Sequential Integrated Inversion of tomographic images and gravity data: an application to the Friuli area (north-eastern Italy)

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ABSTRACT

The three-dimensional pattern of elastic moduli (bulk modulus, Young modulus, shear modulus) of the upper crust (0-10 km depth) has been determined in the Friuli area (north-eastern Italy) from the 3D $V_p$, $V_p/V_s$ and density structures. Firstly, 3D P-wave velocity and P to S velocity ratio were modeled by joint inversion for hypocentres and velocity structure. Then, we apply the tomographic inversion method of Sequential Integrated Inversion (SII) to recover the three dimensional density structure. The pattern of the elastic moduli is characterized by marked lateral and depth variations that reflect the geologic-structural heterogeneity of the area, produced by the superposition of several tectonic phases with different orientations of the principal axes of stress. The bulk ($K$), Young ($E$) and shear ($G$) moduli image a high rigidity body with an irregular shape, at 4-8 km depth. The body is characterized by $G \geq 3.2 \times 10^{10} \text{ N}\cdot\text{m}^{-2}$, $K \geq 6.8 \times 10^{10} \text{ N}\cdot\text{m}^{-2}$ and $E \geq 8.4 \times 10^{10} \text{ N}\cdot\text{m}^{-2}$ and is associated to platform limestones and dolomitic rocks. The seismicity is mainly located along the sharp variations of the moduli pattern, in or adjacent to high rigidity zones. The most severe earthquakes ($M_L$ between 4.5 and 6.4), occurred in the study area from 1976 to the present day, are located in a transition zone from high to low rigidity patterns. Our interpretation is that the elastic moduli variations, closely related to variability in rock mechanical properties, influence the occurrence of earthquakes by processes of stress concentrations. The values of the elastic moduli recently obtained from laboratory measurements on the main lithologic units fall in the middle-high range of the values obtained with the present investigation.

Keywords: Seismic tomography, gravity anomalies, seismicity, elastic moduli, Friuli, NE Italy.

1. Introduction

The local earthquake tomography (LET) has been widely used to image the crustal heterogeneities of the seismogenic zones. Since P and S wave velocities provide some measure of the rock stiffness, the tomographic images allow to investigate the relation between the mechanical properties of the crust and the occurrence of earthquakes. However, velocities of similar value may happen to characterize different composition of rocks and the velocity pattern alone cannot be sufficient to image the variable mechanical properties of the medium. The elastic
moduli can be effective in imaging the heterogeneous mechanical properties of earthquake source zones. Hence, the purpose of the present paper is to determine the pattern of the bulk, Young and shear moduli from 3D $V_p$ and $V_p/V_s$, tomographic inversions integrated with gravity modelling, in the Friuli area (north-eastern Italy) and part of north-western Slovenia. The elastic moduli images are then compared with the seismicity pattern of relocated earthquakes, to investigate the seismogenic characteristics. If correctly interpreted, gravity data offer significant information about density heterogeneity; the difficulty lies in the inherent non-uniqueness of gravity data inversion. Therefore, some form of regularization must be adopted to obtain physically relevant and stable solution. A useful approach can be represented by a method that integrates seismic and gravity information within a likelihood function. The method, developed by Tondi and de Franco (2006) is based on a likelihood function which includes: a) information on the gravity data; b) information given by the velocity parameter adjustment vector resulting from the set of traveltimes inverted to model velocities; c) information on the physical correlation among density and velocity parameters; d) information on the error propagation from the velocity to the density model. The likelihood function gives a measure of how good the starting model is in explaining the data (Tarantola, 2005). Hence, the update to the density model is searched by maximizing this function with respect to the density parameter.

Past studies (Gentile et al., 2000; de Franco et al., 2004) on this area have used traditional inversion of seismic and gravity data with the elaboration of independent models. These studies regarded an area less extended and the acceptable solutions are attributable to restricted zones and depths. The goal of the present investigation is to obtain a more detailed and reliable geophysical model of the area with the use of the new approach Sequential Integrated Inversion (SII) of integrated seismic and gravity data inversion. Laboratory data and sonic log data from a borehole in the area (Faccenda et al., 2007), not available in the past studies, are very valuable for the calibration of physical properties of rocks and to improve the geophysical interpretation.

2. Geological framework

The investigated area is characterized by a poliphase deformational zone (Fig. 1), resulting from the superposition of several Cenozoic-age tectonic phases (Venturini, 1991). The NE-SW Dinaric compression was active during the middle-late Eocene and generated NW-SE oriented thrusts, affecting mainly the central, eastern and south-eastern parts of the study area. From middle Miocene to earliest Pliocene, the N-S trending Alpine compression produced E-W oriented systems of south-verging thrusts and backthrusts, with severe shortening of the upper crust in the central part of the area. During the Pliocene times, a NW-SE oriented compression produced mainly NE-SW trending thrusts and folds. Each tectonic phase inherited and re-activated the geological deformations of the previous phase producing a complex deformation pattern and different tectonic domains, which form at present different seismotectonic zones (Bressan et al., 2003).

The rock stratigraphy (Slejko et al., 1989) consists of sedimentary rocks of Paleozoic to Quaternary age (Fig. 1). Flysch and molasse compose the Cenozoic and Quaternary deposits. The dominant geologic units of the investigated area are limestones and carbonatic rocks of Mesozoic age. The Paleozoic rocks are mainly composed by sandstones, limestones and locally volcanic deposits. The severe shortening of the N-S trending Alpine compression caused the detachment
of the Mesozoic sedimentary cover from the Paleozoic geologic units. The related thrusting produced alternations and repetitions of the Triassic dolomitic rocks and Jurassic limestones (Carulli and Ponton, 1992).

3. Tomographic inversion

The $V_p$ and $V_p/V_s$ images of the upper crust of the study area are obtained with local earthquake tomography (LET), based on the inversion of local earthquake P and S wave arrival times (Thurber, 1983). The iterative simultaneous inversion of hypocentral parameters and 3D
velocity structure, with a damped least squares technique, is performed with the SIMULPS algorithm of Evans et al. (1994). A 1D initial velocity model is assigned to the nodes of a 3D grid. The P and S arrival times are then inverted for earthquake location, \( V_p \) and \( V_p/V_s \) variations. The travel times from hypocentre to station are calculated with an approximate ray tracer, based on pseudo-bending (Um and Thurber, 1987).

The investigated area is represented by a grid (Fig. 2) extended 114 km in the W-E direction (W-E grid nodes at \( X = -60, -50, -35, -25, -15, -7, 0, 7, 15, 25, 36, 45, 54 \) km) and 55 km in the S-N direction (S-N grid nodes at \( Y = -30, -20, -10, -5, 0, 5, 10, 20 \) km). X is positive to the east, Y is positive to the north. The centre of the grid has latitude 46°20’N and longitude 13°05’E. The grid spacing is finer in the central part to account for heterogeneous ray coverage. The depth grid spacing is \( Z = 0, 2, 4, 6, 8, 10, 12, 15, 22 \) km. A layer at negative 3 km depth is included to account for the Earth’s topography.
We use in the inversion 394 events (Fig. 2), occurred from 1988 to 2008, initially located with the HYPO71 program (Lee and Lahr, 1975), with largest azimuthal separation between stations (GAP) less than 180°. The coda-duration magnitude $M_D$ (Rebez and Renner, 1991) ranges from 1.4 to 5.1. The picking accuracy is estimated about 0.05 s for P waves and within 0.1 s for S waves. The data set used for the 3D $V_p$ and $V_p/V_s$ inversion consists of 4337 P arrival times and 3480 S arrival times.

The seismograms from 1988 to 2008 were recorded by 12 digital short-period stations with sampling rate 62.5 sps, 5 digital short-period stations with sampling rate 125 sps (2 operating since 2002), of the Istituto Nazionale di Oceanografia e di Geofisica Sperimentale (OGS). Furthermore, we use the records of 4 broad band stations of the Seismological Survey of Slovenia, with sampling rate of 200 sps, operating since 2004. In the tomographic inversion, we include also the arrival times of the 1998 and 2004 Bovec-Krn sequences, occurred in western Slovenia, recorded by locally temporary stations, characterized by a sampling rate of 125 sps. During the 1998 sequence, 6 broad-band and 4 strong motion stations were operating (Živčič et al., 2000; Bajc et al., 2001). The data of the 2004 sequence were recorded by 5 strong motion stations and 1 short-period seismometer (Živčič, personal communication) and by 2 short-period temporary stations of the OGS network.

The tomographic images are obtained by computing at first a reliable 3D $V_p$ model, that is used as basis for inverting $V_p/V_s$ values. The 3D $V_p/V_s$ data were inverted keeping fixed the $V_p$ values as obtained from the 3D $V_p$ model and allowing hypocentre relocation. This procedure (Miller and Smith, 1999; Husen et al., 2000; Kaypak, 2008) is adopted because the S-wave data are fewer in number and the S-wave pickings have major uncertainties than the P-wave data. Non-uniform distribution of P and S data cause different sampling and ray coverage in the investigated crust. These differences do not allow a simultaneous inversion of P velocity and $V_p/V_s$ ratio structure, since the larger number of P-wave phases could mask significant $V_p/V_s$ variations. A similar approach is also used for modelling the crustal seismic properties from active sources along seismic arrays (Mjelde et al., 2003). Thurber (1993) showed that the inversion of 3D $V_p/V_s$, keeping the $V_p$ fixed, can provide faithful images of significant $V_p/V_s$ variations, that are useful for better constraining the mechanical properties for geological interpretations.

The 1D $V_p$ initial model (Table 1) is defined by considering the models of Scarascia and Cassinis (1997) and of Brückl et al. (2007), who evidenced large-scale discontinuities in the crustal P-wave velocity structure at 10, 15 and 22 km depth. The geological cross-sections of Carulli and Ponton (1992), Merlini et al. (2002) and Poli et al. (2002) are used as reference for the geometry and thickness of the sedimentary units (0-12 km depth).

Scarascia and Cassinis (1997) and Brückl et al. (2007) found $V_p$ values of about 6.3 and 6.4 km s$^{-1}$ in the crustal upper layers at depth less than 10 km. A decrease of P-wave velocity from 6.3 to 5.9 km·s$^{-1}$ is recognized at about 10 km depth (Scarascia and Cassinis, 1997). The 1D $V_p$ and $V_p/V_s$ initial models of the upper layers is defined by considering the best resolved tomographic images of Gentile et al. (2000) and Bressan et al. (2009), the P-wave velocities obtained with sonic log data from a borehole in the area and the $V_p$ and $V_s$ laboratory measurements (Faccenda et al., 2007). The field and laboratory data include the dominant lithologies of the upper crust of the study area.

The values of damping for the $V_p$ and $V_p/V_s$ inversion are selected according to Eberhart-Phillips (1986) by evaluating the trade-off curves between the data and the solution variance. The $V_p$ data are
inverted using damping 10. The 3D rms residual is 0.138 s. The 3D $V_p/V_s$ model is inverted with damping 10, using the 3D $V_p$ model as basis and allowing hypocentre relocation. The 3D rms residual is 0.246 s. Figs. 3a to 3f show the $V_p$ and $V_p/V_s$ images at six depth slices. On the maps are plotted also other 1129 events, occurred from 1988 to 2008 and relocated using the obtained tomographic model. The overall earthquakes plotted are 1523 with $M_D$ ranging from 1.4 to 5.1.

The uneven distribution of both the seismic stations and the events in the study area and the different recording periods of some stations influence the quality and the accuracy of the solution. Eberhart-Phillips and Reyners (1997) claimed that a node is adequately resolved if its resolution is peaked and has not meaningful contribution from nodes not adjacent. The spread function, SF (Michelini and McEvilly, 1991) summarizes the information contained in the resolution matrix and indicates how peaked the resolution is for a node. The spread function represents a succinct way of assessing the resolution because it takes into account all the elements of the resolution matrix. Following Reyners et al. (2006), SF values less than 2.5 indicate that the tomographic model is representative of the volume surrounding the given node. SF values between 2.5 and 3.0 characterize nodes with acceptable resolution, but the velocity may be averaging a larger volume. Nodes with SF values between 3.0 and 3.75 show meaningful velocity patterns, but the size of the velocity variations may be smaller than the actual velocity heterogeneity. Nodes poor resolved are characterized by larger SF values. Figs. 4a to 4f and Fig. 5a to 5f show the SF pattern of the final 3D $V_p$ and $V_p/V_s$ models, respectively. We chose the threshold SF < 3.0 below which the velocity anomalies can be considered adequately resolved.

The resolution patterns of the $V_p$ and $V_p/V_s$ tomographic images appear quite similar. The resolution is poor at 0 km depth because of uneven sampling of ray paths. At 2 km depth, the resolution appears patchy with best resolved zones in the central and eastern part of the grid. The layers at 4 and 6 km depth show best resolved nodes in the X distance about from -25 to 45 km and in the Y distance about from -10 to 20 km. The central part of the grid appears best resolved at 8 km depth within X distance about between -25 and 25 km and Y distance about between -10 and 10 km. The resolution decreases at 10 km depth because of minor number of earthquakes and, consequently, relative poor sampling of nodes by the ray paths.

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>$V_p$ model (km/s)</th>
<th>$V_p/V_s$ model</th>
</tr>
</thead>
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<td>-3.0</td>
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<td>1.90</td>
</tr>
<tr>
<td>0.0</td>
<td>5.50</td>
<td>1.88</td>
</tr>
<tr>
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<td>5.90</td>
<td>1.86</td>
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</tr>
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</tr>
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<td>6.00</td>
<td>1.85</td>
</tr>
<tr>
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<td>6.10</td>
<td>1.83</td>
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<td>6.40</td>
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</tr>
<tr>
<td>22.0</td>
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<td>1.77</td>
</tr>
<tr>
<td>40.0</td>
<td>8.00</td>
<td>1.80</td>
</tr>
</tbody>
</table>
Fig. 3 - Tomographic images at depth slices 0 (a), 2 (b), 4 (c), 6 (d), 8 (e), 10 (f) km. The $V_p$ values are shown as contour lines, the $V_p/V_s$ values are plotted in graded colours. Diamonds: relocated earthquakes (see text at chapter “Tomographic inversion”).
4. Tomographic images

The \( V_p \) and \( V_p/V_s \) images (Figs. 3a to 3f) are characterized by a complex pattern with marked lateral and depth variations that we relate mainly to the lithological heterogeneity and to the different degree of fracturing. The seismicity is mostly located along the sharp variations of high...
$V_p$ in comparison with the surroundings and marked changes of the $V_p/V_s$ anomalies. Shallow molasse and terrigenous deposits (flysch), outcropping in the south-eastern sector of the study area and lying at 0-4 km depth, are characterized by low P-wave velocities ($V_p = 5.3-5.8$ km·s$^{-1}$) and high $V_p/V_s$ values (1.81-2.00). The rock stratigraphy of the investigated crust consists mainly...
of Mesozoic rocks (Fig. 1). We relate the P-wave velocities ranging from 5.5 to 6.3 km s\(^{-1}\) and the \(V_p/V_s\) values between 1.79 and 1.87 to Cretaceous and Jurassic limestones. A wide range of P-wave velocities (6.0-6.8 km s\(^{-1}\)), with high values (6.4 to 6.8 km s\(^{-1}\)) associated to platform limestones and dolomitic rocks, characterizes the Triassic sedimentary rocks. The \(V_p/V_s\) anomalies related to Triassic rocks vary from 1.75 to 1.95. The Paleozoic deposits, lying below 6 km depth in the northern sector of the investigated area, are mainly made up of sandstones and limestones and can be characterized by \(V_p\) ranging from 5.8 to 6.3 km s\(^{-1}\), with \(V_p/V_s\) values 1.78-1.93.

The tomographic images show in the central part of the area, from 4 to 8 km depth, a high rigidity body, characterized by \(V_p\) 6.4-6.8 km s\(^{-1}\), with significant velocity contrasts respect to the northern and southern zones. We relate the high velocity anomaly to the Triassic platform limestones and dolomitic rocks, as resulted also by the \(V_p\) data obtained by sonic-log technique performed by the ENI-AGIP company in the well Cagnacco 1 (Venturini, 2002), located close to the Udine town (Fig. 1). The \(V_p/V_s\) values range between 1.77 and 1.92. Domenico (1984) showed that the \(V_p/V_s\) ratio is sensitive to the rock lithology and reported from laboratory measurements the ranges 1.78-1.84 for dolomite and 1.84-1.99 for limestones. However, we deem that the wide range of \(V_p/V_s\) values is partly due to the variation in lithology between limestones and dolomitic rocks. Since the \(V_p/V_s\) ratio is mostly influenced by the density of cracks and the pore pressure (O’Connel and Budiansky, 1974; Tatham, 1982), we attribute the variability of \(V_p/V_s\) values also to the different level of cracking and fluid pressure. The shape of the high velocity body is not regular and it extends from the central part of the area to western Slovenia. High \(V_p\) spots are found also in the western part of the investigated area.

Generally, the pattern of the \(V_p/V_s\) images is heterogeneous, with spots of high and low \(V_p/V_s\) anomalies mostly located in the central and eastern part of the area. The \(V_p/V_s\) ratio appears sensitive to the rock lithology but largely influenced by the degree of fracturing and therefore by the mechanical strength. As mentioned above, we point out that the area is highly tectonized. This area is characterized (Carulli and Ponton, 1988) by maximum interference and overlapping of Alpine E-W oriented and Dinaric NW-SE oriented faults (Fig. 1) and by the most severe shortening (Venturini and Carulli, 2002) occurred during the Cenozoic-age tectonic phase. The severe shortening of the Mesozoic cover, caused by thrusting, produced frequent repetitions and thickening of the sedimentary units. The variability of P-wave anomalies are partly due to the large rigidity variation between platform limestones/dolomities and the other lithologies. The large lateral heterogeneities of \(V_p\) and \(V_p/V_s\) tomographic images reflect the complexity of the various tectonic domains, characterized by different fracture patterns and mechanical properties. The results confirm the main features emerged in Gentile et al. (2000): a high rigidity body recognizable in the central and eastern parts of the area, lying between 4 and 8 km depth and the high mechanical heterogeneity of the upper crustal layers. The tomographic images of the present study appears more detailed and the \(V_p\) and \(V_p/V_s\) values are better constrained for the related lithologies.

5. The 3D density model

The velocity update (\(\Delta v\)) between the starting 1D velocity model and the seismic tomography (ST) model is used as input for the 3D seismo-gravity integrated inversion (Tondi and de Franco,
Gravity constraint is given using the Bouguer anomaly gravity digital data of the Italian National Gravity Database (Cassano et al., 1989), which have an areal sampling of $3 \times 3$ km. For the inversion, 1720 points of Bouguer anomalies were used.

Relationships between the P-wave crustal velocities and density values are based on $V_p$-density depth profile, elaborated from laboratory data in the Friuli area (Gentile et al., 2000). The relations used are $\rho = 221.6 \ V_p + 1334.5$ for velocities up to 7 km s$^{-1}$ and $\rho = 401 \ V_p + 77.2$ for higher values. The relationship $\rho = 155.1 \ V_p + 2147.1$, taken from 3SMAC global density Earth model (Nataf and Ricard, 1996), is used for the upper mantle, down to 40 km depth. Densities are expressed in kg m$^{-3}$.

The 1D starting density model (Fig. 6), which is used as reference model, is partitioned in 1260 rectangular prisms: 840 internal bodies and 420 external bodies to account for edge effects. The rectangular prisms are characterized by a point origin $\rho_0$ (the node on the bottom left side) and three density gradients in the x, y, z directions with respect to this point [Eq. (15) in Tondi and de Franco (2006)]. The expression derived by Pohanka (1998), for a polyhedral body whose density is linearly dependent on the body coordinates [Appendix C in Tondi and de Franco (2006)], is used for the computation of the three components of the gravity field. The use of polyhedral bodies with linearly varying density allows us to appropriately match the velocity gradients with the density gradients during the seismic gravity modelling. Analogously, as in seismic modelling, we do not fix an a priori Moho depth variation; an approximate Moho signature emerges from the inversion of data sets.

Taking into account the downward extension of the model, the gravity response of the study area is calculated using a background density of 3100 kg m$^{-3}$. Residuals $\Delta g$ between the observed (Fig. 7) and the calculated field are used as input for the recovery of the density parameter adjustment vector $\Delta \rho$ [Eq. (1) in Tondi and de Franco (2006)].

![Fig. 6 - Reference density model used as starting model for the SII optimization process.](image)
\[ \Delta \rho = \left( \left( G^T C_{gg}^{-1} G + C_{mm}^{-1} \right)^{-1} \left( G^T C_{gg}^{-1} \Delta g + \alpha C_{mm}^{-1} \Delta v \right) \right) \] (1)

with \( \Delta v \) the velocity parameter adjustment vector resulting from the seismic tomographic inversion; \( G \) the matrix containing the geometric gravitational coefficients relating each node to each gravity measurement; \( C_{gg} \) the diagonal matrix of the a priori gravity data uncertainties which are set equal to 0.5 mGal for all measurements.

The \( C_{mm} \) matrix takes into account the error propagation from the velocity to the density model. The error propagation is related to the a posteriori uncertainty in \( V_p [\sigma(V_{pm})] \) and to the uncertainties in the coefficients \( \alpha \) and \( \beta \) of the linear \( \rho-V_p \) relationship:

\[ \sigma(\rho_m) = V_{pm} \sigma(\alpha) + \alpha \sigma(V_{pm}) + \sigma(\beta) \quad 1 \leq m \leq M \] (2)

where \( M \) is the number of model parameters.

As we do not have information on the errors in the coefficients (\( \alpha \) and \( \beta \)) that describes the \( \rho-V_p \) linear relationship, the diagonal elements of the \( C_{mm} \) covariance matrix are completely determined by the a posteriori seismic covariance matrix, resulting by the seismic tomography inversion and converted in density using the velocity-density relationships \([\sigma(\rho)=\alpha \sigma(v)]\). A mean density value of 200 kg m\(^{-3}\) is assigned to this matrix in correspondence of null values of seismic hit counts. The \( C_{mm} \) matrix for the first seven layers (0-12 km depths) has a mean density covariance of about 63 kg·m\(^{-3}\) with an average standard deviation for each layer of about 7.5 kg·m\(^{-3}\). The a priori percentage error on the estimated density for the first seven layers is about 10.5%, which is calculated averaging the density value and the corresponding density covariance for each layer.

By appropriately setting the variances on the \( C_{mm} \) matrix, we control the size of parameter adjustments (\( \Delta \rho_m \)) and accordingly, of the minimization of gravity data misfit.

As explained in Tondi and de Franco (2006), the gravity data residuals produced by the optimized density model, give also the possibility to check the quality of the chosen \( \rho-V_p \) linear relationship. Where the relationship overestimates the density values we can observe negative gravity discrepancies (\( g_{OBS} - g_{CALC} \)); on the other hand underestimation brings positive gravity misfit.

Hence, we proceed with a SII and the gravity field produced by the new density model (Fig.

Fig. 7 - Bouguer anomaly map of the study area with areal resolution of 3 × 3 km.
Fig. 8 - Depth slices of the percentage anomalies of the final density model with respect to the reference model.
Gravity residuals reproduced by this 3D density structure (Fig. 9) reveal that Bouguer anomalies are well fitted within the limits of 6 mGal. As discussed above, the positive and negative sign of residuals identify those regions where the chosen $\rho - V_p$ linear relationship should be slightly changed to fit the real structures. Only an anomalous high spot of 12 mGal is recognized at 13.43°E 46.18°N. This anomaly in the misfit is connected to the gravity response of the prisms belonging to the first two layers (0 and 2 km) in this position, for which we have a loss in the seismic ray coverage that corresponds to the a priori $C_{mn}$ value of 200 kg m$^{-3}$.

The update to the density model, $\Delta \rho$, is transformed in the new update to the velocity model through the linear differentiate relationship previously adopted ($\Delta v_{SII} = \Delta \rho / \alpha$) and a new velocity model (referred to as $SII_{vel}$ hereinafter) is obtained.

The statistic parameters are reported in Table 2, showing the difference between seismic tomography (ST) input velocity model and the SII model for the first seven layers. In general, mean and standard deviation of the seismic and SII models are quite similar. The percentage velocity variations do not exceed about 6%. This is also confirmed by the RMS travel time residuals calculated with $SII_{vel}$ model (0.154 s$^{-1}$), similar to those obtained with the seismic tomography (ST) model (0.138 s$^{-1}$).

Nevertheless, the consistent reduction of gravity residuals and statistic parameters (Fig. 9 and Table 2) evidences a more accurate reconstruction of model parameters.

6. Results and discussion

The density pattern appears heterogeneous, with marked lateral variations (Figs. 10a to 10f). High density bodies ($\geq 2.75 \cdot 10^3$ kg·m$^{-3}$) with irregular shape are imaged from 4 to 8 km depth. We relate the high density variations mainly to change in lithology and fracturing. The density increase with depth is mainly due to the lithostatic load. Low densities ($2.40 - 2.65 \cdot 10^3$ kg·m$^{-3}$) are related to shallow molasse and flysch deposits in the southern sector of the investigated area. The density range $2.63 - 2.75 \cdot 10^3$ kg·m$^{-3}$ are attributed to the Cretaceous-Jurassic limestones. The Triassic rocks, mainly made up of dolomitic rocks and limestones, are characterized by a wide

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8) is compared to the observed field. Gravity residuals reproduced by this 3D density structure (Fig. 9) reveal that Bouguer anomalies are well fitted within the limits of 6 mGal. As discussed above, the positive and negative sign of residuals identify those regions where the chosen $\rho - V_p$ linear relationship should be slightly changed to fit the real structures. Only an anomalous high spot of 12 mGal is recognized at 13.43°E 46.18°N. This anomaly in the misfit is connected to the gravity response of the prisms belonging to the first two layers (0 and 2 km) in this position, for which we have a loss in the seismic ray coverage that corresponds to the a priori $C_{mn}$ value of 200 kg m$^{-3}$.

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Table 2 - Statistical parameters related to the estimated P-wave velocity in each layer with seismic tomography (ST) and SII; std: standard deviation over the mean.

<table>
<thead>
<tr>
<th>Layer (km)</th>
<th>mean SII vel (km/s)</th>
<th>std SII vel (km/s)</th>
<th>mean ST vel (km/s)</th>
<th>std ST vel (km/s)</th>
<th>max % vel variation</th>
<th>min % vel variation</th>
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<td>5.51</td>
<td>0.25</td>
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range of density (2.72 – 2.90\times10^3 \text{kg}\cdot\text{m}^{-3}). The highest values (2.80 – 2.90\times10^3 \text{kg}\cdot\text{m}^{-3}) pertain to the dolomitic rocks. The low values are attributable to minor arenaceous layers. The density values in the range 2.65 – 2.78\times10^3 \text{kg}\cdot\text{m}^{-3}, below 6 km depth in the northern sector, can be related to the Paleozoic units (terrigenous sediments, limestones, volcanic and low grade metamorphic rocks).

The bulk modulus ($K$), the Young modulus ($E$) and the shear modulus ($G$) are then calculated using the relationship between elastic moduli and the values of density, $V_p$ and $V_s$. Figs. 11 to 13 show the bulk modulus (Figs. 11a and 11b), the Young modulus (Figs. 12a and 12b), the shear modulus (Figs. 13a and 13b) with the relocated seismicity, at 6 and 8 km depth. These depths are taken as representative of the crustal structure where most of the seismicity is located. The pattern of the elastic moduli appears quite complex and reflects the structural heterogeneity of the crust. As already emphasized, the study area is a poliphase deformational belt, resulting from the superposition of several tectonic phases, that fragmented the crust into different tectonic domains. The bulk, shear and Young modulus images highlight a high rigidity body in comparison with the surroundings, as already revealed by the tomographic images, with an irregular shape. The high rigidity body is characterized by $G \geq 3.2\times10^{10} \text{N}\cdot\text{m}^{-2}$, $K \geq 6.8\times10^{10} \text{N}\cdot\text{m}^{-2}$ and $E \geq 8.4\times10^{10} \text{N}\cdot\text{m}^{-2}$. The bulk, shear and Young moduli are characterized by similar patterns. The seismicity is generally located in zones characterized by marked variations of the moduli pattern. Part of seismicity is located inside the high rigidity bodies, but it appears mostly concentrated along the sharp transitions of the elastic moduli.

Figs. 14a and 14b show the location of the most severe earthquakes ($M_r$ ranging from 4.5 to 6.4) occurred from 1976 to the present day in the study area, on the shear modulus map at 6 and 8 km depth. The locations are from Zonno and Kind (1984), Slejko et al. (1999) and Bressan et al. (2009). They include the strongest shocks of the May 6, 1976 sequence (numbers 1-10), the April 12, 1998 earthquake (K98) and the July 12, 2004 event (K04). All earthquakes are located in a transition zone from high to low rigidity.

The relation between the location of earthquakes and the mechanical properties of the crust is not univocal in many cases, but it should be contextualized in that particular geologic environment. Zhao and Negishi (1998) found that the 1995 Kobe mainshock occurred in a zone characterized by low P and S velocities and high Poisson’s ratio, an effect attributed to fluid-
filled, fractured rocks. Powell et al. (2010) observed in the New Madrid area seismicity associated to low $V_p/V_s$, resulted by low $V_p$ and high $V_s$ anomalies, an effect due to quartz-rich rocks. On the other hand, Hauksson and Haase (1997) observed that earthquakes occur in or close to high velocity zones in the Los Angeles basin area. Vlahovic and Powell (2001) found that

Fig. 10 - Density model at depth slices 0 (a), 2 (b), 4 (c), 6 (d), 8 (e)10 (f) km.
earthquakes were located along the edges of a more competent crust (high velocity zones), surrounded by zones characterized by low velocity, in the New Madrid area. Koulakov et al. (2010) showed that strong earthquakes can occur in the transition areas between higher (rigid blocks) and lower velocities (highly fractured rocks), while weak and moderate seismicity is concentrated in low velocities zones. The results of the present study and past observations (de Franco et al., 2004) about the distribution of seismicity and the pattern of elastic moduli appear quite concordant with the findings of Koulakov et al. (2010). Our favoured interpretation is based on the arguments advanced by Chatterjee and Mukhopadhyay (2002) and by Maxwell and Young (1992). Chatterjee and Mukhopadhyay (2002) investigated the local stress modifications caused by variations in the rock mechanical properties. They found that stress concentration is high and

Fig. 11 - Bulk modulus calculated at 6 km depth (a) and 8 km depth (b). Diamonds: relocated earthquakes (see text at chapter “Tomographic inversion”).

Fig. 12 - Young modulus calculated at 6 km depth (a) and 8 km depth (b). Diamonds: relocated earthquakes (see text at chapter “Tomographic inversion”).
maximum rotation of the principal stresses occurs in zones where greater is the contrast in elastic properties of rocks. Following Maxwell and Young (1992), the zones of marked variation of the elastic properties of rocks modify the stress field by destressing and relaxation processes of the clamping forces, through variation of the effective confining pressure and rotation of the principal axes of stress. The rotation of the stress tensor can reduce the fracture normal stress at the fracture surface, favouring the occurrence of seismicity.

Therefore, we interpret the high rigidity bodies as more competent parts of the crust that have greater capability of storing strain energy. The high rigidity bodies are bordered by sharp variations of the elastic moduli that represent rock volumes with progressive increasing damage, differently able to accumulate strain energy. The stresses stored are likely released in the transition zones, close to high rigidity bodies, generating the seismicity. Therefore, the elastic moduli variability, that is the contrast in rock mechanical properties, influences the occurrence of weak and large earthquakes.

We assign the main lithologic units to the following values of elastic moduli. The Paleozoic rocks are characterized by $E$ ranging between $7.0-9.0 \cdot 10^{10}$ N·m$^{-2}$, $G$ between $2.5$ and $3.5 \cdot 10^{10}$ N·m$^{-2}$, $K$ in the range $5.0 - 7.0 \cdot 10^{10}$ N·m$^{-2}$. A wide range of elastic moduli values are attributed to the Mesozoic units: $E$ in the range $5.5 - 10.5 \cdot 10^{10}$ N·m$^{-2}$, $G$ between $2.0$ and $4.0 \cdot 10^{10}$ N·m$^{-2}$ and $K$ in the range $4.5 - 9.0 \cdot 10^{10}$ N·m$^{-2}$. Flysch and molasse deposits are generally characterized by low values: $E$ between $4.0$ and $6.0 \cdot 10^{10}$ N·m$^{-2}$, $G$ between $1.5$ and $2.5 \cdot 10^{10}$ N·m$^{-2}$ and $K$ in the range $3.3 - 5.6 \cdot 10^{10}$ N·m$^{-2}$.

The values of the elastic moduli obtained from the SII are compared with those obtained on synthetic stratigraphic profiles from laboratory (lab) measurements (Faccenda et al., 2007). The synthetic profiles modelled the sedimentary cover of the Friuli area by including the most representative rock types of the litho-stratigraphic sequence that mainly compose the upper crust of the Friuli area. The lab $E$ ranges from $7.0$ to $11.0 \cdot 10^{10}$ N·m$^{-2}$. The lab $K$ is characterized by values between $6.1$ and $8.8 \cdot 10^{10}$ N·m$^{-2}$. The lab $G$ varies from $2.7$ to $4.2 \cdot 10^{10}$ N·m$^{-2}$. It is interesting to note that the lab values of $E$, $K$, $G$ fall inside about the middle-high range of the

Fig. 13 - Shear modulus calculated at 6 km depth (a) and 8 km depth (b). Diamonds: relocated earthquakes (see text at chapter “Tomographic inversion”).
values found from SII inversion. The highest values of the $E$, $K$, $G$ lab measurements ($E$: $9.4-11.0\cdot10^{10}$ N·m⁻², $K$: $7.8-8.8\cdot10^{10}$ N·m⁻², $G$: $3.6-4.2\cdot10^{10}$ N·m⁻²) are pertaining to the dolomitic rocks that, in our interpretation, mainly form the high rigidity bodies revealed in the area from 4 to 8 km depth.

7. Conclusions

The 3D $V_p$ and $V_p/V_s$ models of the upper crustal layers are characterized by marked lateral and depth variations reflecting the geologic-structural heterogeneity of the study area, caused by the superposition of several tectonic phases, with different orientations of the principal axes of stress. We relate the $V_p$ and $V_p/V_s$ anomalies to the lithological heterogeneities and to the different degree of fracturing and hence to the brittleness. The 3D $V_p$ model is used together with gravity data information as input to the tomographic inversion method SII in order to recover the density model. This density model explains the observed Bouguer gravity map with misfits between $-6$ and 6 mGal and provides a check on the reliability of the 3D $V_p$ model. Hence, the SII algorithm, through the joint use of seismic and gravity information, decreases the degree of ambiguity implicit in the reconstruction of the density model, difficult to determine on the basis of seismic or gravity data alone. Furthermore, it handles the difficulties due to a linear relationship between seismic velocities and densities, through the use of an a priori density covariance matrix.

The density pattern is heterogeneous, with marked variations that we associate mainly to changes in lithology and fracturing. A wide range of density ($2.63 - 2.90\cdot10^{3}$ kg·m⁻³) is attributable to Mesozoic limestones with the highest values ($2.80 - 2.90\cdot10^{3}$ kg·m⁻³) assigned to the Triassic dolomitic rocks. The Paleozoic rocks are characterized by density range $2.65 - 2.78\cdot10^{3}$ kg·m⁻³, while low densities ($2.40 - 2.65\cdot10^{3}$ kg·m⁻³) are related to shallow molasse and
terrigenous sediments. The density model sets reliable constraints on the material properties within the study area.

The pattern of the elastic moduli (bulk modulus, Young modulus and shear modulus) reflects the structural heterogeneity of the investigated crust.

The most important feature revealed by the bulk, shear and Young moduli pattern is a high rigidity body \( G \geq 3.2 \cdot 10^{10} \text{ N}\cdot\text{m}^{-2}, K \geq 6.8 \cdot 10^{10} \text{ N}\cdot\text{m}^{-2} \) and \( E \geq 8.4 \cdot 10^{10} \text{ N}\cdot\text{m}^{-2} \) in comparison with the surroundings, recognizable from 4 to 8 km depth, with an irregular shape. The earthquake occurrence appears related to the material heterogeneities. The seismicity is mainly located along the sharp variations of the moduli pattern, in or adjacent to the high rigidity zones. The most severe earthquakes \( (M_r \text{ ranging from 4.5 to 6.4}) \) occurred in the study area from 1976 to the present day are located in a transition zone from high to low rigidity patterns. The marked differences in mechanical properties of the crust significantly affect the stress distribution and different rigidity zones suggest different capability to store strain energy. High transition between higher and lower rigidity zones favours stress concentration, causing the occurrence of seismicity. The role played by crustal heterogeneity, that can significantly influence the stress trajectories and the loading stresses, suggest that care must be taken in modeling the stress propagation and stress transfer with uniform isotropic elastic medium for investigating the earthquake occurrence.

Laboratory measurements, not available in the past studies of the Friuli area, are meaningful for the calibration of geophysical properties of rocks obtained with the SII method. The values of \( E, K, G \) found from laboratory measurements correspond to the middle-high range of the values obtained with the sequential integrated inversion. The differences may indicate in-situ fracturing since laboratory data are often biased high because of the highest quality of samples for experiments.

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