The Friuli (NE-Italy) tilt/strain gauges and short term observations

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Abstract

The tilt/strainmeter network of the Department of Earth Sciences of the University of Trieste, has by now a long history of records, the Trieste Grotta Gigante horizontal pendulum station having been set up in 1959 and the Friuli tilt/strainmeter stations in 1977. Since then the stations have been continuously recording the strain-rate in one of the most seismic areas of the Alpine arc, giving invaluable information on crustal deformation in a tectonically active area. Although maintaining essentially the same mechanical features from the time of installation, the instrumentation has undergone modernization, in order to apply recent technical developments to the network. This regards mainly data acquisition, which now, except for one station, is digital. The data are all available and are stored in the Deformation-Database of the Department of Earth Sciences. At first a description of the essential technical and mechanical properties of the instrumentation constituting the network is given. The mean power spectrum of all instruments covering five decades is presented, which is a powerful means to compare the quality of different stations. Following theoretical considerations of the expected pre- or coseismic deformation accompanying local events, the observations of those events is presented, which ought to give the greatest signals. The coseismic steps are modeled for those events for which a fault plane solution was available.

Key words *tiltmeter – strainmeter – short term observations*

1. Introduction

The tilt/strainmeter network of the Department of Earth Sciences, University of Trieste is installed in the seismic zone of Friuli and south to this area, near Trieste, where the observation of deformation of the crust is of extreme interest with regard to the research on preseismic phenomena and in general on the actual movements in seismic zones (Zadro, 1978, 1980; Ebblin and Zadro, 1979; Ebblin, 1986).

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The instrumentation which monitors deformation in the testfield includes at best 10 tiltmeters, 3 strainmeters and 2 horizontal pendulums sited in Trieste. Currently (1998), the tilt/strainmeter network operates a reduced number of instruments (2 tiltmeters, 3 strainmeters, 2 horizontal pendulums) due to budget shortages. The seismicity has been recorded by the microseismicity network of Osservatorio Geofisico Sperimentale (OGS) since 1977, complete to magnitude 1.8 in Friuli and by the Istituto Nazionale di Geofisica (ING) national seismologic service. Recently (1995), the Department of Earth Sciences set up a network of strong motion seismographs in the same area. A vast number of meteorological stations and groundwater measuring stations are kept up by the Regione Friuli Venezia Giulia and the Ufficio Idrografico e Mareografico di Venezia. Since 1995 radon concentration in soil has been measured continuously in one of the stations (Vil-

lanova), and a second radon station was installed at Arta Terme in collaboration with ING (Braitenberg *et al.*, 1997b; Garavaglia *et al.*, 1998).

The first instruments of the network to be operative were the Trieste horizontal pendulums, which recorded episodes of long period oscillation (4-10 min) resembling surface waves starting from 1973 (Zadro, 1978). The episodes lasted up to several hours, with an increase in amplitude and occurrence frequency up to the catastrophic 1976 Friuli earthquake (epicentral distance about 100 km). Observation of earth tides with all the instrumentation allows the study of the crustal elasticity structure and its variation with time. This is particularly interesting in connection with the 1976 earthquake, which preceded the installation of the Friuli stations by only a few months, and thus offered the possibility to observe postseismic crustal movements (Zadro, 1980) in these stations. Furthermore, it was possible to detect variations in the mechanical properties of the crust connected to the healing of the ruptured material (Mao *et al.*, 1989; Braitenberg *et al.*, 1995). More than a decade of records acquired to date have allowed the study of long term deformation of the zone, which has given very interesting results regarding the geodynamics involved (Zadro and Rossi, 1991; Rossi and Zadro, 1996; Rossi *et al.*, 1999). Another problem under study is the interpretation of short term anomalies on the instruments of the tilt/strainmeter network in the light of seismic events and the stress-conditions connected to them (Ebblin *et al.*, 1980; Mao *et al.*, 1990; Zadro, 1992; Dal Moro and Zadro, 1999). A further topic of interest concerns the influence of hydrologic factors on seismicity and on the deformational records, which has been treated in Zadro *et al.* (1987), Brussa Toi and Ebblin (1988), Braitenberg *et al.* (1993); Braitenberg and Zadro (1996), Dal Moro and Zadro (1998) and Braitenberg (1999).

2. Geologic and tectonic setting

Friuli is located on the NE border of the Adriatic plate, in the eastern part of the Southern Alps, where the overlap with Dinaric structures takes place (Barbano *et al.*, 1985). The Southern Alps are interpreted, in global tectonics, as the result of the continuation of relative motion between the European plate and the Adriatic microplate. These movements are still active, as demonstrated by neotectonic structures and by seismic activity. The seismicity is currently limited to the upper 15 km of depth. The northward movement of the Adriatic plate in the Neogene and Quaternary involved increasing areas of the Southern Alps. This should have caused intense shortening, mainly in the post-Hercynian cover, with a build up of overthrusts verging towards the south and its probable detachment from the Paleozoic basement in the more southern sector. In the pre-alpine area especially, Dinaric overthrusts have been split up and incorporated in the South-Alpine overthrusts and partly re-utilized in South-Alpine tectonogenesis, causing very complicated structural relationships. The most important regional features are the E-W striking South-Alpine overthrusts (mainly in the NW) and the SE-NW striking Dinaric overthrusts (mainly in the SE and buried in the Friuli plane). These structures are intersected by subvertical faults, striking about N-S and often showing strike-slip motion. Both structures (overthrusts and strike-slip faults) are compatible with the actual stress field, with NNW-SSE to N-S maximum compressive stress (Zanferrari *et al.*, 1982).

3. The observational sites of the tilt/strainmeter network

The tilt/strainmeter network (fig.1) consists of 5 tiltmeter stations each equipped with two Marussi tiltmeters, located in different parts throughout the seismic Friuli zone. In one of these stations (Villanova) three Cambridge-wire strainmeters are installed. The Trieste station, 80 km to the south of the seismic zone is set in the Grotta Gigante of the Trieste Karst, unique for its high vault, is equipped with two long period Zöllner type horizontal pendulums.

The aim of the tilt/strainmeter network sets a number of definite requirements on the selection of the observation sites and on the characteristics of the instrumentation. Regarding the selection of the sites, special attention had to be
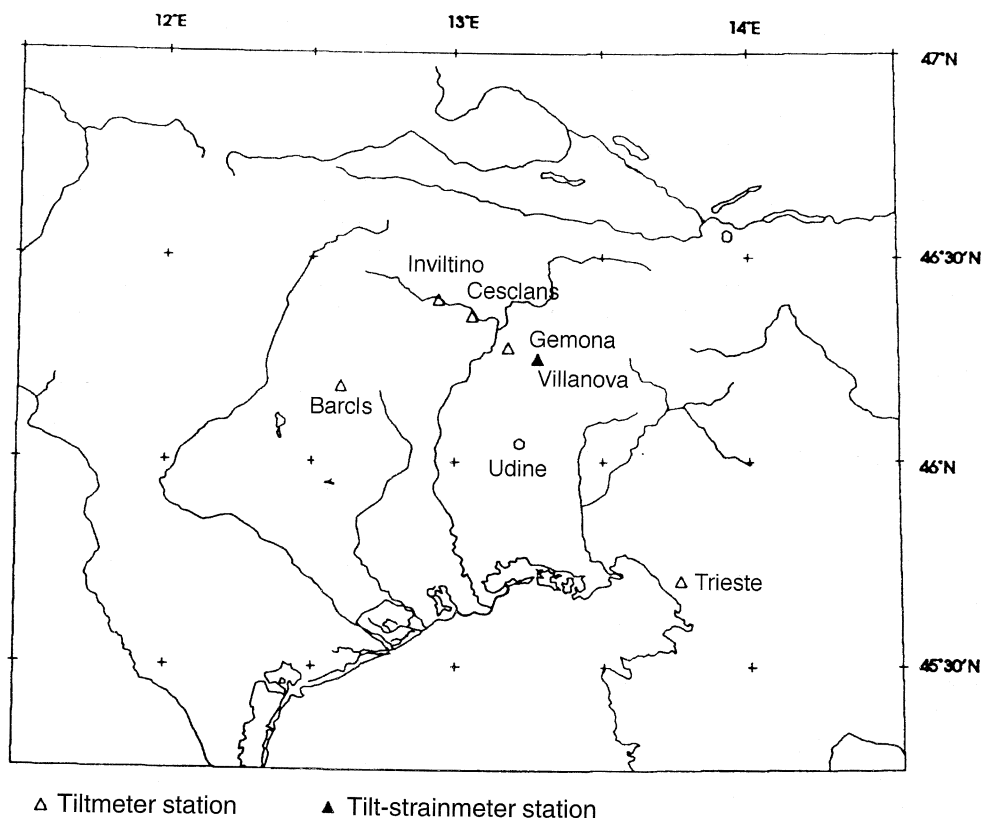
Δ Tiltmeter station   ▲ Tilt-strainmeter station

Fig. 1. Friuli region with the location of the stations of the tilt/strainmeter network. The stations equipped with a pair of Marussi tiltmeters are Barcis (BA), Invillino (IN), Cesclans (CE), Gemona (GE) and Villanova (VI). The VI station also houses the horizontal strainmeters and a radon monitoring device. The Trieste (TS) station is equipped with the large horizontal pendulums of the natural «Grotta Gigante» cave.

paid to thermal stability and insolation to human noise. This was obtained by the installation of the stations in natural caves or old military fortifications from the 30's. The description of the stations and their geographic coordinates are given in tables Ia,b.

4. Instrumental description

In the following the relevant instrumental constants are given. The data acquisition was transformed from analogical on paper to digital in all Friuli-stations by the year 1990. A block scheme of the configuration of the up to date data acquisition in the VI station is given in fig. 2. The set-up in the other Friuli tiltmeter stations is similar, except for the fact that they do not include the parts attaining to the strainmeters and the radon counter. The electronic system is switched on every hour for 4 min, during which data acquisition is performed, giving hourly data values after application of a software antialiasing filtering process. The eigen periods of the tiltmeters are discussed in the next section.

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Table Ia. Geographical location and geological setting of the stations and instrumentation.

Geographical location	Code	Year of installation	Geologic setting	Instrumentation	dT/year	Remarks
Villanova delle Grotte	VI	Strainmeters: Dec. 1978 Pendulums: Jan. 1997	Natural cave in slab of limestone overthrusted into Eocene flysch	Strainmeters: ST2:N128E ST3:N27E ST4:N67E Tiltmeters: VINS, VIEW	1.5 °C/yr	South of main overthrust of Periadriatic system; highly fractured area
Cesclans	CE	July 1977	Old military fortification built in the 1930's	Tiltmeters: CENS, CEEW	3-4 °C/yr	Zone highly fractured by overthrusts oriented E-W and by vertical faults striking N-S. Very near 1976 earthquakes
Invillino	IN	July 1979	Old military fortification built in the 1930's	Tiltmeters: INNS, INEW		
Gemona	GE	Sept. 1980		Tiltmeters: GENS, GEEW	2-5 °C/yr	
Barcis	BA	Dec. 1978	Natural cave some meters deep in Paleogene limestone	Tiltmeters: BANS, BAEW		Close to Periadriatic overthrust WSW of main seismic area
Borgo Grotta Gigante	TS	May 1959	Natural cave «Grotta Gigante» of the Trieste Carst. Vault 112 m high	Long Period horizontal pendulums: TSNS, TSEW	Less than 1.°C/yr	Reference station 80 km SE of seismic zone

Table Ib. Coordinates and altitude of the stations of the Friuli tilt/strainmeter network and the Trieste Grotta Gigante station.

Code	Latitude	Longitude	Altitude
VI	46°15′23″N	13°16′53″E	616 m
CE	46°21′23″N	13°03′30″E	351 m
IN	46°24′10″N	12°57′09″E	480 m
GE	46°16′19″N	13°09′50″E	825 m
BA	46°11′09″N	12°36′01″E	520 m
TS	45°42′30″N	13°45′48″E	274 m

4.1. *Marussi tiltmeters*

The Marussi tiltmeters are traditional horizontal pendulums with Zöllner type suspension, constructed at the Istituto di Geodesia e Geofisica (now merged into the Department of Earth Sciences) University of Trieste. Figure 3 shows a schematic view of the instrument and table II gives the corresponding geometrical dimensions and essential mechanical characteristics.

The determination of the magnification factor is accomplished considering the expressions which give the period of oscillation of the pendulum in the horizontal and vertical plane. With the moment of inertia J_0 with respect to the actual axis of rotation, R the distance of the centre of gravity from the actual axis of rotation, i the angle of inclination of the actual axis of rotation from the vertical, and g the local gravity value, the period of oscillation T_o in the horizontal plane of the pendulum is equal to
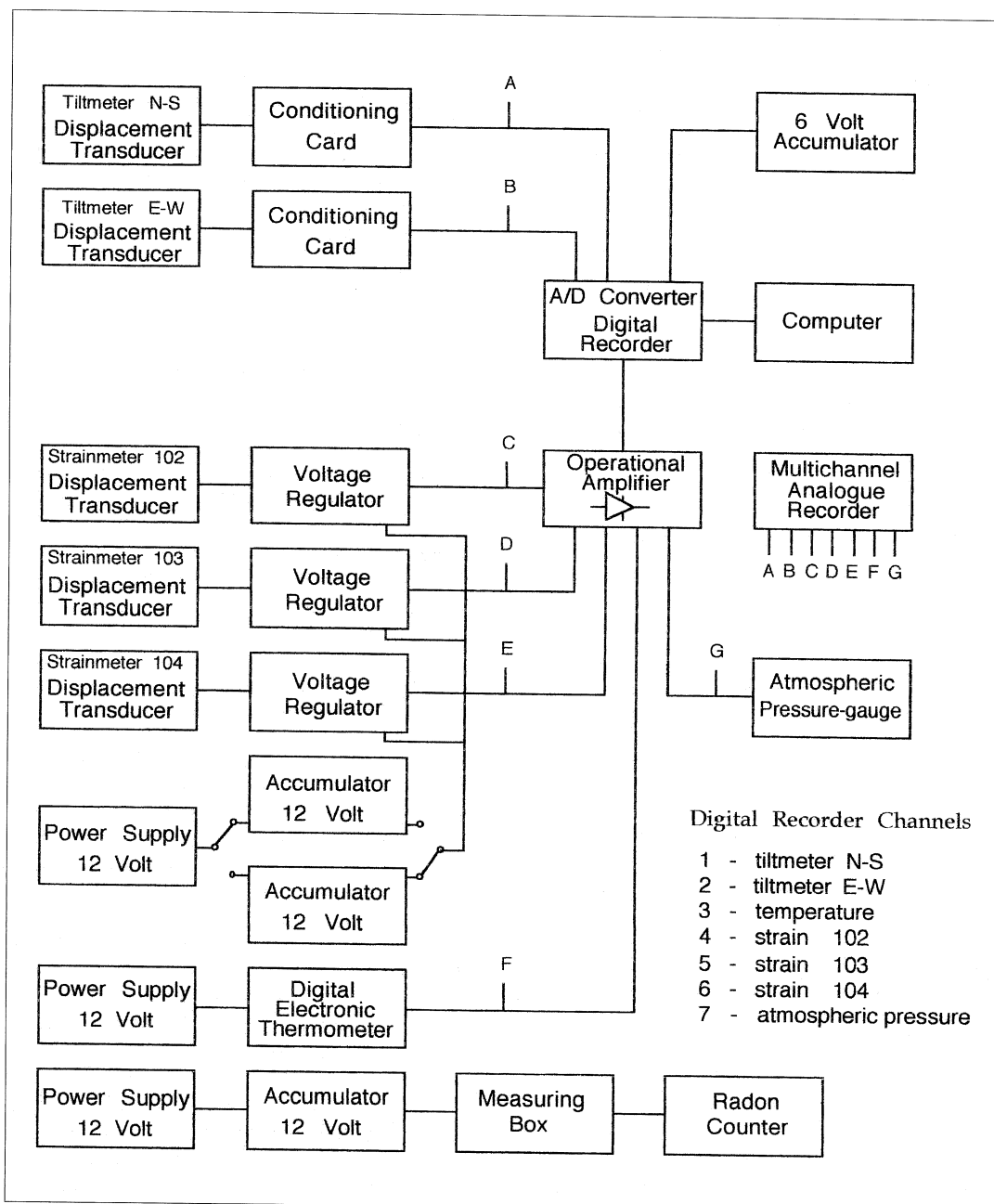
$$T_o = 2\pi\sqrt{\frac{J_0}{mgR\sin(i)}}.$$

Fig. 2. Block diagram of the data acquisition system of the Friuli station Villanova.

Ra = Distance centre of gravity from rear suspension.

d = Distance actual axls of rotation from rear suspension.

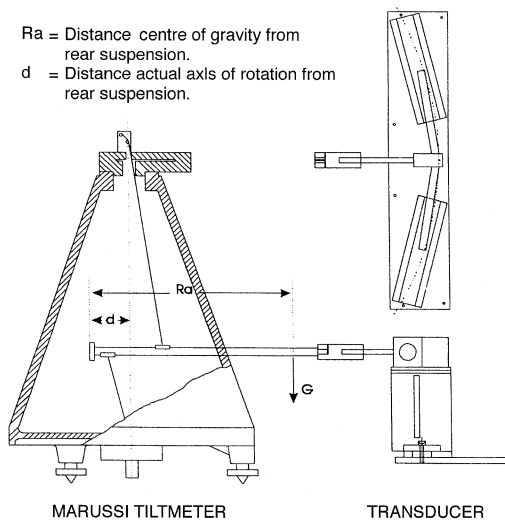MARUSSI TILTMETER   TRANSDUCER

Fig. 3. Schematic figure of the Marussi tiltmeter. Cross-section of the transducer built by Gruppo Tecniche Avanzate, Trieste.

With R_a the distance of the centre of gravity from the rear suspension, the period of oscillation T_a in the vertical plane of the pendulum fixed at the rear suspension is given by

$$T_a = 2\pi \sqrt{\frac{J_0 + mR_a^2 - mR^2}{mgR_a}}.$$

From the last two equations, we obtain the expression for the static magnification A of the instrument, given by

$$A = \frac{1}{\sin(i)} = \frac{T_o^2}{(2\pi)^2} \frac{gR}{\dfrac{T_a^2}{(2\pi)^2} gR_a + R^2 - R_a^2}.$$

The magnification of the measured tilt on the recording is determined by the distance between the recording device and the actual axis of rotation. For the analogical (up to 1990) recording this was determined by the distance recording pen-actual rotation axis and was equal to

$R_d = 0.5157$ m. For the digital recording it is given by the distance between inductive displacement transducer and actual rotation axis, which is equal to $R_d = 0.5166$ m. The linear variable differential transformer was built by Gruppo Tecniche Avanzate in Trieste with good linearity and temperature coefficients (table III). The values given in table III are those for a typical recording situation (magnification $A = 5000$).

4.2. *Cambridge wire-strainmeters*

The three strainmeters installed in the Villanova cave are of the type Cambridge Wire Strainmeter. A wire strainmeter of this type is described in detail in King and Bilham (1976). It consists of an invar wire held in tension by a frictionless balance. Strain in the ground appears as rotations of the balance which are detected using an inductive displacement transducer. Table IV gives the technical details of the instrument.

4.3. *Marussi horizontal pendulums of Grotta Gigante*

The horizontal pendulums of the Grotta Gigante station (TS), were extensively described in 1959 on occasion of the 3rd International Earth Tides Symposium, held in Trieste (Marussi, 1959, 1960). The instruments were subsequently completely revised in 1966. The Grotta Gigante pendulums are horizontal pendulums with Zöllner type suspension, as are the Friuli tiltmeters, though very different in two principal aspects: the first is the fact that for the horizontal pendulums in the Grotta Gigante the upper and lower attachments are fixed to the solid rock of the cave, ensuring great stability of the wires which support the pendulums. This causes a principal difference in the measurement of local deformation, a point discussed in Ebblin and Zadro (1980). The second fact regards the natural period of the pendulums, as the exceptional height of the cave (112 m), allowed the Grotta Gigante instruments to attain a long natural period, thus protecting the instruments from the action of short period micro-

Table II. Technical description of Marussi tiltmeters.

Quantity	Symbol	Value
Distance between upper and lower mountings		0.5 m
Weight of housing case		45 kg
Total weight of the pendulum, including the wires	m	0.679 kg
Distance of the centre of gravity from the rear suspension	R_a	0.32 m
Distance of the actual axis of rotation from the rear suspension	d	0.039 m
Moment of inertia with respect to the actual axis of rotation	J_0	$5.5 \cdot 10^{-4}$ kg m^2
Period of oscillation in the vertical plane, determined experimentally	T_a	1.3 s
Period of oscillation in the horizontal plane	T_o	Maintained to 90 s
Recording: – before 1990: analogical – after 1990: digital	Hourly sampling rate – scintillation on sparker paper – inductive displacement transducer	

Table III. Properties of the tiltmeter transducer and resolution of the analogical and digital recording. Values for magnification $A = 5000$.

Measuring range of the transducer	$8 \ 10^3$ ms
Temperature coefficient of the transducer	4 ms/°C
Resolution of the transducer	Better than 1 ms
Recording: until 1990:	Analogical on paper. Amplification 80 ms/mm
after 1990:	Digital. 1 ms/digit

Table IV. Technical description of the strainmeters. Values for a strainmeter of 10 m length.

Orientation and length: ST2 ST3 ST4	 Azimuth = N128E Length = 13.06 m Azimuth = N27E Length = 12.63 m Azimuth = N68E Length = 14.33 m
Diameter of invar wire	0.56 mm
Tension of wire	Adjustable in the range 15N-20N
Temperature coefficient	$5 \ 10^{-7}$ strain/°C
Pressure coefficient	10^{-9} strain/mbar
Stability	Better than 10^{-7} strain/yr
Resolution	About 10^{-9} strain
Recording: until 1990:	Analogical on paper. Amplification $2 \ 10^{-9}$ strain/mm
after 1990:	Digital and analogical on paper. 10^{-10} strain/digit

Table V. Technical description of Marussi type horizontal pendulums.

Quantity described	Symbol	Pendulum A (EW)	Pendulum B (NS)
Distance between upper and lower mountings		94.9 m	95.5 m
Total weight of the pendulum, including the wires	m	18.374 kg	18.340 kg
Distance of the centre of gravity from the rear suspension	$\cdot R_a$	1.418 m	1.417 m
Distance of the actual axis of rotation from the rear suspension	d	0.173 m	0.175 m
Moment of inertia with respect to the actual axis of rotation	J_0	30.72 kg m^2	30.55 kg m^2
Period of oscillation in the vertical plane, determined experimentally	T_a	2.46 s	2.46 s
Period of oscillation in the horizontal plane, maintained to	T_o	360 s	360 s
Operation mode		Critical damping	Critical damping
Recording		Analogical on photo-graphic paper	Analogical on photo-graphic paper

Table VI. Resolution of the Marussi type pendulums.

Period T_o measured	Static magnification	Magnification on the recording	Resolution
360 s	24 000	0.9 ms/mm	0.5 ms

seisms. In the following, the essential mechanical characteristics of the Grotta Gigante pendulums are given.

Inserting the values given in table V in the expressions given in paragraph 4.1, we obtain for the period of oscillation in the horizontal plane T_o and the static magnification A the values given in table VI.

The modernization of the acquisition to a digital system was started in 1996 in collaboration with Dr. G. Romeo and Dr. Q. Taccetti from the Istituto Nazionale di Geofisica, Rome. The reflected light of a solid-state laser beam pointing on the mirrors also used for the analog system is recorded with a CCD. The sampling was fixed at 25 samples/s, and was tested on the NS component. We have obtained very good recordings of several large magnitude events ($M > 5$) which agree very well with the time integral of the recordings of a VBB seismograph installed at the same site.

4.4 *Data availability*

In order to show the availability of the entire data set, in fig. 4 for each instrument the horizontal line indicates the functioning of the instrument, while interruptions are shown in blank. Regarding all data, it may be seen that the network has produced a very important datum set in a highly seismic region. Integration of our recordings with those of the local seismic network (OGS) and the meteorological stations of the Regione Friuli-Venezia Giulia gives invaluable information for the study of geodynamic crustal processes. The entire data set is available with an hourly sampling rate.

5. Spectral analysis

A way to describe the quality of the records of the different stations is to calculate the mean

AVAILABILITY OF THE RECORDS
OF THE FRIULI TILT/STRAINMETER NETWORK (YEARS '76-'96)
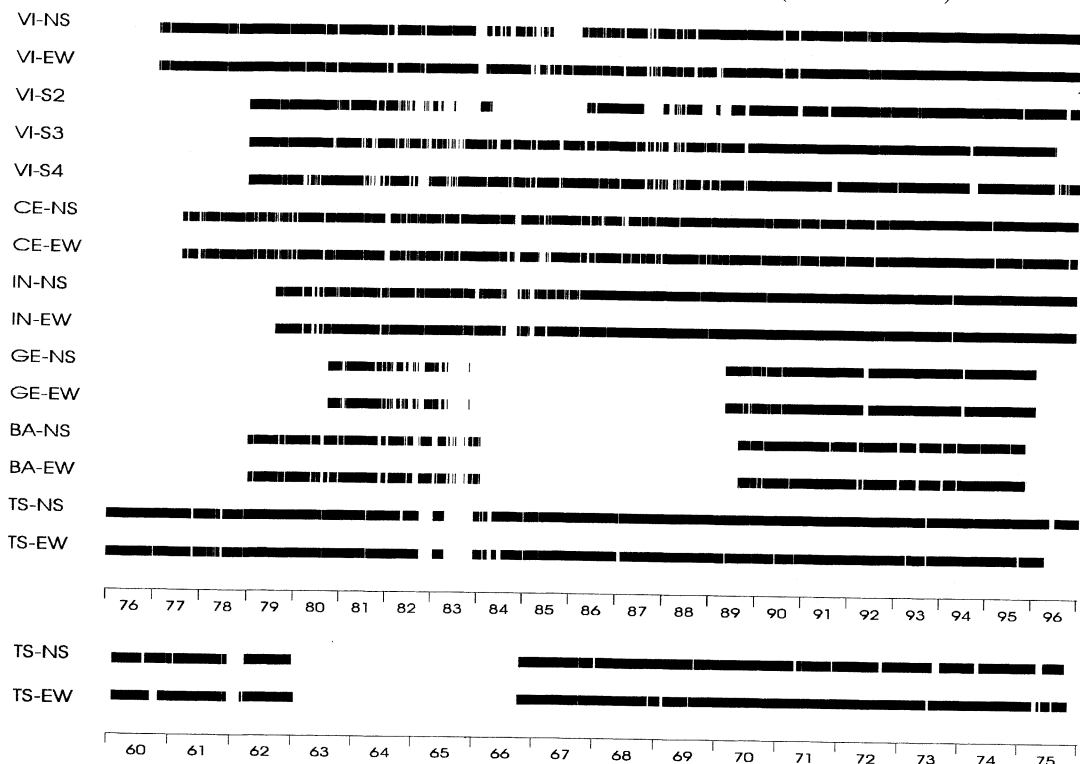AND OF THE GROTTA GIGANTE (TRIESTE) PENDULUMS (YEARS '60-'96)

Fig. 4. Availability of the entire data set of the tilt/strainmeter network for the years '60-'96. The codes for each station and instrument are given in tables Ia,b.

Fourier spectrum, sometimes also called noise spectrum. According to the data availability (see fig. 4), the spectrum can be calculated on up to 5 decades in frequency. The spectrum is calculated on two different data sets for the different frequency bands: one year of hourly data are used in the band $2.10^{-7}-10^{-4}$ Hz, and the mean spectrum over several years is calculated. We chose to use only the years with digital data acquisition, *i.e.* the years 1991-1996. The low frequency part of the spectrum is calculated on daily sampled data (Rossi and Zadro, 1996). The longest series pertains to the Grotta Gigante pendulums (1967-1996). The other low-frequen-

cy spectra are evaluated on shorter time periods, according to data availability (VI tilt: 1978-1996; VI strain ST2: 1987-1996; VI strain ST3 and ST4: 1979-1996; CE tilt: 1978-1996; IN-tilt: 1980-1996; GE and BA tilt: 1990-1995). The overall spectra, obtained connecting the low frequency part with the high frequency part are given for the NS and EW tilt spectra of all stations in fig. 5a,b respectively, and for the three strainmeters in fig. 5c. The two parts of the spectrum connect together very well for the Grotta Gigante pendulums. A small downward shift of the high frequency part with respect to the low frequency part is observed for the other
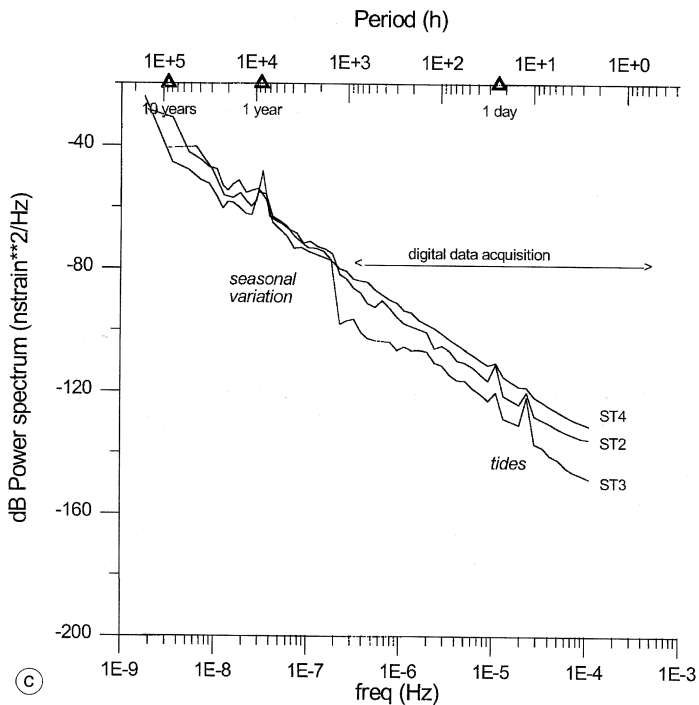
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(a)

(b)

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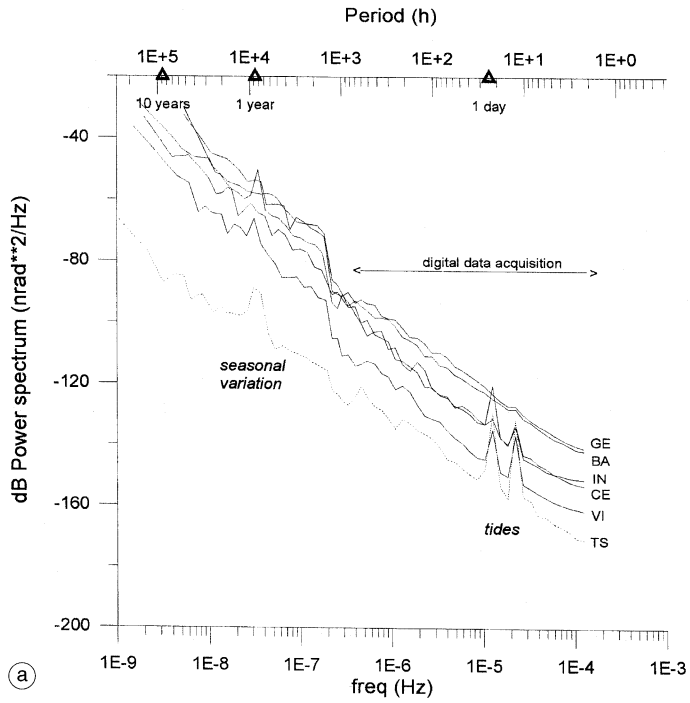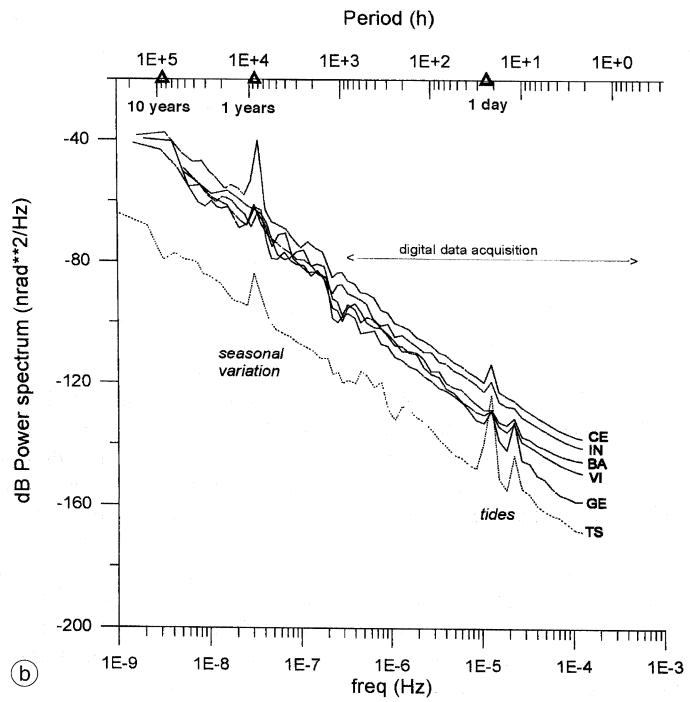
Fig. 5a-c. Mean spectra of tilt and strainmeter records. a) NS-tilt spectrum; b) EW-tilt spectrum; c) extensometer spectrum for strainmeters ST2, ST3, ST4.

tilt stations, which can be attributed to the fact that the high frequency part is calculated on digitally recorded data, the noise level of which is lower than the previous recording system. All spectra show a linear decay of spectral energies in the log-log representation. The mean linear factor (m) is -10.7 ± 0.7 dB/log (frequency) for the Friuli tilt stations. This is equivalent to approximately a $1/f$ decay of the spectral energy with frequency. The decrease is slightly less steep (-8.5 ± 0.4 dB/log (frequency)) for the Grotta Gigante pendulums and is equivalent to a decay of $1/f^{0.8}$. This result is in agreement with the tilt and strain spectra given by other investigations (Berger and Levine, 1974; Wyatt and Berger, 1980; Wyatt *et al.*, 1984). The most evident feature in all curves is the spectral energy at tidal frequencies (diurnal and semi-diurnal band), and at the yearly cycle. By far the quietest records are furnished by the Grotta Gigante pendulums, which can be attributed to their deep underground installation and long eigen period.

The analysis of the spectral density is interesting in connection with the problem of detecting tectonic signals, in particular the pre-, co- or postseismic deformations. For a signal to be significant it should be greater than some multiple of the noise level. Apart from the tidal frequencies, the calculated mean spectra can be interpreted as the noise spectra. As seen from the spectrum, the noise level is frequency dependent, showing that a fast tectonic signal is more likely to be detected than a slow signal of the same amplitude. The integration of the spectral density over a certain frequency band determines the noise level over that band.

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6. Short period signals

6.1. *Seismic events and the expected deformational signal*

The short period deformation signal to be expected in connection with a seismic event may be estimated in different ways, based on either theoretical or empirical considerations.

The theoretical models considering pre- or coseismic deformations divide into two groups, one considering the fault constitutive (stress-strain) relations as a function of slip or slip rate, neglecting any change in the physical properties of the crustal material, as the model of Tse and Rice (1986) of a strike slip fault. Stuart and Tullis (1995) applied the theory to estimate the borehole dilatational preseimic deformation at Parkfield (California). The model predicts the accelerated deformation rate in the days before the main shock, with an amplitude above the noise levels of today's dilatometers. The other group is based on the bulk rock constitutive relations, which predict changes in the physical-mechanical properties of the material surrounding the fault, involving essentially the development of cracks, with or without the influence of fluids (Scholz *et al.*, 1973; Hudson and Knopoff, 1989; Mjachkin *et al.*, 1975). The presence of the earthquake preparation volume can be considered a soft inclusion in the medium, which produces a strain signal, if a certain regional stress is assumed. This strain signal was estimated by Dobrovolsky *et al.* (1979), who sets a relation between the maximum hypocentral distance (R_ε, in km) at which the signal may be detected as a function of the earthquake magnitude (M). Assuming a variation of the shear modulus of 10% and assuming a strain detectability limit of 10 nstrain, he finds $\log(R_\varepsilon) = 0.43\,M$. The above relation is valid for distances greater than 2 fault lengths, as in the near-field the strainfield is very inhomogeneous and azimuth dependent, and no simple magnitude-distance relation can be used to estimate the expected deformation.

An empirical magnitude-distance relation for precursory signals was defined by Takemoto (1991), after collecting and analyzing the strain-tilt measurements made in Japan. Considering all strain observations available at the time of an earthquake in Japan, and by classifying the observations according to whether a precursor was observed or not, he came to the conclusion that a precursor may be observed if the hypocentral distance (r in km) is less than the empirical relation $\log r = 0.5\,M - 1.96$. From the relations which relate the magnitude to the source dimension, Takemoto follows that no precursor is liable to be observed in Japan outside the radius of about two source dimensions.

Comparison of the magnitude-distance relations of Dobrovolsky and Takemoto show that the theoretical calculations of Dobrovolsky predict a much greater radius in which precursors are liable to be observed, than the limit distance obtained from the observations collected by Takemoto (1991).

The distance dependence of coseismic permanent deformation was studied by Wyatt (1988). Using a dislocation model of a strike-slip fault he found that outside the source regime, the deformation is given by the expression: $\log(\varepsilon_{rms}) = 1.5\,M_w - 3\log r - 11\ (\pm 0.3)$, with ε_{rms} either the root mean square (rms) of several independent components of strain at one location or the rms of any strain component, when averaged around a circle of constant hypocentral distance. r is the hypocentral distance (in km), and M_w is the moment magnitude (Hanks and Kanamori, 1979). An important result of the above relation is that the coseismic deformation decays with the third power of the hypocentral distance.

Common to the relations of Dobrovolsky and Wyatt is the slope of the linear relation of log-distance (in km) to magnitude, for a constant deformation. This slope is equal to 0.5, whereas the slope of the empirical relations of Takemoto is slightly inferior and equal to 0.43.

Regarding the Friuli seismicity, the hypocentral distance to the nearest functioning tilt or strainmeter for all local events with $M \geq 3$ is shown in fig. 6 for the years 1977-1997, plus the important Slovenian event of 1998. The eight events with the most favorable magnitude-distance relation are marked with the year of occurrence. The seismicity refers to the aftershock-depleted catalogue, according to the standard method of Reasenberg (1985). A cluster of earth-
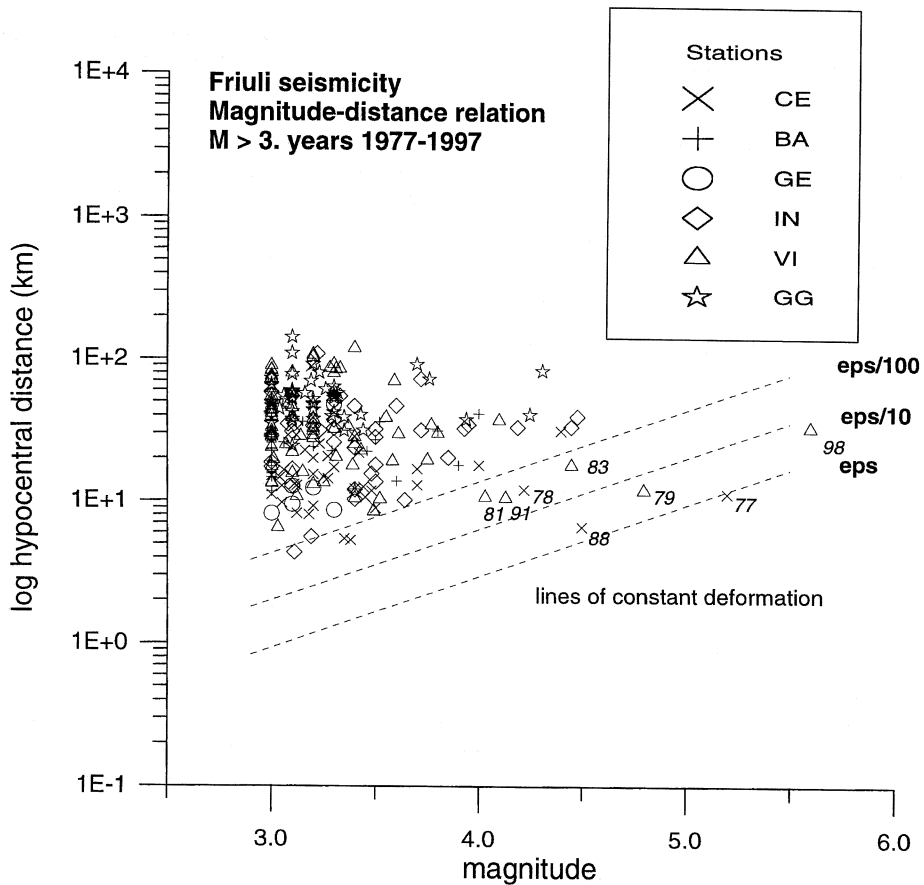
Fig. 6. Hypocentral distances to nearest functioning station for Friuli events with $M \geq 3$ for the time interval (1977-1997) plus the Slovenian event of 1998. Also shown are lines of constant deformation for the deformation level eps, and eps/10, eps/100, with eps = 2 10^{-7}.

quakes is replaced with a single event, representative of the total released energy. The OGS-catalogue was used, adopting for the years 1977-1988 the revised edition (Renner, 1995). Also shown are the lines of constant pre- or coseismic deformation according to Wyatt (1988), at a level of deformation eps, eps/10 and eps/100. Eps is equal to 2 10^{-7}, the expected deformation of the 1977 event, the most favorable event according to the distance-magnitude relationship. None of the events falls into the range of distances, for which according to Takemoto precursors could be observed.

6.2. Observation and modeling of selected events

In the following, the events with the most favorable magnitude-distance relation are discussed, and the strain signal observed before, after and with the event is shown. The hypocentral determinations of the events and the distance to the nearest stations are reported in table VII. In order to conform to the existing literature, both the hypocentral determination and magnitude of the mainshock and that of the equivalent event are reported in the table.

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Table VII. Eight most favorable events (depleted catalogue) with respect to the magnitude-distance relationship: the fault mechanisms refer to: (B) = Barbano *et al.* (1985); (K) = Kravanja *et al.* (1994). Magnitude (*M*), depth (*h*), hypocentral distance to nearest station (*d*). Hypocentral determinations belong to a) OGS and b) equivalent event obtained from the aftershock analysis (where possible).

Event	North.	Lat.	East.	Lon.	M h (km) year mo day time	d (km)	Mechanism
1a	46°	16.12'	12°	59.99'	5.2　8.24 1977 09 16 23:48		EW-inverse or NW-SE strike slip (B)
1b	46°	18.70'	12°	58.4'	5.2　8.2	11.2	
2a	46°	17.92'	13°	09.46'	4.2　4.4 1978 04 03 10:49		
2b	46°	19.32'	13°	10.33'	4.2　7.2	12.0	
3a	46°	20.28'	13°	17.14'	4.8　6.7 1979 04 18 15:19		NW-SE strike-slip (B)
3b	46°	22.06'	13°	15.38'	4.8　9.42	11.8	
4a	46°	20.6'	13°	07.28'	3.38　1.53 1980 12 01 09:20		
5a	46°	24.25'	13°	0.38'	3.1　1.24 1986 02 09 17:51	5.3	
6a	46°	20.88'	13°	04.56'	4.2　5.1 1988 02 01 14:21	4.3	EW overthrust (K)
6b	46°	21.5'	13°	05.84'	4.5　5.8	6.6	
7a	46°	14.58'	13°	18.54'	3.9　19.5 1991 10 05 14:56		
7b	46°	15.41'	13°	18.54'	4.1　10.3	10.5	
8a	46°	18.6'	13°	39.6'	5.6　9.8 1998 12 04 10:55		NW-SE strike slip (OGS)
8b	46°	17.25'	13°	40.8'	5.66　8.76	29.7	

We can attempt to model the observed co-seismic deformation by a dislocation model. This model consists of a homogeneous elastic halfspace characterized by equal Lame's parameters λ and μ in which a rectangular plane surface (length L and width W) posed at a certain depth (d) is dislocated by an amount (u). From the combination of gravity and seismic data the rigidity in the Friuli area in the upper 5 km was found to be about 18 GPa (Dal Moro *et al.*, 1998). The product of fault area and average slip is proportional to the seismic moment M_0, as seen in Brune's relation $M_0 = \mu u S$ (Brune, 1968). For the Friuli events though only the local magnitude is given, which can be related to the seismic moment through the seismic moment magnitude M_w which in turn is rela-ted to the local magnitude by $M_w = M_L - 0.07$ (Wyatt, 1988), and to the seismic moment (in Nm) by $\log M_0 = 1.5 M_w + 9.1$ (Hanks and Kanamori, 1979).

One problem is to define the geometry of the fault plane and the average slip. For the Friuli events, no published constraints are given, besides the magnitude and the fault-plane solutions for some greater events. We may though refer to general earthquake scaling relations, in order to confine the class of models, compatible with the seismological record. Scholz (1990), p. 182 discusses this problem, and shows that on average the logarithm of the rupture length (L) is linearly connected to the logarithm of the seismic moment, and may be approximated by $\log L = 0.33 \log M_0 - 5.1 \pm 0.3$ (L in km, M_0 in

Nm). This empirical relation is in accordance with the more recent compilations by Wells and Coppersmith (1994), but they limit the analysis to medium and large-sized earthquakes ($M \geq 4.8$). It follows that the average slip (u) can be obtained from Brune's relation, when some assumptions on the aspect ratio (ratio of the rupture length to width) of the fault are posed.

a) *16 September 1977, M = 5.2*

The event occurred 16 months after the disastrous May 6, 1976 ($M = 6.4$) Friuli earthquake, and shortly after the first tilt meters were installed, the only operating stations being VI and CE. The event has had some attention in the past, as the nature of the fault mechanisms created some dispute (Barbano *et al.*, 1985), on whether it was Alpine (EW-strike) or Dinaric (NW-SE strike). The aftershock sequence was studied with a method termed the principal parameters analysis (Ebblin and Michelini, 1986; Rossi and Ebblin, 1990). The method is based on the assumption that the spatial distribution of groups of successive hypocenters provides some information on the geometry of the rupturing fault system. On a sliding time window the scatter matrix is computed and interpreted as defining an ellipsoid, whose axes are the positive square roots of the eigenvalues. The principal axes of the ellipsoid are aligned with the three eigenvectors, respectively.

The analysis showed that the first part of the sequence defines a Dinaric strike with shallow NE dip. We tested the compatibility of the different fault plane solutions obtained from the seismologic records and the fault orientation inferred from the aftershock distribution with the tilt measurements. The coseismic deformation was recorded by the NS tilt component of VI and both NS and EW tilt components of CE (see fig. 7), at hypocentral distances of 26 km and 11 km, respectively. Both the fault mechanisms obtained from the seismologic records (Dinaric strike slip and alpine thrust) are incompatible with the tilt observations. In fact, using a mean fault size and slip (according to Scholz, 1990), wrong amplitudes and signs are predicted. Table VIII gives event number, as defined in

table VII, the fault strike (E from N), depth, dip (downwards from horizontal), the length and width, the components along strike and apposite dip of the slip vector (positive upwards) and the modeled and observed tilt and/or strain steps. The fault orientation obtained from the aftershock distribution can partly explain the tilt observations. As the rake of the slip vector is not constrained, the rake (upwards from horizontal) best approximating the observations was searched for and was 98°. The fault length and width was estimated from the ellipsoid approximating the aftershock distribution to be about 5 km. A slip of 0.7 m then predicts the tilt signal at CE well (see table VIII, event 1c). The orientation of the NS tilt at VI is correct, but the observed value is more than one order of magnitude greater than the predicted one. Figure 7 shows the recordings of the NS and EW tilt components in the stations CE and VI, together with the rainfall (daily sampling) and the water table variation (3 days sampling). The modeled tilt steps are added to the graphs. In the days following the event, tilt steps are observed in one or both stations. They occur at times of aftershocks, but could not be modeled due to the uncertainty on the mechanism.

b) *3 April 1978, M = 4.2*

This event did not give way to significant short period signals although it has a favourable magnitude-distance relationship. The same can be said of the events of 1 December 1980, $M = 3.4$ and 9 February 1986, $M = 3.1$, the last two with magnitudes too small.

c) *18 April 1979, M = 4.8*

The observations in VI and CE (12 and 19 km distance, respectively) during this event are shown in fig. 8. Only the components which recorded the event are shown. The strain signal which occurred 3 days before the event is due to a distant earthquake (649 km) in Montenegro of $M = 7.3$ (Lat. 42.10, Lon. 19.21; April 15, 1979, 08:20. NEIC). For the event of April 18 a distinct coseismic dislocation was observed in CE
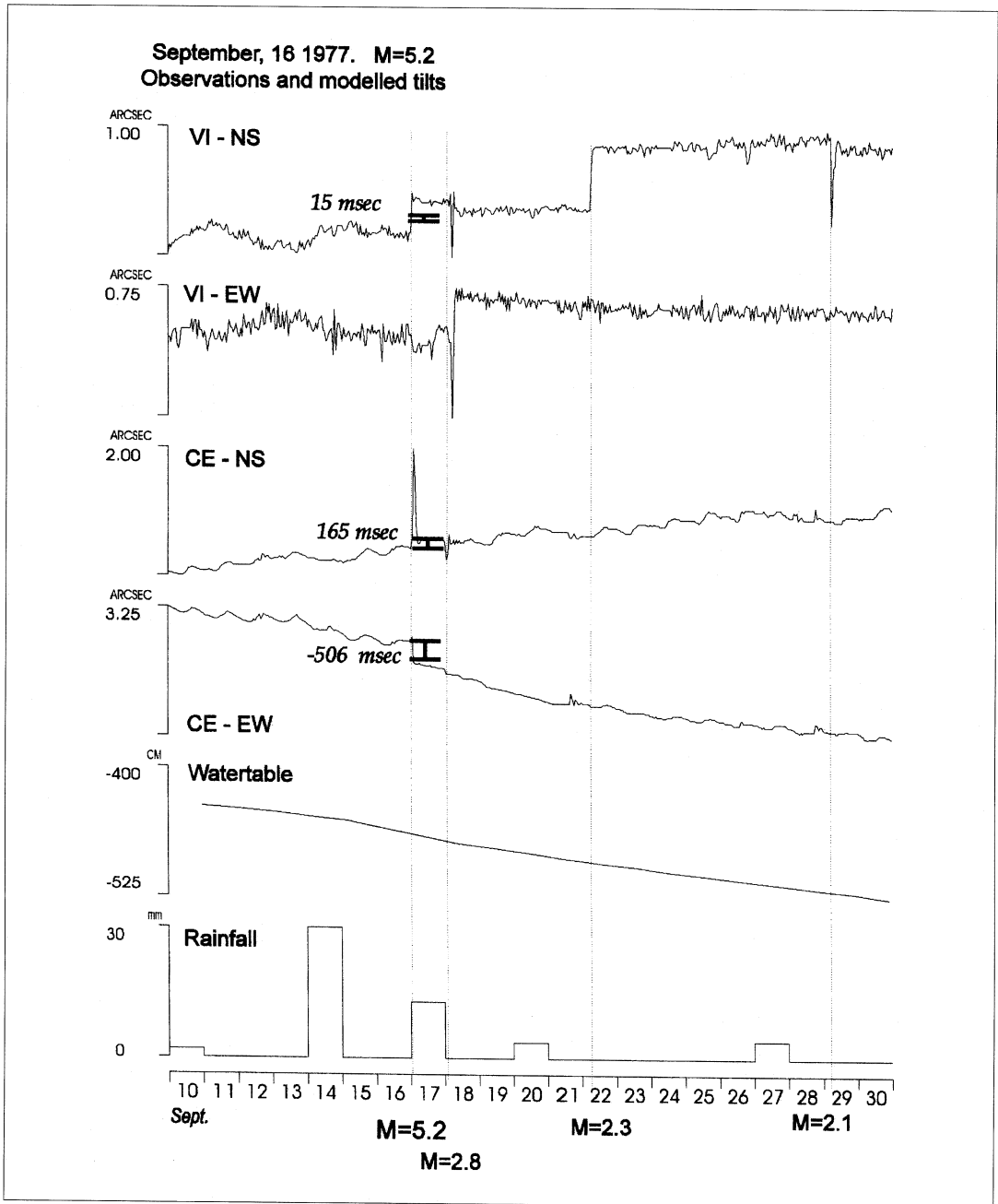
September, 16 1977. M=5.2
Observations and modelled tilts

ARCSEC
1.00 **VI - NS**

15 msec

ARCSEC
0.75 **VI - EW**

ARCSEC
2.00 **CE - NS**

165 msec

ARCSEC
3.25

-506 msec

CE - EW

CM
-400 **Watertable**

-525

mm
30 **Rainfall**

0

10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 26 27 28 29 30
Sept.

M=5.2   M=2.3   M=2.1

M=2.8

Fig. 7. Tilt recordings in VI and CE during the occurrence of the $M = 5.2$, 16 September 1977 event. The short term tilt signals following the main shock are associated with aftershocks. Also shown are the modeled tilt steps. Tilt in arcsec, watertable in cm below the surface, and rainfall in mm/day.

Table VIII. Observed and modeled coseismic strain-tilt steps for those events, for which the mechanism is available. For event 1 two mechanisms have been proposed from first arrivals: a) overthrust and b) strike slip. The analysis of the aftershock distribution furnishes a third mechanism (c). L = fault length; W = fault width, u = mean dislocation on fault ($u1$ along strike direction, $u2$, along dip direction, thrust: $u2$ positive), M_0 = seismic moment. Tilt in ms, strain in nstrain; component names as in table Ia.

Event	Strike E from N	Depth (km)	Dip	L (km)	W (km)	u1 u2 (m)	M_0 (Nm)	Model tilt NS tilt EW St2 St3 St4	Observed tilt NS tilt EW St2 St3 St4
1a	234°	8.3	22°	3.4	3.4	0 .37	$7.9\ 10^{16}$	−0.9 – (VI) 160 275 (CE)	240 – (VI) 170 −510 (CE)
1b	349°	8.3	68°	3.4	3.4	−.37 0	$7.9\ 10^{16}$	−1.3 – (VI) 506 −24 (CE)	240 – (VI) 170 −510 (CE)
1c	350°	8.3	8°	5	5	−.12 .82	$3.7\ 10^{17}$	15 – (VI) 165 −506 (CE)	240 – (VI) 170 −510 (CE)
6	76°	4	30°	2.6	2.4	−.09 .181	$2\ 10^{16}$	451 −694 (CE)	440 −690 (CE)
8	320°	8.5	66°	11	7	−.68 .60	$1.1\ 10^{18}$	16 26 103 −250 −130	73 165 130 −70 −234

and a small signal is found in the strainmeters of VI. The EW component of CE went out of range on 15 April and did not record during the subsequent days. Unfortunately, the event was not recorded by the tiltmeters of VI, as it falls into the time interval during which the recording paper was changed. Due to the scarce number of components, which recorded this event, the modeling of this event by a dislocation was not done. A model, which uses the fault mechanism obtained from the seismologic recording, does not satisfy the observations. This case did not generate a consistent aftershock sequence, so that no inferences on the geometry of the fault plane could be made from methods as the principal parameter method. The compressive strain deformation signal observed between April 24 and 30 is induced by rainfall. It is an example for the typical hydrologic induced signal, characterized by a rapid compression following a near to exponential recovery.

d) 1 February 1988, M = 4.4

One peculiarity of this event is the long aftershock sequence it generated, non-respective of the fact that the magnitude of the main shock was little more than 4 ($M = 4.2$). In fact, the aftershock sequence lasted for 26 days, counting 34 events with magnitudes $M > 2.5$. From the analysis of the aftershock sequence (Braitenberg *et al.*, 1997a) the length and width of the fault plane could be estimated). The fault plane solution was determined by Kravanja *et al.* (1994) by the method of *P*-wave first motion polarities. The method of first motion polarities gives a strike of 20° to 110° according to the model and inversion program used, of a subhorizontal thrust mechanism on a south dipping fault, which is equivalent to an alpine (E-W) mechanism. Tilt steps were observed on both NS and EW tilt components in the nearest station (CE, hypocentral distance 6.6 km). The recordings of the two nearest stations CE and IN (13.4 km distance) are shown in fig. 9. A dislocation model can explain the tilt steps observed in CE on a fault compatible with the alpine thrust mechanism (see fig. 9 and table VIII). In the two months preceding this event a southward tilt in the CE-station was observed (not shown), which was interpreted with a viscoelastic model in Dal Moro and Zadro (1999).

e) 5 October 1991, M = 3.8

One of the most complex and puzzling strain observations was recorded during the occurrence of this event. The tilt and strain records
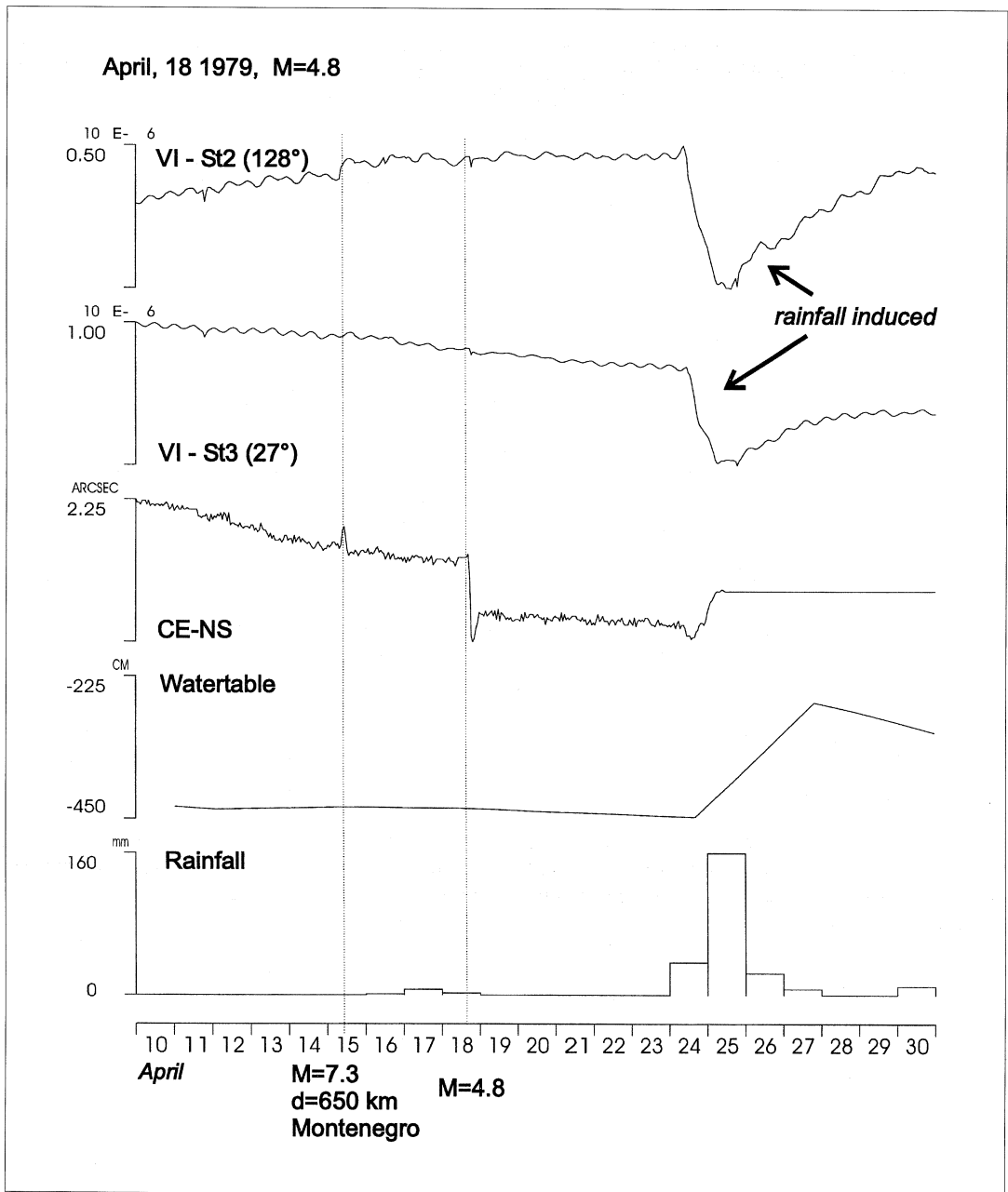
Fig. 8. Tilt and strain records of the April 18, 1979, M = 4.8 event. Also shown is the watertable variation and rainfall (see fig. 7). The short term signal on April 15 is associated to a distant earthquake of magnitude M = 7.3. The dislocation associated with the main shock is clearly observed as tilt steps, whereas the signal is small for the strainmeters. A typical rain-induced strain-signal occurs between April 24-30.
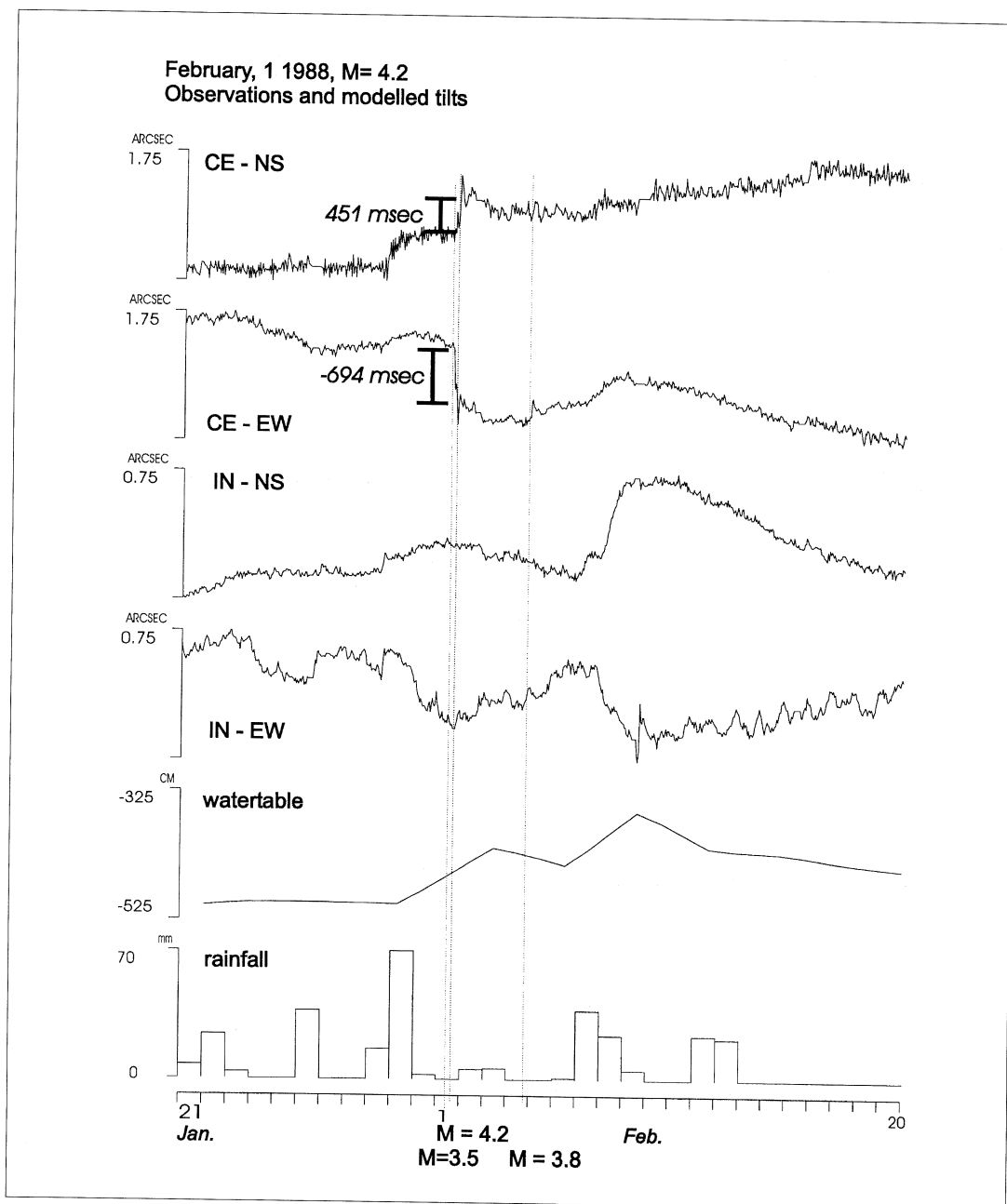
Fig. 9. Tilt records of the February 1, 1988, $M = 4.2$ event. Also shown is the watertable variation and rainfall (see fig. 7). The modeled tilt signals for station CE are shown by the inlayed vertical bars. The more distant IN station did not give any coseismic step. The main shock together with a foreshock and one aftershock are shown.
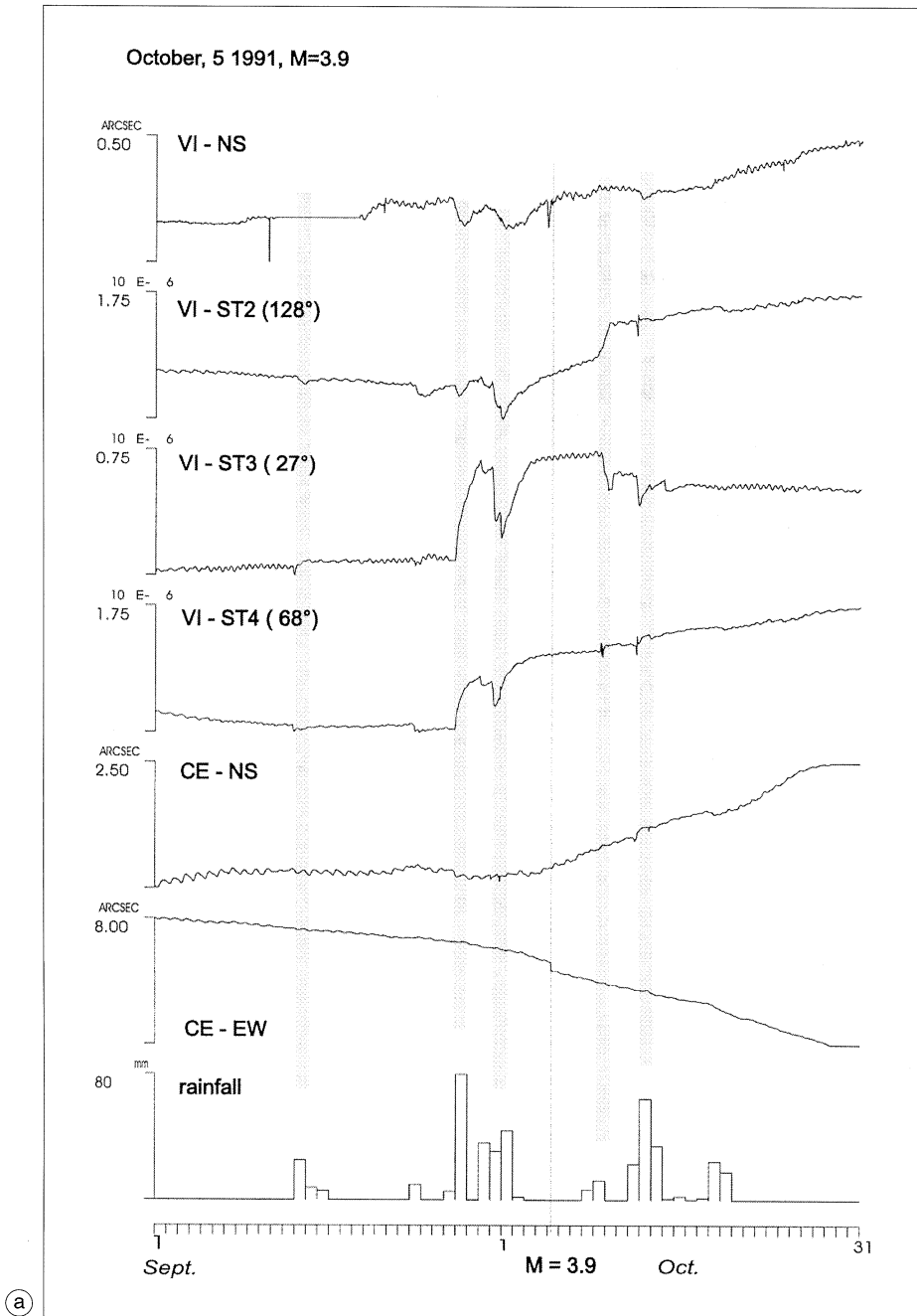
Fig. 10 a,b. Tilt (stations VI, CE, IN, GE, BA) and strain (station VI) records at the time of the October 5, 1991, *M* = 3.9 event. Also shown is the watertable

October, 5 1991, M=3.9
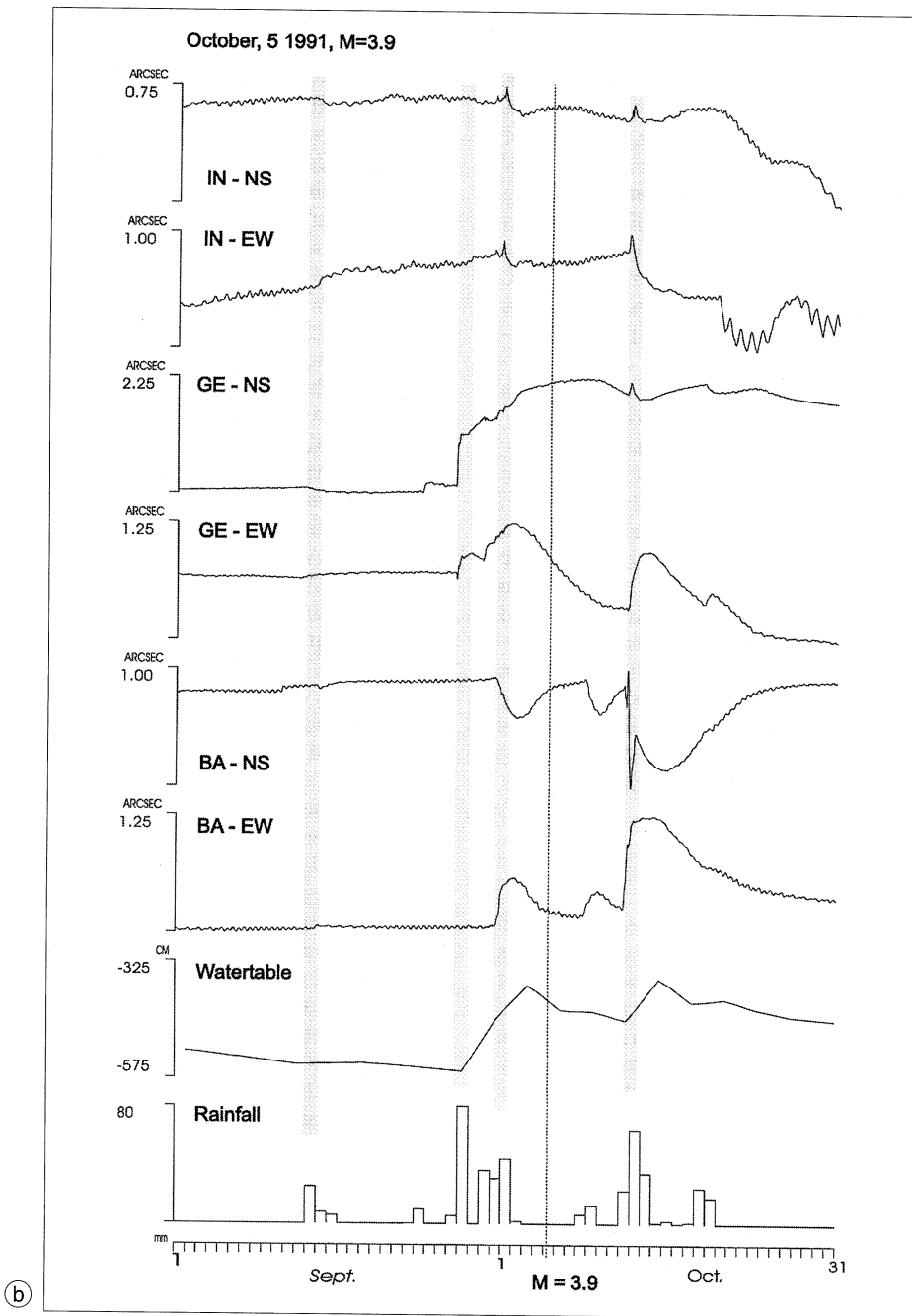
variation and rainfall (see fig. 7). The days of greater rainfall are identified as grey bars.

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of all stations of the network are graphed in figs. 10a,b together with rainfall and water table variation (Venzone). Hypocentral distances at VI, CE, IN, GE, BA are 11, 25, 33, 15 and 56 km, respectively. Grey bars mark the days of greatest rainfall. A particularity, relevant to strain/tilt measurements was an extraordinarily strong precipitation at and near VI preceding the event by 5 days. The strainmeters ST3 and ST4 gave way to an extensional deformational signal, contrary to the normal rain response, which is generally compressional, for all three components (see fig. 8). This last signal can be modeled in general terms using a statistical approach (*e.g.*, Braitenberg, 1999) or by poroelastic theory (*e.g.*, Weise *et al.*, 1999). At the same time, the tiltmeters in GE recorded a dislocation. One tilt component (NS) of VI recorded a short-term signal during the 8 h before the event. The coseismic deformation was only recorded in the CE-station. The coseismic signal could not be modeled, as no reference regarding the focal mechanism was found. Using general considerations regarding the local seismotectonics the compatibility of the observed signal with the dislocation model was tested and the anomalous deformation signal modeled by different rheological models in Dal Moro and Zadro (1999).

f) *12 April 1998, M = 5.2*

This event is the strongest local shock, which occurred after the seismic crisis of the years 1976-1977. The only operating Friuli station at the time of the event was the VI station (30 km hypocentral distance), and the records are shown in fig. 11. A distinct coseismic permanent deformation is observable on all tilt and strain components. The amplification factors of the strain records have been adjusted to the theoretical earth tides, as an absolute calibration of the instruments has currently not been made. The rapid compressive deformations with slower recovery recorded on all three strainmeters on the days 7, 12, 16, and 30 of April are due to rainfall (compare with fig. 8). The consequent trend leads to an uncertainty in the observed coseismic deformation, which is though limited to about 10% of the coseismic signal. At present

only the preliminary hypocentral locations of the main shock and aftershocks have been released by the OGS network. The fault mechanism from first arrivals has been determined to be of Dinaric type (OGS), with a dextral strike slip mechanism on a nearly vertical NW-SE striking fault (azimuth 38°W from N). From the relations outlined above (Hanks and Kanamori, 1979), the seismic moment is estimated to be $M_0 = 6.3 \ 10^{17}$ Nm, and the fault length 6.8 km. Assuming a square fault, the dislocation then amounts to 0.77 m. This seismic moment is in accordance with that determined seismologically which is given between $1.4 \ 10^{17}$ Nm and $1.1 \ 10^{18}$ Nm (NEIC).

Also in this case we tested the spatial distribution of the aftershocks in order to obtain further information on the possible orientation of the dislocation, which could explain the measurements. As described above, the ellipsoid, which fits N successive aftershocks, is determined at consecutive overlapping time windows shifted by an amount of K events. According to the method, the orientation of the shortest principal axis is orthogonal to the active focal plane during the aftershock sequence. In the analysis a window of 50 events, shifted by 25 events was used. In fig. 12a the orientations of the principal axes are shown in lower hemisphere stereographic projection. The orientation of the principal axes is stable throughout the sequence, the vector with the smallest eigenvalue being horizontal with azimuth N30E-N150W. The eigenvectors with the largest and intermediate eigenvalues vary in orientation, but remain in a nearly vertical plane. It can be inferred that the aftershock sequence is confined to a subvertical plane with Dinaric (NW-SE) orientation. The lengths (in km) of the principal axes are given in fig. 12b for the successive windows of analysis. The lengths of the largest and intermediate principal axes estimate the spatial extension of the aftershocks on the active fault. They can be therefore used to give some inference on the size of the fault. In this case the aftershocks which occur in the first window of analysis (first 50 events) are distributed on an area of about 10 by 6 km. Figure 12b shows that the sizes of the ellipsoids calculated on successive windows of analysis decrease. The fault size estimated from the af-
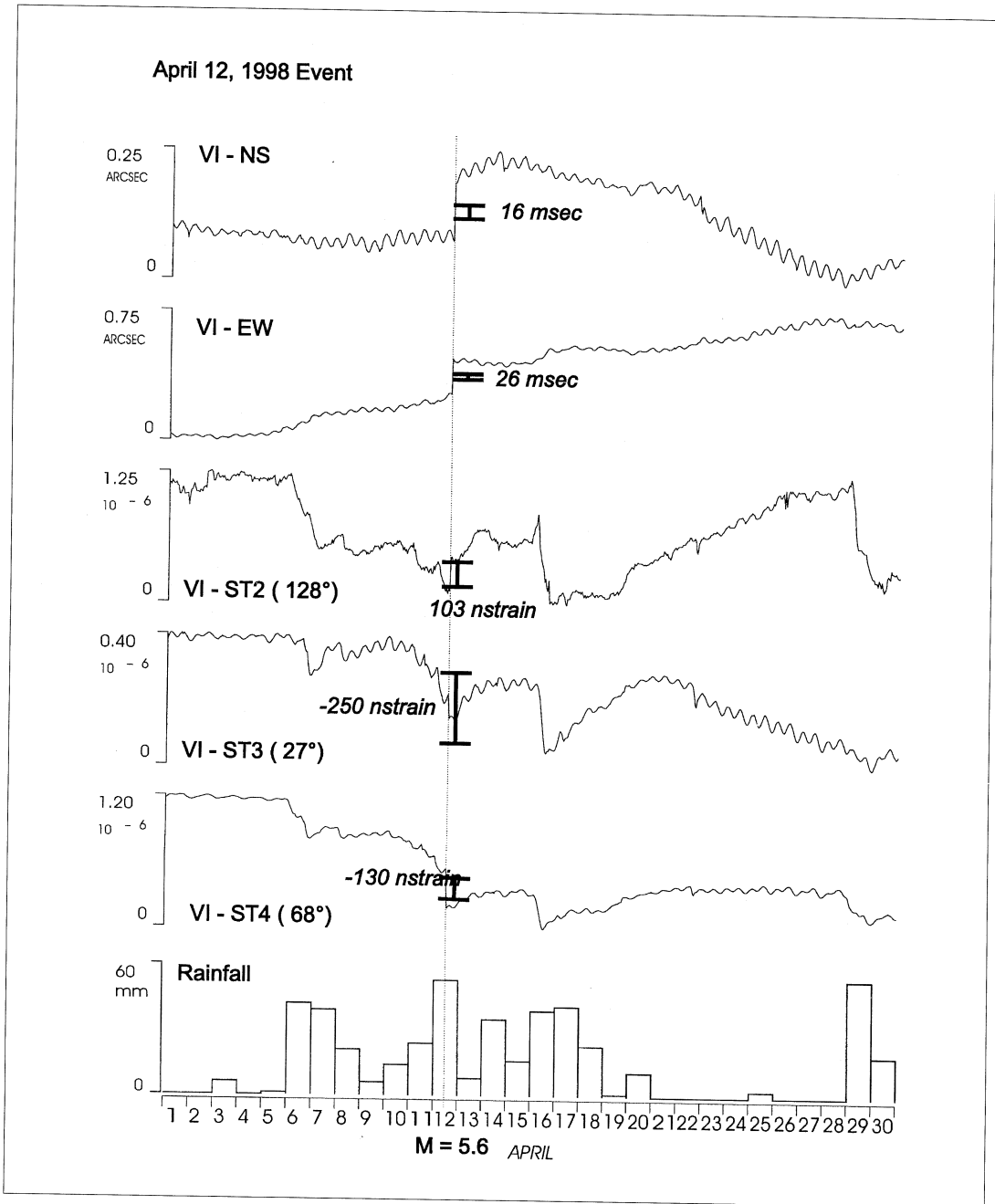
Fig. 11. Tilt and strain records of the April 12, 1998, $M = 5.6$ event. The event is the largest local event since installation of the Friuli tilt-strain network. A clear step-like signal is observed for both tilt- and all three strainmeters. The modeled signals are shown as vertical bars.

April 12, 1998 event
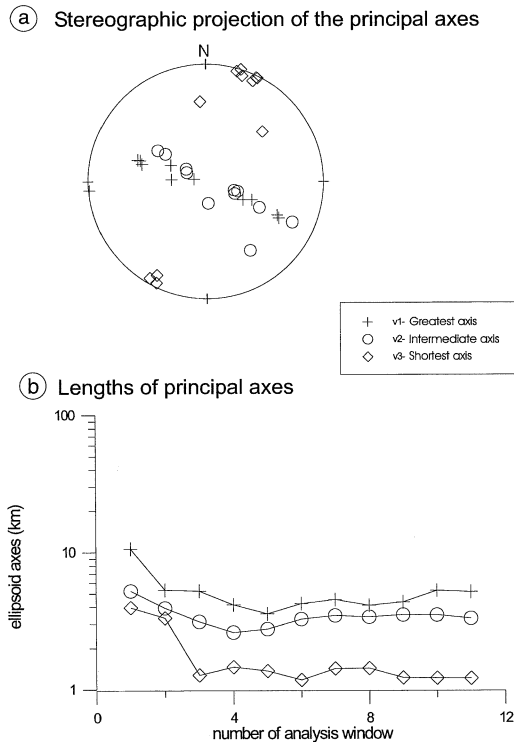Aftershock distribution
N=50, d=25

(a) Stereographic projection of the principal axes

(b) Lengths of principal axes

Fig. 12a,b. Analysis of the aftershock sequence of the April 12, 1998, $M = 5.6$ event by calculating the ellipsoidal distribution of n ($n = 50$) successive aftershocks on windows of analysis shifted by $k = 25$ events. a) Orientations of the three principal ellipsoid axes in lower hemisphere stereographic projection. b) Lengths of the principal axes for each window of analysis (in km).

tershock distribution is comparable to that obtained from the theoretical scaling relations.

We use the Dinaric fault model as a starting point for modeling the strain and tilt observations by a dislocation model. For a fixed range of azimuth and dip angles, the squared error between modeled and observed strain and tilt data is minimized, by allowing the rake to vary in certain bounds. Azimuth, dip and rake are allowed to vary by ± 45°, starting from azimuth

N30W, dip 90° and rake 180°. This range comfortably covers the mean uncertainty in the fault plane solutions. In fig. 13 the root mean square (rms) error is shown for different azimuth and dip. Couples of azimuth and dip, for which the signs of the modeled tilt and strain values are wrong, are blanked. A broad zone of extremal values is centered at an azimuth of N40W and a dip of 65°, leaving an uncertainty range of about 20° in azimuth and 30° in dip. The rms-error is smallest when the dislocation is allowed to have a thrust component. This solution is in agreement with the preliminary fault-plane solutions for what concerns dip and azimuth; the thrust component of the slip vector is not confirmed by the seismological observations (P. Suhadolc, personal communication). In the future the modeling shall be repeated constraining the solution with the tilt-strain observations and the to be published fault-plane solutions obtained from seismological observations. It shall be also considered in which extent local conditions at the observation site can influence the observed deformations. The corresponding deformation field is given in fig. 14a-e, where a dislocation model defined by the parameters given in table VIII, event 8 was used. The fig. 14a-e pertain to the fields of the two tilt components (NS and EW, in ms) and the three strainmeters (ST2, ST3, ST4, in nstrain) respectively. The x-axis is oriented along the strike. The black dot shows the position of the Villanova station, the origin of the coordinate system being centered on the epicenter.

7. Conclusions

The strain-tiltmeter network of Friuli (installed in 1977) and the horizontal pendulums station of Trieste (installed in 1959) have by now a long history of recordings. Some instrumental revisions and modernizations made a summary and description of the existing instrumentation necessary.

The overall spectral properties of the strain and tilt records were calculated and revealed the power law decay of spectral energy with frequency, the coefficient of which seems stable in different strain measurements made throughout
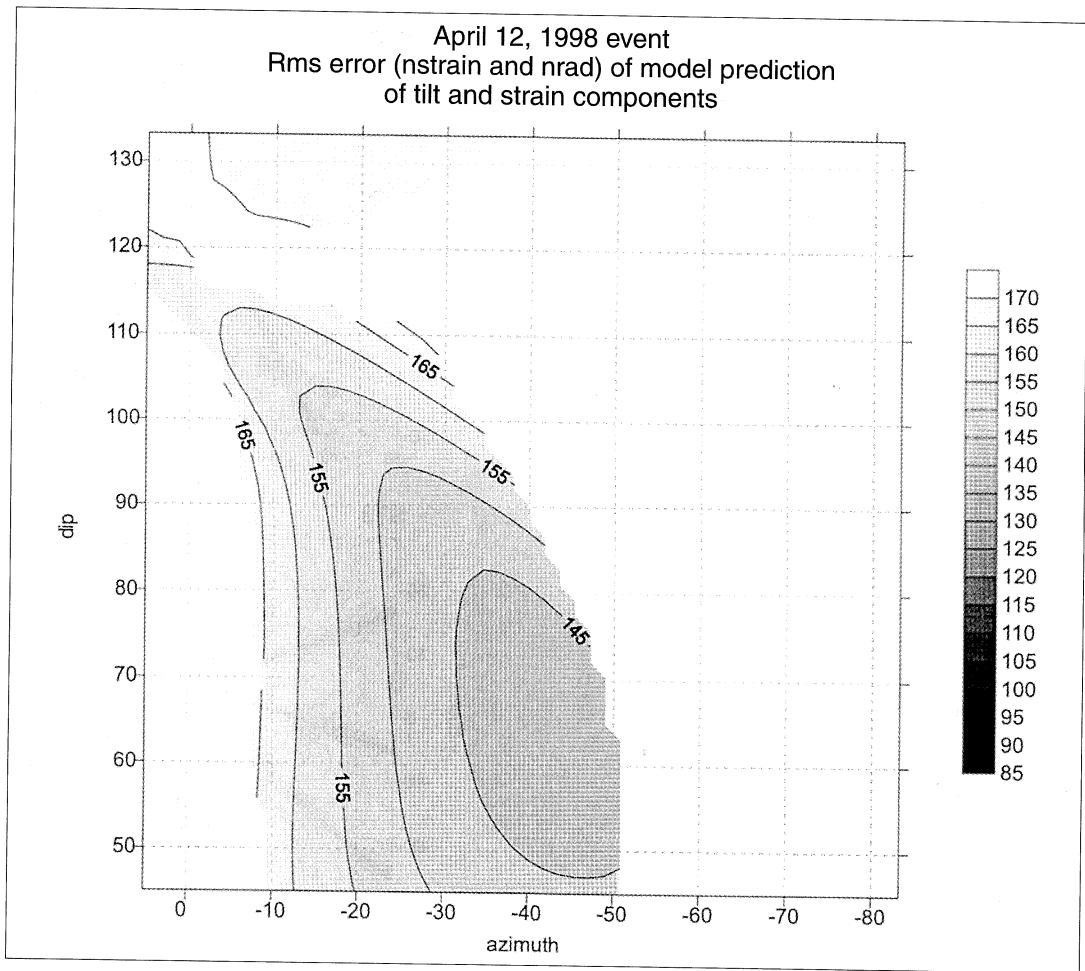
Fig. 13. Root mean square prediction error of observed tilt and strain at VI, for different couples of azimuth and dip of the fault model. The rake is confined to the range of 180 ± 45°. Grey code of prediction error on the right. Azimuth positive from N to E.

the world. The evaluation of the mean spectrum is relevant for estimating the noise level of the measurements, important in connection with the search for pre-, co- or postseismic strain signals. Due to their exceptionally large building, the Grotta Gigante horizontal pendulums have been shown to have extremely low noise levels, comparable to that of the long-base tiltmeters.

The local seismicity was studied with respect to the magnitude-distance relation to all stations of the network. For the events satisfy-ing the most favorable magnitude-distance relation, the short term strain tilt observations were shown. Most of the events selected by this criterion were accompanied by pre-, co-, or post-seismic deformation. In some cases it was possible to model the coseismic steps by a disloca-tion model. The simplification inherent in the model (homogeneous halfspace, single disloca-tion) may be the cause for discrepancies in the amplitude between the observed and modeled signals. Problems of this sort had also arisen in
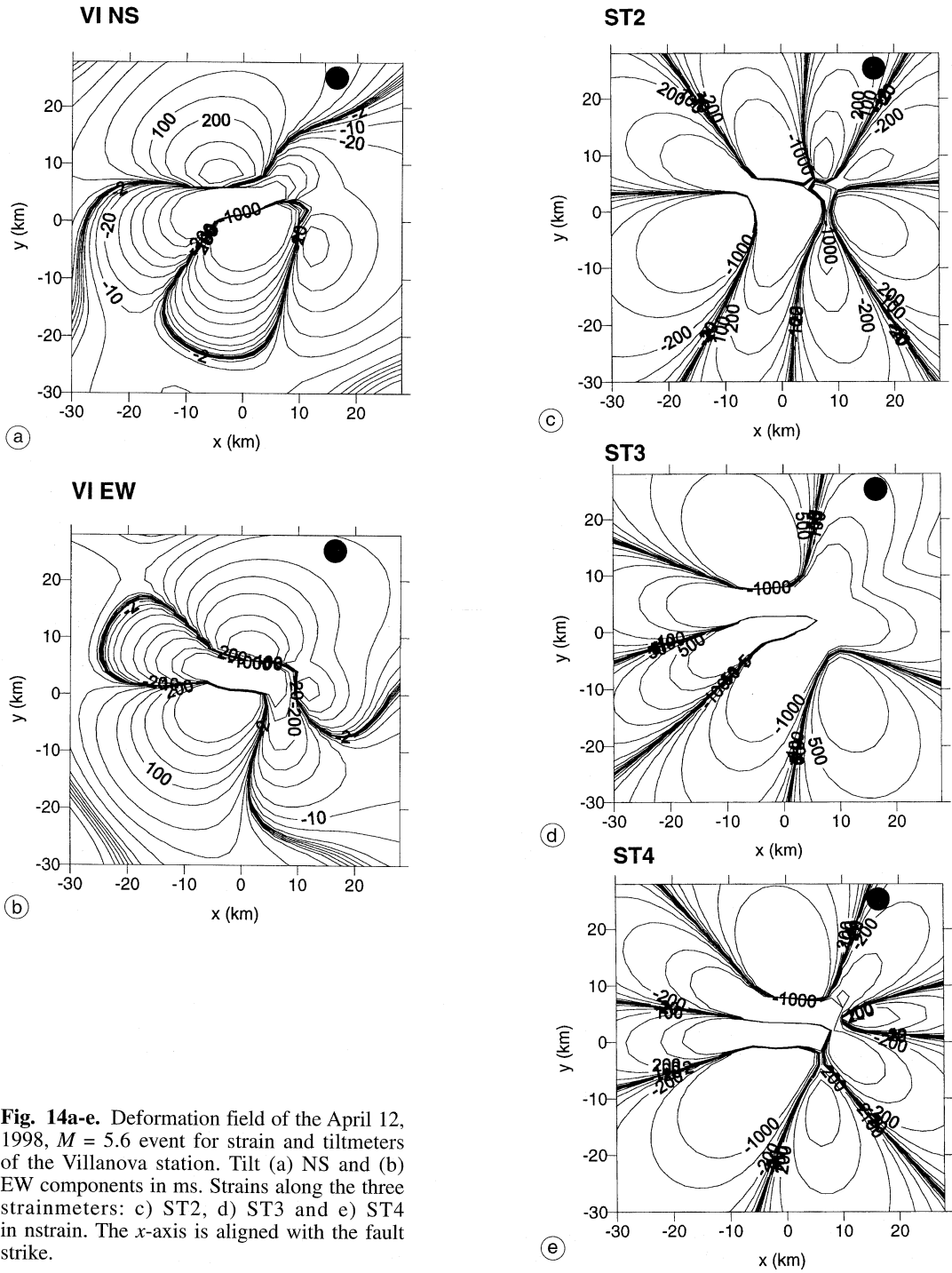
VI NS

ST2

VI EW

ST3

ST4

Fig. 14a-e. Deformation field of the April 12, 1998, *M* = 5.6 event for strain and tiltmeters of the Villanova station. Tilt (a) NS and (b) EW components in ms. Strains along the three strainmeters: c) ST2, d) ST3 and e) ST4 in nstrain. The *x*-axis is aligned with the fault strike.

works which attempted to explain observations in California (Johnston *et al.*, 1987; Wyatt, 1988; Linde and Johnston, 1989)

We may conclude that the magnitude-distance relation, which predicts that a strain signal associated with a seismic event, be it pre-, co- or postseismic, decays outside the focal volume with the inverse of the third power of the hypocentral distance, is a good means to select events for which strain signals should be expected. We have though encountered some cases, for which a strain signal was observed although the event was distant in relation to its magnitude. Several causes could explain this fact: inhomogeneity of the medium, or underestimation of the magnitude by the seismological record.

Acknowledgements

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