip regime on Dinaric faults. The Friuli Italian region is characterized by thrust tectonic, mainly EW oriented and south verging in the centre and NE-SW-oriented and SE-verging thrust in the western part (Bressan et al., 2003).

Fig. 1 shows both the historical and instrumental seismicity in the area. Brown squares correspond to earthquakes in CPTI04 catalogue (Gruppo di lavoro CPTI, 2004) from 1100 with intensity ≥IX. The other symbols correspond to instrumentally recorded seismicity. In particular, stars correspond to the events with magnitude ≥5 in the area. See figure caption for more details. Red symbols are the earthquakes used in this work. They are the 156 earthquakes with duration magnitude ≥ 3 recorded in the area from 1994 to 2011. In particular, a cluster of seismicity can be easily detected in the northeastern part in correspondence with the two Kobarid 1998 and 2004 seismic sequences. Smaller clusters of lower magnitude seismicity were recorded also after Sernio 2002 mainshock and during and in Claut region (for seismic sequences in the area see Gentili and Bressan, 2008). The instrumentally recorded seismicity roughly follows the historical earthquakes distribution.

Fig. 1 – Seismicity in the area. Brown squares: historical seismicity (earthquakes with intensity ≥IX are shown with squares dimensions proportional to earthquake intensity). Red symbols: seismicity with magnitude >3 from 1994 to 2011. Stars: earthquakes with magnitude > 5. Yellow stars: 1976 swarm. Green star: 1977 Trasaghis earthquake. Blue star: 1979 Lusevera earthquake. Red stars: Kobarid and Sernio earthquakes. Dashed lines: Sections shown in Fig. 3.
Attenuation tomography in Friuli Venezia Giulia region. To invert the $k$ data we used the same grid used by Bressan et al. (2012) in 3D tomographic inversion of velocity in the area. The grid extends 114 km in E-W direction and 55 km in N-S direction. The grid centre has latitude 46.33 N and longitude and 13.08 E. The W-E grid nodes are at X: -60, -50, -35, -25, -15, -7, 0, 7, 15, 25, 36, 45, 54 km; the S-N grid nodes are at Y: -35, -20, -10, -5, 0, 5, 10, 20 km; the Z nodes are at depth 0, 2, 4, 6, 8, 10, 12, 15, 22 km with a layer at negative 3 km depth to account for the Earth’s topography. The X-axis is positive to the east, the Y-axis is positive to the north. The velocity of S waves in the medium is parameterized by assigning velocity values obtained in Bressan et al. (2012) paper at the nodes of the grid. Bressan et al. used 394 events from 1988 to 2004 with duration magnitude between 1.4 and 5.1 and performed an iterative simultaneous inversion of hypocentral parameters and 3D velocity structure with a damped least squares technique using a previous version of the code SIMULPS2000, i.e. SIMULPS12 (Evans et al. 1994). We used 156 events from 1994 to 2011 with duration magnitude between 3.0 and 5.7 for a total of 980 3-D records. The smaller number of earthquakes analyzed in this study is due on the constraints on the minimum magnitude we selected.

Like in tomographic applications in which velocity is inverted, an adequate starting model for the inverted parameter is important before attenuation tomographic inversion, due to the non-uniqueness of the solution of linearized inverse problem. A wrong starting model can lead to blunders and biases in the result. No previous knowledge for a valuation of the $Q_s$ field in the Friuli area are available; so for finding the 1D starting crustal model a search heuristic was employed, that does not use linearization method. We adopted a genetic algorithm technique (Goldberg, 1989). In particular, the David L. Carroll GAFortran code with MICRO-GA enabled (Carroll, 2001). A micropopulation of five individuals was utilized using as cost function the same SIMULPS2000 code minimizing the total weighted RMS; 600 generations were generated.

Fig. 2 – Checkerboard test for depth 4-10 km. Circle size correspond to the percentage difference respect to the mean value.
Starting from the previous 1D $Q_s$ Earth model we performed the 3D inversion using the program SIMUL2000 (Thurber, 1993; Eberhart-Phillips, 1993, Thurber and Eberhart-Phillips, 1999), starting from the values of $k$. The code applies an iterative damped least squares method. In the inversion, as suggested in SIMUL2000 code, the earthquakes hypocenters and origin times were kept fixed. The $Q_s$ crustal values are inverted using a damping value of 0.002, chosen evaluating the trade-off curve of the data variance versus model variance achieved after single iterations (Eberhart-Phillips, 1986, 1993).

**Results.** A refined estimate on the data reliability was performed through the comparison between the checkerboard test and a spread function test (Michelini and McEvilly, 1991). We performed the checkerboard test with a mean $Q_s$ of 200 with variations of $\pm 40\%$; we added to synthetic $t^*$ a noise comparable to the residual level of real data. We choose such a value because it’s near to the RMS obtained by the genetic run (Chen and Clayton, 2012). Fig. 2 shows the checkerboard test, for depth 4-7 km, superimposed an isoline of the spread function $S$ value of 7.

Fig. 3 – Vertical cross sections 3-5 of the 3D $Q_s$ distribution.
Fig. 3 shows the resulting $Q_s$ for sections 3, 4, 5. The earthquakes are indicated as circles whose size corresponds to the magnitude. On the generic nth section characterized by y coordinate equal to $y_n$, the earthquakes corresponding to y in the range $[(y_{n+1} + y_n)/2$, $(y_{n+1} + y_n)/2]$ are projected. A cluster of earthquakes can be detected in the rightmost area of sections 4 and 5, corresponding to the Kobarid sequences. In the same area, a large decrease of the value of $Q_s$, and therefore a larger attenuation, can be detected, in good agreement with the results by Gentili and Franceschina, 2011. Another cluster of seismicity can be detected in section 5 on the left, with a depth between 3 and 15 km. This seismicity belongs to the Claut area, where several low moderate magnitude earthquakes happened in the past (Gentili and Bressan, 2008).

To the right of the cluster, there is on the top a region characterized by high values of $Q_s$ (low attenuation) and on the bottom with low value of $Q_s$ (high attenuation). Gentile et al. (2000) detected in the same region on the top an area of low $V_p/V_s$ respect to the surrounding (1.76) and high $V_s$ velocity (6.2). Considering their results, together with the high value of $Q_s$ obtained in this paper, we hypothesize that it can correspond to un-fractured dolomitic limestones path, inside a more fractured one. On the bottom, Gentile et al. (2000) detect an area of $V_p/V_s$ in the range 1.84-1.89. This region is interpreted by Gentile et al. (2000) as an highly fractured zone that corresponds to the Tramonti-But Chiarsò fault. This is compatible with our low attenuation results, even if the $Q_s$ anomaly extends deeper than the $V_p/V_s$ one and, unlike for the high $V_p/V_s$ zone, does not reach the surface, but is interrupted by a low attenuation zone.

Both Bressan et al. (1992) and Gentile et al. (2000) detected an high P waves velocity body in the central part of the area with depth between 4 and 8 km. This is coherent with the detected high values of $Q_s$ (1100-1500). The even higher values of $Q_s$ at 15-18 km depth is not well resolved by both this work and by Bressan et al. (1992) and Gentile et al. (2000) works.

References


Gentili S. and Bressan G.; 2008: *The partitioning of radiated energy and the largest aftershock of seismic sequences occurred in the northeastern Italy and western Slovenia*. J. Seismol., 12, 343-354.


