Evidence for variations of mechanical properties in the Friuli seismic area

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Abstract


Tidal strains of the earth were extracted from three horizontal strainmeter records (1979–1986) in the Friuli seismic area by the “filter method”. The areal strain factor, ignoring the tidal loading effects, shows that its time variation is significant: their amplitudes in 1986 are about 50% of those in 1979. The modifications of mechanical properties, estimated in terms of the local shear and bulk modulus variations, were obtained considering the envelope of the tidal strain signals. The seismic wave velocity determined by the simultaneous inversion of the arrival time data of the local seismometric network displayed comparable changes.

A major change both in the seismic velocities and the elastic parameters started in March 1982, which was about 11 months before an earthquake of magnitude 4.1, the largest event from 1979 to 1986, which occurred within the seismic network on Feb. 10, 1983.

1. Introduction

The literature concerning tidal effects dates back to Lord Kelvin times and is very extensive (Melchior, 1978). Traditionally, the efforts in this field were mainly devoted at gaining an insight into the physics of the Earth’s interior.

However, theoretical arguments made clear, already at an early stage, that numerous corrections have to be applied to Earth’s tide measurements to make them suitable for interpretation with respect to such problems. In fact, after the elimination from the records of the long-period and very-short-period terms and of the aperiodicities, the effects of the atmospheric-pressure and water-table variations have to be also subtracted from them. Moreover the gravitational and loading effects of the ocean tides have to be estimated for each measurement site together with the indirect effects due to the gravity potential modification caused by the induced crustal tilting.

Once properly corrected, Earth tide data are compatible with rigidity moduli of the uppermost crustal layers. However, the elastic parameters inferred from the analysis of the records differ greatly from place to place (Nishimura, 1950; Tomaszek, 1957; Marussi, 1960), supporting the conclusions drawn earlier from the empirical estimates of the loading effects alone (Takahasi, 1929; Nishimura, 1950) that heterogeneities are responsible for the marked deviations from the theoretical values (Jobert, 1956).

Subsequently, in the last fifteen years it has become apparent that additional factors influence the measurements, producing local data inhomogeneities which make the estimates of the whole-Earth tide magnitude difficult (Lennon and Baker, 1973; Baker and Lennon, 1976).
Further observations at seven stations in the United States (Beaumont and Berger, 1975) showed that, when corrected for ocean loading, the amplitude differences between the observed and theoretical strains on the average reach 10% in the $M_2$ constituent and 12% in the $O_1$.

King and Bilham (1973) pointed out that tidal strains deform the cavities in which the observations are made and Lecolazet and Wittlinger (1974) computed such effects. Also Harrison (1976) calculated the deformations produced by a number of cavities of peculiar shapes and stressed the perturbing role of the topography and of the elastic heterogeneities of the medium.

Although properly corrected, tidal data still exhibit notable anomalies, often large and very-frequent. Evans et al. (1979) reported 35 measurements of Earth tidal strains from 16 near-surface sites in Great Britain. Corrected for both cavity and topographic effects, their data still exhibit variations up to 50% in tidal admittance; again such large anomalies were attributed by them to regional variations in the elastic moduli.

In general the available data indicate that the computed Love numbers cannot be considered representative of the whole-Earth properties and thus that other methods are more suitable for the investigation of the global physical structure of the Earth’s interior.

On the contrary, as suggested by Harrison (1985), Earth tide measurements may be used to study local geological problems and changes in the tidal parameters with time. Theoretical studies made by Beaumont and Berger (1974) indicate that the amplitude of tidal strain and tilt could change by as much as 60% together with a seismic P-velocity decrease of 15% in a dilatant zone embedded within the crust. Tanaka (1976) estimated the possible changes of the loading tides due to dilatancy assumed in coastal regions: his results show that amplitude changes in the $M_2$ wave could reach the order of 10% of its normal value. A similar estimate given by Mao (1984) indicates that changes of loading tidal tilt depend strongly on the station locations, the possible changes in $M_2$ wave before an earthquake of the magnitude 7 in coastal regions being between 20% and 30%. However, only few observations of Earth tide amplitudes have been reported to date in relation to earthquakes (Latynina and Rizaeva, 1976; Mikumo et al., 1978).

When Crosson (1976), Aki and Lee (1976) developed the technique of simultaneous inversion of hypocentral location and seismic velocity structure by the seismic wave arrival times, Aki and Lee (1976) pointed out that, with improved resolution and accuracy, the technique would have enabled the detection of precursory velocity changes, if any, for smaller earthquakes. Since smaller earthquakes are more frequent, it is possible to apply the simultaneous inversion method to shorter-term earthquake prediction. Fitch and Rynn (1976) also described an inversion method in which localized volumes of anomalously low seismic velocity could be identified using arrival time data from local earthquakes. They suggested that the inversion method can resolve near-source effects, whereas the past methods require the velocity change to be regional in scale in order to be detected.

In the present work, the time variations of the tidal strain in the Friuli seismic area, Northeast Italy, were investigated and ascribed to modifications of the effective elastic moduli of the uppermost crust which occurred in connection with the 1976 earthquake. Moreover, the seismic wave velocities were inverted by the technique of the simultaneous determination of hypocenter locations and the velocity structure to ascertain whether, in the different time intervals, the velocities were compatible with the moduli changes assumed above; the data relative to the Friuli seismic network operated by the Osservatorio Geofisico Sperimentale, Trieste (OGS) were used.

2. Observations

Since November 1978, three Cambridge invar-wire strainmeters have been recording the extensional rates in a natural cave near the village of Villanova together with a couple of Marussi horizontal-pendulum tiltmeters.

The part of the cave where the instruments are installed is a pseudo-elliptical cavity (Ebbin, 1986). The largest dimension of the room is about 20 m in NNE direction, roughly 4/3 of the inter-
mediate one; the maximum height is about 7 m. The annual temperature variation in the room is not more than 1° C.

The cave site is located at lat. 46.253° N and long. 13.279° E at an elevation of about 570 m a.s.l. It lies at about 60 m depth below the surface in the eastern part of the seismically active zone of central Friuli where the E-W-trending structure of the Alps merge into the northwest termination of the NW–SE-trending Dinarides yielding a peculiar geodynamic situation (Carulli, 1986).

The three strainmeters, for convenience called STRN2, STRN3 and STRN4, are horizontal, oriented N128° E, N27° E and N68° E. Their lengths are 13.06, 12.63 and 14.33 m, respectively.

The electronically amplified strainmeter data are pen recorded on paper and then digitized with a sampling interval of one hour, punched on paper tape and stored on magnetic discs. The instruments are calibrated once a month by changing the input signal artificially. Their sensitivity is about 4 × 10^{-10}/0.2 mm.

Unfortunately, owing to technical problems, STRN2 has not produced useful records from January 1984 to March 1986.

The seismometric network has been established in the seismically active zone of central Friuli by OGS in May 1977. One-second period seismographs of the single vertical component are now in operation in 15 stations in a region of about 100 by 100 km² (Fig. 1). Nine of them are radio linked in digital format with the data acquisition center in Udine, where the data are recorded on magnetic tape by means of a Racal Geostore recorder.

3. Tide analysis

Three longitudinal strain measurements, in non-parallel directions on a plane, define completely the strain tensor on that plane if homogeneity is assumed.

Two different periods out of 8 years of tidal data recorded by the strainmeters in the Villanova observatory have been used to calculate the tidal factors of the areal strain. Unfortunately the tilt data could not be included in the analysis since the tiltmeter sensitivity barely reaches the tidal amplitude magnitude.

The theoretical strain tides can be evaluated using the Love numbers $h_n$, $l_n$, and the tide-generating potential $W_n$. The components for latitudinal, meridional and shear strains in spherical polar coordinates $E_{\theta\theta}$, $E_{\lambda\lambda}$ and $E_{\theta\lambda}$ are given by

\[
E_{\theta\theta} = \sum \frac{l_n}{ag} \frac{\partial^2 W_n}{\partial \theta^2} + \sum \frac{h_n}{ag} W_n,
\]

\[
E_{\lambda\lambda} = \sum \frac{l_n}{ag \sin^2 \theta} \frac{\partial^2 W_n}{\partial \lambda^2} + \sum \frac{h_n}{ag} \cot \theta \frac{\partial W_n}{\partial \theta}
\]

\[
E_{\theta\lambda} = \sum \frac{2l_n}{ag \sin \theta} \frac{\partial^2 W_n}{\partial \theta \partial \lambda} - \sum \frac{2l_n}{ag \sin \theta} \cot \theta \frac{\partial W_n}{\partial \lambda}
\]

where $g$ is gravity at the Earth's surface, $a$ is the Earth's radius, $\theta$ and $\lambda$ are co-latitude and longitude of the point of interest, respectively.

The horizontal strain in any direction, $e_i$, is:

\[
e_i = E_{\theta\theta} \cos^2 \alpha_i + E_{\lambda\lambda} \sin^2 \alpha_i - E_{\theta\lambda} \sin \alpha_i \cos \alpha_i
\]

where $\alpha_i$ is the azimuth of the strain observation.

The observed records were band-pass filtered to extract the frequencies between 0.5 and 6 cycles
per day and then compared with the theoretical strains (Fig. 2).

There are three points worth noting in the comparison between the observed and theoretical signals (Fig. 2): (1) There are evident amplitude discrepancies between the observed and theoretical tidal strains. (2) The amplitude discrepancies in the three different directions are not the same. (3) The three components of the observed tidal strains have undergone a remarkable reduction with time.

Points 1 and 2 are general phenomena certainly caused by the various effects mentioned above as well as to the local heterogeneities. Here, our main attention is focused on point 3.

a. Variation in Love numbers

Kuo (1969) and Smith and Jungels (1970) chose to work with the areal dilatation as calculated from a three-axis strain gauge since this was shown to be unaffected, to the first order, by ocean or atmospheric loading. The M₂ tidal strain effect produced by the Adriatic Sea has been calculated.

The results show that the amplitudes of $E_{\theta\theta}$ and $E_{\lambda\lambda}$ are $1.24 \times 10^{-9}$ and $1.34 \times 10^{-9}$ respectively and the phase difference is about 180°. Therefore, the load areal strain for this closed basin is indeed approximately zero.

The areal strain of the diurnal waves and semi-diurnal waves are given by:

$$
\sigma_i = E_{\theta\theta} + E_{\lambda\lambda} = \frac{2(h - 3l)W_{2i}}{ag} \quad (i = 1, 2)
$$

where $W_{21}$ and $W_{22}$ are the tide generating potentials of diurnal and semi-diurnal waves respectively, and $h - 3l$ is called the areal strain factor. $\sigma_i$ can also be expanded into the summation of the tidal constituents, that is:

$$
\sigma_i = \sum_n A_{ni}\cos(\omega_{ni}t - \beta_{ni})
$$

where $n$ denotes different diurnal and semi-diurnal constituents, $t$ is time, $A_{ni}$, $\omega_{ni}$, and $\beta_{ni}$ are their amplitude, frequency and phase, respectively.

From eqns. (3) and (4) the amplitude and phase spectra can be obtained at the various frequencies, if the same numerical Fourier analysis is used for the observed and theoretical tidal areal strains for
TABLE 1

<table>
<thead>
<tr>
<th>Wave</th>
<th>Areal strain T Am.</th>
<th>Areal strain O Am.</th>
<th>h-3l</th>
<th>Phase lag 0</th>
</tr>
</thead>
<tbody>
<tr>
<td>1979 June-November</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>M2</td>
<td>1.178</td>
<td>1.026</td>
<td>0.314</td>
<td>14</td>
</tr>
<tr>
<td>O1</td>
<td>1.085</td>
<td>0.717</td>
<td>0.238</td>
<td>4</td>
</tr>
<tr>
<td>1986 May-October</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>M2</td>
<td>1.163</td>
<td>0.473</td>
<td>0.147</td>
<td>6</td>
</tr>
<tr>
<td>O1</td>
<td>1.064</td>
<td>0.476</td>
<td>0.165</td>
<td>11</td>
</tr>
</tbody>
</table>

* T and O denote theoretical and observed tides, respectively. The unit of areal strain amplitude is $10^{-8}$.

The time interval of interest. The areal strain factors may be obtained by comparing them directly. Two different numerical processes can be adopted for the areal strain factor computation. In the first one the areal strain factor is obtained frequency by frequency band from the spectral values of the three strain components. In the second one the areal strain is calculated from the three strain components in the time domain and the spectral analysis is carried out only for the areal strain time series. The numerical errors due to spectral contamination are smaller in the latter case since only one spectral analysis is performed.

Thus the latter approach is applied to two different 6 months data periods, one from 1979 June 19, at 00h, and another from 1986 May 1, at 00h. Two reliable constituents $M_2$ and $O_1$ are separated from the areal strain data, the areal strain factors are given in Table 1. The advantage of this process is that some artificial errors in the data analysis are avoided (Smith and Jungels, 1970).

The results from the two time intervals show significant differences, the areal strain factors of 1986 being less than 50% of those of 1979. Such differences cannot be attributed to cavity, topographic and loading effects, which are obviously the same in the two time intervals and therefore they might be ascribed to (1) variations of the local elastic parameters and/or to (2) calibration errors. The phase of the areal strain is affected directly by errors in the relative calibrations of the three strain gauges. In our results, the phase lags appear to be small, suggesting that the instrumental calibrations are reliable. Therefore the variations are not likely to be related to calibration errors but rather to local variations of the mechanical properties of the rocks. The latter hypothesis is also supported by the fact that the three components of strain exhibit all the same reducing trend with time (Fig. 2).

b. Variation in elastic parameters

In order to obtain satisfactory results, the spectral method requires long and continuously observed data sequences; record interruptions cause strong limitations to its applicability. In contrast, the commonly used method of harmonic analysis is suitable to determine Love numbers from short sequences of data. However, mechanical parameters, which have a direct physical meaning, give a more immediate description to the properties of the rocks than Love numbers. Here, from the tidal strain data, we attempt to give a new method to investigate the time variations of the elastic parameters.

In order to estimate the tidal strain amplitude of the perigean and apogean oscillations, the envelope of the tidal signals with nearly half-months period was found (Figs. 2 and 3). Such an envelope can be calculated through the following steps.

If $f(t)$ is a time series, the instantaneous amplitude $A(t)$ and phase $\phi(t)$ are defined as (Dzewonksi and Hales, 1972):

$$A(t) \cdot e^{i\phi(t)} = f(t) + i f_1(t)$$

where $f_1(t)$ is the quadrature signal of $f(t)$. The quadrature signal is defined as:

$$f_1(t) = FT^{-1} \{ FT \{ f(t) \} \cdot s(\omega) \}$$

where $s(\omega)$ is a sign function. Therefore the envelope of $f(t)$ is given by:

$$A(t) = \left\{ f^2(t) + f_1^2(t) \right\}^{1/2}$$

and the instantaneous phase is:

$$\phi(t) = \tan^{-1} \left[ f_1(t)/f(t) \right]$$

In our case, $f(t)$ is the band-pass filtered tidal curve. We suggest that the band width is from 11 to 13 hrs (Fig. 3). The peak values of the three
observed and theoretical strain components and their fitting curve obtained by least-squares method are presented in Fig. 4.

Assume now the Earth medium to be isotropic and linearly elastic, and the tidal stresses to be equivalent for the real Earth and the Earth model. In the spherical reference system the relations between the elastic parameters and the strains are:

\[
\frac{\mu^o}{\mu^T} = \frac{e_{\theta\theta}^T - e_{\lambda\lambda}^T}{e_{\theta\theta}^o - e_{\lambda\lambda}^o} \tag{9}
\]

\[
K^o
K^T = \frac{2\Delta^T}{3\left[(e_{\theta\theta}^o + e_{\lambda\lambda}^o) - \Gamma(e_{\theta\theta}^T + e_{\lambda\lambda}^T)\right] + 2\Gamma\Delta^T} \tag{10}
\]

where the superscript T stands for the theoretical tide, o for the observed one; \(\mu\) and \(K\) are shear and bulk moduli, respectively; \(\Delta\) is the cubic dilatation, on the surface:

\[
\Delta = -\frac{\lambda}{\lambda + 2\mu}(e_{\theta\theta} + e_{\lambda\lambda})
\]

\[
\Gamma = \frac{\mu^T}{\mu^o} \tag{11}
\]

Here we use a local coordinate reference system bisecting the principal axes to obtain big, stable normal strain components.

The strain tensors are computed from the peak values of the envelopes of the three observed and theoretical longitudinal strains, then the reference systems are rotated to bisect the principal directions of the observed and theoretical strains and a new, rotated strain tensor is obtained by:

\[
E^N = R^T(a)ER(a) \tag{12}
\]

where \(R\) is the rotation tensor, the rotation angle
\( \alpha \) is equal to the half of the summation of the observed and theoretical strain principal angles.

With the strain components in the new systems, the time variations of the shear modulus \( \mu \) and of the bulk modulus \( k \) can be calculated by eqns. (9) and (10). The results are given in Figs. 5a and b, respectively. The shear modulus \( \mu \) increases to become, in 1986, 2.5 times larger than in 1979.

The variation is consistent with that of the areal strain factors by the spectral method for the two different periods (Table 1). The bulk modulus reached a peak value in 1983. It should be noted that the calculation of the shear modulus is affected to a great extent by the loading tides since there is the term \( e_{\theta \theta} - e_{\lambda \lambda} \) in eqn. (9).

The advantages of this way are that (1) the peak values of envelope are more reliable, since the derivative of the time function is small near the peak value and its relative error is small, too. (2) Only the records near the peak values are important in the analysis, so the data interruptions beyond \( |t_p + L| \) do not affect the analysis results, where \( t_p \) is the time corresponding to the peak value and \( L \) is time lag in the filter process. (3) The mechanical parameters of the rocks have a more direct physical meaning than tidal Love numbers.

4. Arrival time inversion

In the section above, the time variations of the rock mechanical properties in a Friuli site were found by analyzing tidal strain data. As is well known, seismic wave velocities also depend on the mechanical parameters of the media. At any change in the media, the seismic wave velocities
will vary accordingly. Theoretical relations between the earth tides and seismic wave velocity variation have been studied by Beaumont and Berger (1974). However, so far only few simultaneous analyses of the observed results of tidal strain and seismic velocity changes have been reported (Mikumo et al., 1978). This may be because it is not easy to obtain long time tidal strain observations because of the limitation of the instrument, station and geographic conditions. Furthermore, it is very rare to have three or more component strainmeters and a microearthquake network in the same seismic area.

Since the development of the technique of simultaneous inversion for velocity and hypocenter of the P and S wave arrival time, it has been possible to determine $V_P$ and $V_S$ separately. In this section, we attempt to investigate time variation of seismic wave velocity by the simultaneous inversion theory, and compare these results to the results obtained from the tidal strain analysis.

a. Inversion method

The arrival time data recorded within the seismic network contain information about the earthquake space–time location and the seismic velocity structure of the Earth. Consider $N$ earthquakes occurring within a network of $M$ stations, the linearized equations for simultaneous inversion relating the arrival time residuals $r$ to the model parameters changes can be written as:

$$ r_{ij} = \sum_{k=1}^{4} \left( \frac{\partial T_{ij}}{\partial h_{jk}} \right) \delta h_{jk} + \sum_{k=1}^{P} \left( \frac{\partial T_{ij}}{\partial v_k} \right) \delta v_k $$

$$ (i = 1, \ldots, N; \ j = 1, \ldots, M) $$

where $T$ is the travel time, $P$ the number of the velocity model parameters, $h_{jk}$ the hypocentral parameters, $v_k$ the velocity parameters. Equation (13) may be written in matrix form as:

$$ r = \mathbf{G} \, \mathbf{d} \mathbf{x} $$

where $r$ is an $MN$-dimensional vector of travel-time residuals, $\mathbf{d} \mathbf{x}$ is a $(4N + P)$-dimensional vector of hypocenter parameter and velocity parameter adjustments, $\mathbf{G}$ is a $MN \times (4N + P)$ matrix which consists of the partial derivatives of the travel time $T$ with respect to hypocenter and velocity parameters.

By means of the damped least-squares method (Crosson, 1976), the generalized inverse of the matrix $\mathbf{G}$ is:

$$ \mathbf{G}^\dagger = \mathbf{V} \left[ (\mathbf{S}^2 + \theta^2 \mathbf{I})^{-1} \mathbf{S} \right] \mathbf{U}^T $$

where $\mathbf{U}$ is a $MN \times MN$ orthogonal matrix which consists of $MN$ eigenvectors of $\mathbf{G} \mathbf{G}^T$, $\mathbf{V}$ is a $(4N + P) \times (4N + P)$ orthogonal matrix which consists of $4N + P$ eigenvectors of $\mathbf{G}^T \mathbf{G}$, $\mathbf{S}$ is a $MN \times (4N + P)$ diagonal matrix which consists of non-negative square roots of the eigenvalues of $\mathbf{G}^T \mathbf{G}$, $\mathbf{I}$ is the identity matrix and $\theta^2$ is the damping constant.

Following the generalized inverse notation, the estimated results (denoted by a "hat"), the resolution matrix $\mathbf{R}$ and the covariance matrix $\mathbf{C}$ of the solutions can be calculated by:

$$ \mathbf{d} \hat{\mathbf{x}} = \mathbf{G}^\dagger \mathbf{r} $n$$

$$ \mathbf{R} = \mathbf{G}^\dagger \mathbf{G} $$

$$ \mathbf{C} = \sigma^2 \mathbf{G}^\dagger (\mathbf{G}^\dagger)^T $$

(Wiggins, 1972), where $\sigma^2$ is the observation variance.

Since the coefficient matrix $\mathbf{G}$ in eqn. (14) is a function of the model, the problem must be solved iteratively. An initial guess of the velocity model and hypocenters has to be iterated to refine the estimates using eqn. (16). The $\theta^2$ is an adjustable parameter for controlling the trade-off between resolution and covariance (Aki and Richards, 1980). As $\theta^2 \to 0$, $\mathbf{R} \to \mathbf{I}$, the parameters to be determined are perfectly resolved. As $\theta^2$ increases, the covariance values decrease, while the resolution matrix $\mathbf{R}$ is departing more strongly from the identity matrix.

b. Data selection

We use the arrival time data obtained from the Friuli seismometric network by the OGS. The original data have been analyzed using a standard local earthquake location program HYPO71 (Lee and Lahr, 1975) by the OGS staff. The edited data have been written on magnetic tape. The crust
TABLE 2
The Friuli velocity structure used by the OGS in the location procedure

<table>
<thead>
<tr>
<th>Velocity (km/s)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.85</td>
<td>22</td>
</tr>
<tr>
<td>6.80</td>
<td>17.5</td>
</tr>
<tr>
<td>8.00</td>
<td>∞</td>
</tr>
</tbody>
</table>

velocity model used in such analysis is shown in Table 2. The ratio $V_p/V_s$ is fixed at 1.78.

P waves of 371 earthquakes and S waves of 345 earthquakes were chosen from among the events occurring within the network in the time interval from January, 1978 to August, 1986 (Fig. 1). The selected events strictly satisfy the following criteria: (1) events recorded on four or more stations; (2) A, B, or C quality solution; (3) transmitting system radio-linked; (4) events located in a region of about 20 km by 20 km within the network for consideration of similar ray paths along which the velocities are analyzed, and (5) events having small arrival time residuals ($r < 0.5$ s) in order to control large observational errors. The choice of the particular value in this case has been made on the basis of the quality of the data.

c. Seismic wave velocity variations

Since our aim is to study seismic velocity variation with time, i.e. relative velocity variations, we select a further 268 events with source depths limited between 5 km and 10 km and divide them into thirteen different time intervals, respectively. Each interval includes 21 events (Table 3). Thus the events we use are distributed stochastically within a block which is $20 \times 20 \times 5$ km$^3$. In fact, it is reasonable to use any acceptable geometrical

TABLE 3
The relative variation of P and S wave velocity

<table>
<thead>
<tr>
<th>Time interval</th>
<th>Ray number</th>
<th>Wave</th>
<th>Standard deviation</th>
<th>Velocity (km/s)</th>
<th>Standard error</th>
</tr>
</thead>
<tbody>
<tr>
<td>78.01–78.10</td>
<td>93</td>
<td>P</td>
<td>0.096</td>
<td>5.77</td>
<td>0.026</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.23</td>
<td>3.17</td>
<td>0.015</td>
</tr>
<tr>
<td>78.11–79.07</td>
<td>96</td>
<td>P</td>
<td>0.085</td>
<td>5.77</td>
<td>0.038</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.21</td>
<td>3.19</td>
<td>0.014</td>
</tr>
<tr>
<td>79.08–80.08</td>
<td>96</td>
<td>P</td>
<td>0.089</td>
<td>5.74</td>
<td>0.032</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.18</td>
<td>3.16</td>
<td>0.010</td>
</tr>
<tr>
<td>80.09–81.06</td>
<td>103</td>
<td>P</td>
<td>0.093</td>
<td>5.74</td>
<td>0.032</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.22</td>
<td>3.18</td>
<td>0.012</td>
</tr>
<tr>
<td>81.07–82.01</td>
<td>98</td>
<td>P</td>
<td>0.092</td>
<td>5.73</td>
<td>0.034</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.24</td>
<td>3.15</td>
<td>0.013</td>
</tr>
<tr>
<td>82.02–82.12</td>
<td>107</td>
<td>P</td>
<td>0.098</td>
<td>5.84</td>
<td>0.020</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.24</td>
<td>3.19</td>
<td>0.011</td>
</tr>
<tr>
<td>83.01–83.08</td>
<td>117</td>
<td>P</td>
<td>0.10</td>
<td>5.89</td>
<td>0.020</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.31</td>
<td>3.22</td>
<td>0.010</td>
</tr>
<tr>
<td>83.09–84.01</td>
<td>120</td>
<td>P</td>
<td>0.11</td>
<td>5.74</td>
<td>0.020</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.27</td>
<td>3.13</td>
<td>0.009</td>
</tr>
<tr>
<td>84.02–84.06</td>
<td>117</td>
<td>P</td>
<td>0.10</td>
<td>5.76</td>
<td>0.021</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S</td>
<td>0.32</td>
<td>3.14</td>
<td>0.010</td>
</tr>
<tr>
<td>84.07–84.10</td>
<td>122</td>
<td>P</td>
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model to investigate the relative velocity variations. The station positions are fixed and if the media through which the rays pass are perturbed, the arrival times will also show a perturbation, no matter what the geometrical configuration of the model is. For simplicity, in this study we use a homogeneous half-space geometrical model.

The inversion method described above requires a reasonably good initial trial model. Here we use the velocity model (Table 2) and hypocenter solutions by OGS as the initial trial models. In order to reduce the errors to reasonable values, we have chosen as threshold 2% of the maximum singular value in each inversion as the adjustment parameter $\theta^2$, considering small singular values and covariance. The iterative process is continued until the model fits the data to within the observational error. This is measured by the reduced $\chi^2$ misfit (Leonard and Johnson, 1987):

$$\frac{\chi^2}{NM} = \sum_{i=1}^{NM} \left( \frac{r_i}{\sigma_i} \right)^2$$

where $NM$ is the number of travel times used in the inversion, $r_i$ is the $i$th travel time residual, and $\sigma_i$ is the standard deviation of $i$th travel time. An acceptable model is obtained when $\chi^2/NM$ is approximately equal to one. In calculating, $\sigma$ is assigned as 0.1 s for the P arrival time. Equation (17) has an explicit physical sense, i.e., when $\chi^2/NM$ is fixed, the travel-time residuals are determined by the observational errors. For a homogeneous half space model, the velocity is perfectly resolved. The standard errors of the model parameters are given by the diagonal elements of the covariance matrix $C$. The estimated velocities, standard errors and calculating standard deviations of travel time are given in Table 3.

From Table 3 it is found that the observational errors of P waves are smaller than the predicted value of 0.1 s, whereas those of S waves are larger than the 0.1 s. This is because of the difficulty of determining accurately the S wave phase. In order to find more correct $V_s$ velocities, we have used the following method. Owing to the fact that the first P wave phase can be easily and accurately determined, the hypocenters and origin times found using the first arrival times $t_p$ were first taken as real. Then the unique unknown, velocity $V_s$, can be obtained by the given hypocenters, origin times and the arrival times $t_s$. This method not only saves much computational time but also gets reasonable S wave velocities.

Error analyses are an important standard for appraising the inversion results, i.e., in this case, whether the velocity variations are representative of mechanical parameter variations. The standard errors of the velocity given in Table 3 which are about one order of magnitude smaller than the velocity variations, suggest that the velocity variations calculated here are meaningful (Fig. 5c). The standard deviations of travel times also show that the observed random noises can be eliminated by the inversion technique.

5. Discussion

In the present work, three different methods and two kinds of data were used to investigate the mechanical parameter time variations in the Friuli area in the years following the destructive 1976 earthquake. In the case of the Earth tides, since the tidal stresses can be estimated, the variations of the observed strains reflect those of the effective elastic parameters. Since the latter ones induce changes in the seismic velocities, the arrival time data from microearthquake networks have been studied.

It is usually difficult to detect the near-field seismic velocity changes by the microearthquake data because small errors of the hypocentral location and the arrival measurement for single event will bring about a big velocity uncertainty. However, the velocity changes in a seismic area may provide important earthquake precursor information. The results above indicate that the technique of the simultaneous inversion for hypocenter location and seismic velocity can be used to determine the near-field seismic velocity changes.

The bulk modulus changes (Fig. 5) show a good agreement with the seismic velocity changes, whereas the shear modulus changes do not. This may be related to the following possible reasons: (1) the calculation of $\mu$ is affected by the loading tides (see section 3b, and the loading tides change with time; (2) the estimated parameter $\mu$ is unstable owing to the closeness of the values of two
normal strain components of the observed tidal strains; (3) the velocity changes are average values within the microearthquake network, whereas tidal strains mainly reflect the medium properties around the Villanova station. However, since the tidal areal strain factor (section 3a) and the seasonal shear strain (Mao et al., 1988) in the Villanova station exhibit changes similar to those of the shear modulus $\mu$, it seems reasonable to conclude that the disagreement between $\mu$ and the velocity changes is caused by the local heterogeneities of the elastic medium. It is anyway suggested that the first two effects must be considered before using eqn. (9).

Finally it seems quite interesting that both the velocities and the bulk modulus $K$ started to change about 11 months before an earthquake of magnitude 4.1, the largest event from 1978 to 1986, which occurred under the microearthquake network on Feb. 10, 1983. When the earthquake occurred the change of the velocities and of the bulk modulus $K$ had reached about 3% and 50%, respectively. After the earthquake, the velocities and $K$ started to turn to the situation before the event and the recovery time was nearly equal to that of the deviation. Similar results were also reported by Mikumo et al. (1978). It also appears noteworthy that the earthquake occurrence is marked by a bulk modulus increase in clear contradiction with the dilatation hypotheses. Many earthquakes of $M < 4$ occurred under the Friuli microearthquake network during 1978 to 1986, and did not bring about any evident changes in velocity and elastic parameters. This seems to indicate that, in the Friuli seismic area, the velocities and tidal strain amplitudes display significant changes only prior to earthquakes of $M > 4$ and that the tidal strain changes are more sensitive than the velocity changes, supporting the theoretical results obtained by Beaumont and Berger (1974).

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