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Space geodetic data (GPS) and earthquake forecasting: examples from the Italian geodetic network

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ABSTRACT Some considerations are given about attempts at predicting impending earthquakes by the use of space geodetic (GPS) observations in the Italian peninsula. Both short (weeks to days) and middle/long-term (months to years) presumed precursors are considered. At present, none of the numerous published studies seems to identify significant and systematic earthquake precursors. At most, geodetic data may be used to quantify the rate of tectonic deformation and its lateral variations, in order to identify the zones where strain accumulation is fastest. However, the connection between the level of strain and the probability of earthquakes is not clear yet. We argue that the most promising way to use geodetic data is monitoring post-seismic relaxation, that is the perturbation of the strain field triggered by a major shock. The examples recognized so far of interaction between seismic sources suggest that the highest probability of induced earthquakes at a mature fault corresponds to the arrival time of the highest strain rate values. The numerical simulation of post-seismic relaxation indicates that the expected variation of velocity, strain and strain rate in the zones surrounding the triggering shock may be significantly higher than the resolution of GPS observations. Thus, one may expect that geodetic monitoring, integrated by a deep knowledge of the ongoing tectonic setting, may allow the identification of the zones where the probability of an induced earthquake is undergoing a significant increase.

Key words: earthquake forecasting, GPS observations, Italy.

1. Introduction

In the last decade, satellite-based geodetic techniques such as Thermal Infrared survey (TIR), Synthetic Aperture Radar (SAR) and Global Positioning System (GPS) have considerably developed and their applications to seismotectonic investigation are now manifold (e.g., Tronin, 2010). In particular, due to a combination of relatively low cost, handiness and significantly improved accuracy of the ground-based receivers, dense networks of GPS stations, often installed in permanent sites, are now available [see e.g., Seeber, (2003) for a description of the Global Positioning System technique].

The analysis of this kind of observations, carried out in zones just hit by major shocks, has allowed the definition of pre-seismic, co-seismic and post-seismic kinematic patterns (e.g., Pollitz *et al.*, 2006; Anzidei *et al.*, 2009; Cheloni *et al.*, 2010; Delouis *et al.*, 2010; Ozawa *et* *al.*, 2011; Cenni *et al.*, 2012), which may help to constrain the geometry and kinematics of the activated faults and the redistribution of stress and strain after a seismic event (e.g., Decriem and Arnadottir, 2012). This last information can then be used to gain insights into the possible role of post-seismic perturbation in triggering further seismicity (e.g., Viti *et al.*, 2003, 2012, 2013; Lou and Liu, 2010).

This work first reviews the attempts made so far at exploiting the above results for short and long-term prediction of earthquakes and then discusses the possibility of recognizing, by the help of geodetic monitoring, where and when the probability of induced earthquakes is highest in the zones surrounding a major seismic event.

2. GPS observations and earthquake prediction: selected examples

2.1 Short-term precursors

GPS receivers operate in a frequency range that make them more suitable to continuously monitor the ionospheric total electron content (TEC), with respect to conventional ionosondes (e.g., Liu *et al.*, 1996). Thus, GPS measurement can be used to reconstruct time series of TEC over seismic areas and to detect eventual sudden deviations from the mean trend, which are suspected to be linked with seismic activity. For instance, vertical ground displacements, induced by co-seismic elastic rebound and Rayleigh surface waves, are transmitted and strongly amplified in the atmosphere, due to the exponential decrease of air density with height. These perturbations may cause considerable post-seismic TEC disturbances (e.g., Artru *et al.*, 2001; Liu *et al.*, 2006).

More controversial is the presumed relationship between ionospheric anomalies and the physical processes that precede strong earthquakes. Some authors (e.g., Kim *et al.*, 1994; Pulinets and Boyarchuk, 2004) have suggested that TEC anomalies may be induced in the ionosphere by the transmission of electromagnetic fields and/or transport of ions produced by radon escaping from seismogenic zones. The studies of this phenomenon (e.g., Pulinets, 2006) suggest some considerations about the connection between observed anomalies and preparatory earthquake processes:

1) only earthquakes with magnitude $M \geq 5.0$ may induce detectable ionospheric disturbances;

2) the induced anomalies usually appear few days (1 to 5 on the average) before the mainshock;

3) the size of the ionospheric zone affected by anomalies increases with magnitude.

Relying on the above hypotheses, several TEC anomalies, detected by GPS networks, have been interpreted as seismic precursors (e.g., Plotkin, 2003; Liu *et al.*, 2004b, 2009; Tsai *et al.*, 2006; Singh *et al.*, 2009; Jhuang *et al.*, 2010; De Agostino and Piras, 2011; Hasbi *et al.*, 2011; Kon *et al.*, 2011; Contadakis *et al.*, 2012; Ma and Wu, 2012 and references therein). However, the above approach has not received unanimous consensus. The criticism so far advanced concerns some major aspects:

- in some cases, as for the 2008 Wenchuan (China) earthquake, the spatial extent of the observed ionospheric anomalies is much larger than, or scarcely related to, the zone of earthquake preparation expected by theoretical models (e.g., Dobrovolsky *et al.*, 1979);
- some studies (e.g., Dautermann *et al.*, 2007) point out that TEC anomalies and seismic

events are poorly correlated, both in time and space. For instance, it is not clear if and how TEC anomalies occur in aseismic areas also (e.g., Rishbeth, 2007);
- some processes, as the variability in geomagnetic activity, may explain the detected TEC perturbations better than seismogenic phenomena, whose connection with the distant ionosphere cannot easily be understood (e.g., Rishbeth, 2006);
- all the attempts so far made at applying this approach refer to retrospective predictions, as often happen with earthquake precursors (e.g., Mulargia and Geller, 2003);
- in spite of the fact that many GPS stations operate worldwide since the end of the last century, no one of the recent major earthquakes (e.g., 2004 Sumatra M=9.0, 2008 China M=7.9, 2010 Haiti M=7.0, 2010 Chile M=8.8 and 2011 Japan M=9.1) has been predicted by GPS-TEC analysis or other precursors. So, the above approach still needs much more validation, for instance by the use of contingency tables (i.e., counting of successful and failed predictions) as suggested by Rishbeth (2006).

Short-term forecasting has often involved detection of pre-seismic surface deformation, such as uplift and subsidence, ground tilting and changes in strain rate (e.g., Rikitake, 1976; Lomnitz, 1994; Kanamori, 2003). The fact that this kind of approach has not yet provided significant and systematic results may be due to the difficulty in detecting the very small ground displacements that generally precede a seismic event. In this regard, one must consider that before the advent of satellite-based geodesy, these effects had to be monitored by expensive and time-consuming techniques, such as laser-ranging, repeated surface levelling and observations by tiltmeters and strainmeters (e.g., Mogi, 1985; Zadro and Braitenberg, 1999; Rikitake and Hamada, 2001). A compilation of the results so far obtained by this kind of investigation [(Table 6) in Cicerone *et al.*, 2009], shows that in a half century (1944-1993) only 20 possible ground anomalies, related to 12 earthquakes, have been tentatively detected. The observed vertical movements took place months to days before the main shock, within about 100 km from the epicentre, and were characterized by amplitudes increasing with earthquake magnitude. However, one might remark that all the above analyses are retrospective. Moreover, it must be pointed out that ground deformation, such as subsidence, fissuration and also true faulting, may be caused by non-seismogenic processes, as groundwater withdrawal (e.g., Jachens and Holzer, 1979). Overlooking this possibility may bias the interpretation of ground anomalies and limit their recognition as seismic precursors.

At present, the observational techniques are considerably improved, since high-frequency GPS measurements can promptly detect even minor changes in the relative position of a station site (e.g., Ge *et al.*, 2000). In particular, space geodesy, also including SAR interferometry, may reveal local variations in the rate of horizontal and vertical surface strain, as indicated by analyses of time series acquired before recent strong shocks (e.g., Yu *et al.*, 2001; Jade *et al.*, 2003; Liu *et al.*, 2004a, Gu *et al.*, 2011). The recent work by Chen *et al.* (2013) on the 2010 Jashian (Taiwan) M=6.4 earthquake shows a clear reorganization of the azimuths of horizontal GPS velocity vectors around the epicentral zone in the week before the main shock. Again, Kamiyama *et al.* (2012) claim a noticeable increase in the slope of averaged GPS time series shortly before the great March 11, 2011 Japan earthquake. In particular, the above authors point out that for a number of GPS stations located along the north-eastern coast of Hokkaido, the horizontal displacement vectors started pointing towards the future epicentral region since 3 days before the main shock.

Although interesting, these retrospective analyses have not pointed out unambiguous and systematic seismic precursors yet. In particular, it is not clear whether short-term variations in the kinematic field may happen in absence of significant seismic activity. On the other hand, it would be opportune to check if all recent major earthquakes have been preceded by such kind of anomalies. As said earlier, a systematic analysis based on contingency tables (Rishbeth, 2006) would be very useful. Insights into the reliability of this approach could be gained by comparing pre-seismic and co-seismic kinematic fields, determined by GPS and SAR measurements, in order to check the compatibility of the precursory signals with the possibly related seismic source.

A relatively recent research field is the study of slow earthquakes and related phenomena, often indicated as episodic non-volcanic tremor, low and very-low frequency earthquakes, slow-slip events, creep events or silent earthquakes (e.g., Schwartz and Rokosky, 2007; Beroza and Ide, 2011; Gao *et al.*, 2012; Vidale and Houston, 2012). A regular seismic cycle can be described by three distinct phases: a prolonged, steady-state accumulation of elastic strain at the seismogenic fault, a sudden co-seismic slip with significant seismic energy release, and a relatively slow post-seismic perturbation. Although slow earthquakes are compatible with shear faulting, their duration is generally much longer than the one of regular seismic events of comparable seismic moment (e.g., Ide *et al.*, 2007). In particular, co-seismic slip usually completes in seconds or minutes, while slow events may last for tens or even hundred of days (see e.g., Table 1 by Schwartz and Rokosky, 2007). This implies that slow earthquakes generate much less seismic radiation than regular shocks (e.g., Ide *et al.*, 2007).

Detecting slow earthquakes may be hampered by technical constraints, as bandwidth limitations and saturation, that affect strainmeters, tiltmeters, tidal recorders and seismometers. However, GPS devices operating at fast sampling rate seem to be particularly suitable for recording slow-slip events (e.g., Ge *et al.*, 2000). Indeed, several slow earthquakes have been recognized since the installation of dense networks of continuous geodetic stations (CGPS; Schwartz and Rokosky, 2007).

Most of slow-slip events and periodic tremors have been recognized along circum-Pacific subduction zones as Cascadia, Cocos and Japan (e.g., Schwartz and Rokosky, 2007; Holtkamp and Bruzinski, 2010; Jiang *et al.*, 2012; Hirose *et al.*, 2012). Their sources are located at depth of 20-40 km, i.e., in the brittle/frictional portion of the subduction fault, but well below the hypocentral zone of large subduction earthquakes [around 15 km, e.g., Schwartz and Rokosky (2007) and references therein]. The physical processes underlying slow earthquakes are still debated. Several explanations have been proposed involving frictional properties and pore fluid pressure (e.g., Amoruso *et al.*, 2004; Bilek *et al.*, 2004; Lowry, 2006; Audet *et al.*, 2009; Kaproth and Marone, 2013; Audet and Burgmann, 2014). For instance, Gao *et al.* (2012) suggest that slow earthquakes occur in sectors of the subduction fault where small asperities with high shear strength are distributed within a low-strength zone, possibly related to near-lithostatic pore pressure. So, slow fault sliding at depth could enhance stress in the still locked, seismogenic shallow subduction fault, eventually causing its breaking.

In this view, some authors have suggested that slow earthquakes may represent a seismic precursor (e.g., Bouchon *et al.*, 2011, 2013; Kato *et al.*, 2012). In particular, the analysis of CGPS time series might reveal slow-slip events that precede large regular earthquakes by months to hours (e.g., Jiang *et al.*, 2012; Protti *et al.,* 2014). However, these claims are based on retrospective analysis of seismological and geodetic data referred to a few case histories; no true prediction has been proposed so far, also due to the relatively recent development of CGPS networks. Furthermore, an unequivocal relationship between slow and regular earthquakes has not yet been recognized; silent events have been detected in both interseismic and postseismic phases of the seismic cycle (e.g., Schwartz and Rokosky, 2007). Finally, several large earthquakes ($6.0 \leq M \leq 8.4$) at various tectonic boundaries were not preceded by silent seismic phenomena (Roeloffs, 2006). Freymueller *et al.* (2008) also note the scarcity of slow-slip events detected at the Alaskan-Aleutian subduction zone in the 1995-2007 time interval. This behaviour strongly differs from that of other subduction boundaries, such as the Ryukyu Islands (southern Japan), where slow earthquakes and tremors seem to occur with biannual periodicity.

Due to ambiguities in the observed phenomenology and the proposed physical mechanisms, continuous geodetic monitoring could take several decades to assess the reliability of slow earthquakes as seismic precursor (Vidale and Houston, 2012).

For the Italian region, the research on slow earthquakes mainly concerns the central Apennines, where suitable instrumental devices have been operating for over a decade (Crescentini *et al.*, 1999; Amoruso *et al.*, 2002; Scarpa *et al.*, 2008). Although both slow-slip events and low-frequency tremors have been observed in that zone, such phenomena seem to be quite rare (Scarpa *et al.*, 2008). Moreover, the same authors cast doubt on the possibility that all slow-slip events may be detected by GPS sensors, since that the related ground displacement is often negligible. Indeed, slow fault slip following the April 6, 2009, L'Aquila earthquake (M=6.3) has been detected by very high sensitivity laser strainmeters, located in a deep tunnel about 20 km from the epicentre (Amoruso and Crescentini, 2009). On the other hand, no significant signal seems to have preceded such shock (Amoruso and Crescentini, 2012).

In summary, the relevance of slow earthquakes as seismic precursor is still debated, particularly in seismotectonic contexts not related with subduction zones.

Recognizing anomalies related to eventual pre-seismic ground displacements is difficult also because the underlying physical mechanism is not known. In the prediction of volcanic eruptions, one can rely on the fact that ascent and emplacement of magma implies opening of tensile fractures and crustal inflation, that in turn induce detectable surface uplift, tilting of volcano flanks and volcanic tremors (e.g., Mogi, 1958; Sparks, 2003).

In case of tectonic earthquakes, it is often assumed that stress increase causes rock microfracturing, which results in anelastic, volumetric increase (dilatancy) of the crustal volume centred around the future hypocentre (e.g., Dobrovolsky *et al.*, 1979). This volume of cracked rock at depth (preparation zone) would be the source of precursor signals, that must propagate in the surrounding crust in order to be observed at surface around the epicentre (precursor area). This model implies that short-term precursors of strong shocks would be observed over a large area, increasing with earthquake magnitude. For instance, the expected radius of the precursor area is 380 km for magnitude M=5.0, and 1023 km for M=7.0 (Dobrovolsky *et al.*, 1979). In our opinion, the above model may be oversimplified, as it assumes that the crust surrounding the preparation zone is perfectly homogeneous, so neglecting the possible lateral anisotropy of real tectonic contexts (Sgrigna and Conti, 2012). This may explain partly why significant pre-seismic strain signals are often lacking, as happened for the 2004 Parkfield, California and the 2009 L'Aquila shocks (Bakun *et al.*, 2005; Amoruso and Crescentini, 2012).

One should take into account as well that the nature of earthquake nucleation is still largely unknown, as underlined by the debate about the role of rock fracture and fault friction (including modern rate-and-state laws) in controlling seismogenesis (e.g., Marone, 1998; Reches, 1999; Lockner and Beeler, 2002; Beeler, 2004; Mulargia *et al.*, 2004). In this respect, it should be noted that dilatant behavior and accelerating strain before rupture have been clearly documented in laboratory experiments, but not verified at the crustal scale yet (e.g., Mulargia and Geller, 2003; Tiampo and Shcherbakov, 2012). Also, in contrast with Dobrovolsky *et al.* (1979) model, the fracture-driven (or friction-controlled) nucleation region might involve only a very limited volume around the fault surface, so reducing the possibility of detecting precursory signals at the surface (e.g., Amoruso and Crescentini, 2012).

2.2 Long-term forecasting

Attempts at using geodetic data for recognizing the zones most prone to next earthquakes mainly involve the estimate of the ongoing strain rate field in a given zone. This information is then used to tentatively evaluate the time span required to accumulate the elastic deformation necessary for a strong earthquake, which is compared with the average recurrence interval inferred from seismic history (e.g., Bilham *et al.*, 1997; Liu *et al.*, 2000, 2007; Friedrich *et al.*, 2003). Similar attempts have been performed by Jenny *et al.* (2006), Serpelloni *et al.* (2010) and Angelica *et al.* (2013) for the Italian region. However, this procedure can hardly provide significant information when the amount of strain stored in individual faults before the starting of GPS measurements is not known. Since the information about fault surface properties, pore fluid pressure and stress amplitude is usually poor, we do not know which faults are prone to seismic activation.

For instance, Ergintav *et al.* (2014) use the GPS velocity field to compute the rate of elastic strain in the Marmara Sea fault systems (western Turkey). By estimating the slip accumulation at each fault segment since the last strong earthquake, they try to recognize the fractures that could slip first. However, no precise temporal constraints about the occurrence of the next large shock is provided. Moreover, such kind of approach needs a detailed knowledge of the seismotectonic setting, reliable historical information (i.e., a long and complete seismic catalogue), and accurate localization of epicentres, in order to assign each known shock to a specific fault segment.

Other investigations have compared the pre-seismic and co-seismic GPS velocity field to recognize eventual seismic gaps in plate boundaries [e.g., Ruegg *et al.*, (2009); Moreno *et al.*, (2010) and Metois *et al.*, (2012, 2013, 2014) for the Chilean subduction zone]. However, this approach may encounter important difficulties, as discussed in a number of papers (e.g., Kagan and Jackson, 1991, 1995). A debated issue is the possibility of simultaneous rupture of several segments of a given tectonic boundary. In Japan, where a dense GPS network with hundreds of permanent stations operate since 1994 (e.g., Sagiya, 2004), many efforts have been directed towards identification of zones where the crust is deformed at largest rates (e.g., Suwa *et al.*, 2006; Loveless and Meade, 2010). In particular, along the Kurile-Japan subduction boundary, Hashimoto *et al.* (2009) have recognized at least six "slip-deficit" regions, where the seismic slip rate (deduced by the pattern of historical seismic activity) is largely insufficient to accommodate observed geodetic velocities. Thus, the above authors envisaged that future strong shocks might occur at those zones of enhanced strain rate. Unfortunately, in that work no explicit conclusion was drawn about the possibility of a megathrust earthquake along the above subduction zone, as actually occurred just three years later (2011, Tohoku, Japan, M=9.1). In this case, GPS data have not significantly influenced the assessment of seismic hazard for eastern Japan, which was arbitrarily based on the assumption that adjacent segments of the Japan Trench subduction fault would never activate simultaneously (e.g., Stein *et al.*, 2012).

Recently, several papers have addressed the problem of how deformation accumulates at tectonic zones, in the framework of the seismic cycle theory developed earlier (e.g., Savage, 1983; Thatcher, 1984, 1993 and references therein). Some authors (e.g., Fialko, 2006; Shen *et al.*, 2007) assume that the higher is the rate of observed geodetic strain, the more frequent or the larger the future earthquakes must be. Moreover, elastic modelling of inter-seismic deformation (e.g., Stein and Wysession, 2003) predicts that the surficial shear strain rate is maximum near the fault trace and rapidly diminishes outwards. In this view, detecting the zones where strain accumulation is fastest would be useful to identify the seismic sources most prone to next shocks. However, we do not know how many potentially seismogenic fractures are near to (or far from) failure. A fault whose actual tectonic load is low (or whose shear strength is high, or both) may need a long time to generate the next strong shock, even if the geodetically detected strain rate is large, and vice versa. Moreover, the relatively short time span of available GPS data sets (10-20 years at most) casts doubts on the hypothesis that such information may provide a realistic representation of the pre-seismic loading rate at tectonic boundaries. Furthermore, the geodetic strain is the result of both elastic and inelastic crustal deformation, whose respective contributions are usually poorly known. Thus, the hypothesis that a local maximum of geodetic strain rate corresponds to a fast accumulation of elastic strain and stress may be incorrect, unless one can exclude that significant inelastic (i.e., aseismic) strain takes place.

A point of view opposite to that mentioned above is suggested by Riguzzi *et al.* (2012, 2013), who claim that a fault close to the end of its seismic cycle presents the lowest shallow strain rate. This interpretation is based on the elastic and viscoelastic modelling of fault zones (Thatcher, 1983; Doglioni *et al.*, 2011) and on the statistical correlation recognized between the spatial distribution of recent seismicity in the Italian region and the GPS-derived strain rate field (Riguzzi *et al.*, 2012). However, some remarks should be made about the reliability of the above approach, as discussed in the following:

- Doglioni *et al.* (2011) adopted a clearly oversimplified modelling approach, involving a purely elastic behavior. A more realistic simulation of the inter-seismic fault loading has been carried out by Thatcher (1983), which considers the coupling between the shallow elastic-brittle lithosphere and the underlying viscoelastic layer. In this case, the steady motion of the viscoelastic layer (driven by plate motion) causes the deformation of the upper elastic layer at a rate that decreases with time and is minimum at the end of the seismic cycle. However, Thatcher's (1983) modelling only considers a single, vertical strike-slip fault. Such as an isolate feature does not interact with the surrounding structures that would exist in a more realistic context;
- stress transfer by fault interaction following strong earthquakes (e.g., Harris, 1998; Freed, 2005; Steacy *et al.*, 2005; Lou and Liu, 2010) may drastically change the loading pattern envisaged by Thatcher (1983). Examples of interaction among seismic sources located

along the Apennine chain, due to post-seismic relation processes, are discussed in Viti *et al.* (2012, 2013). The related numerical experiments show that the strain rate near a seismogenic fault may suddenly change, as an effect of the perturbation induced by earthquake previously occurred in surrounding structures;

- modelling of buried fractures (e.g., Stein and Wysession, 2003) shows that in the pre-seismic phase the surface strain rate decreases with the fault burial depth. So, a low strain rate measured at the surface (as provided by geodetic measurements) does not necessarily imply slow strain accumulation near the blind fault;
- the geodetic evidence cited in support of the above view (Riguzzi *et al.*, 2012, 2013) should be subjected to rigorous checks. Above all, the accuracy and resolution of the strain rate field that can be determined by the GPS network available in the central Mediterranean play a key role;
- the GPS network considered by Devoti *et al.* (2011) presents a number of zones poorly covered by stations, for instance along the eastern coast of the Adriatic Sea. Furthermore, the geographical configuration of the monitored area (including vast marine zones, as the Tyrrhenian and Adriatic seas) could bias the evaluation of strain rate in the peripheral zones of the study area. In particular, one could note that, in several zones hit by earthquakes in the considered period, the distribution of GPS sites can hardly guarantee an accurate estimate of the strain rate (in particular, the Dinaric area, where almost half of the reported earthquakes have occurred). This problem could make less significant the statistical analysis used to empirically demonstrate the validity of the proposed approach.

In order to provide further insights into the usefulness of strain rate assessments for earthquake prediction, the next section describes the results we have obtained by the study of this problem in the Italian area, which is actually covered by a fairly dense network of GPS permanent stations.

3. Geodetic strain rate assessments and recent seismicity in Italy

In order to reconstruct the present-day kinematic pattern in the Italian peninsula, we have analyzed the data acquired by 403 GPS permanent stations during the period January 1, 2001 - December 31, 2013. The network considered and the time series analysis adopted to obtain the residual horizontal velocity values shown in Fig. 1 are described by Cenni *et al.* (2012, 2013).

The pattern of average horizontal velocities indicates that the outer (Adriatic) sector of the Apennine belt moves faster than the inner (Tyrrhenian) side of the belt, in line with the long-term kinematics indicated by the Quaternary deformation pattern (Mantovani *et al.*, 2009, 2010, 2012; Cenni *et al.*, 2012, 2013).

We estimated the horizontal strain rate pattern related to the above velocity field by a weighted least-square method. In particular, the 2D strain rate tensor has been evaluated on a regular grid, applying the algorithm developed by Shen *et al.* (1996). The contribution of each station velocity to the strain rate computed on a given node of the grid is weighted by using the associated uncertainties and the exponential scaling function exp($-d_{ik}/D$), where di_k is the distance between the *i*th node of the grid and the *n*th GPS station, and *D* is the distance decay factor. In order to improve the reliability of the results, we have integrated the weighted least-

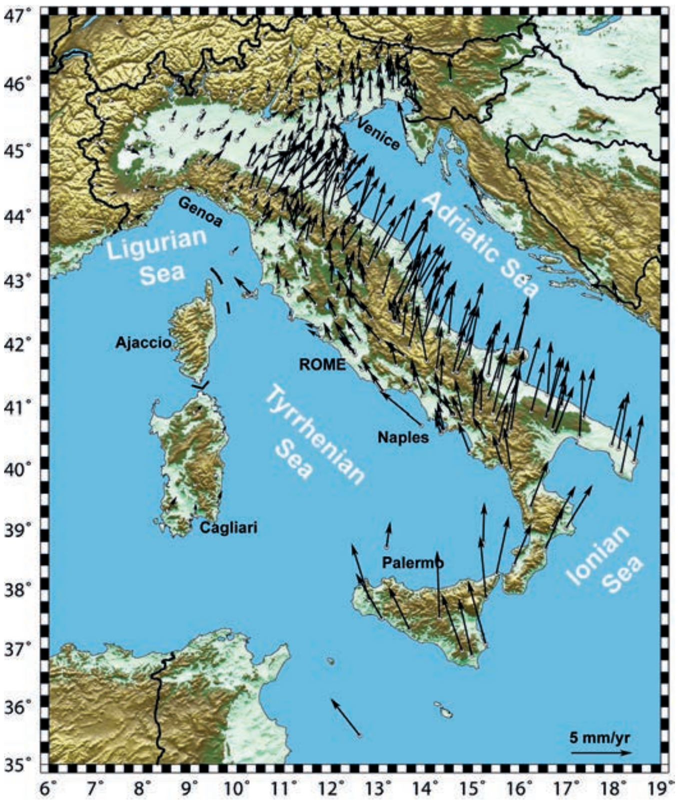

Fig. 1 - Residual horizontal GPS velocities with respect to an Eurasian fixed frame [Euler pole at 54.23° N, 98.83° W, ω= 0.257°/Ma, Altamimi *et al*., (2012)]. See Cenni *et al*. (2012, 2013) for details about the GPS network and data analysis.

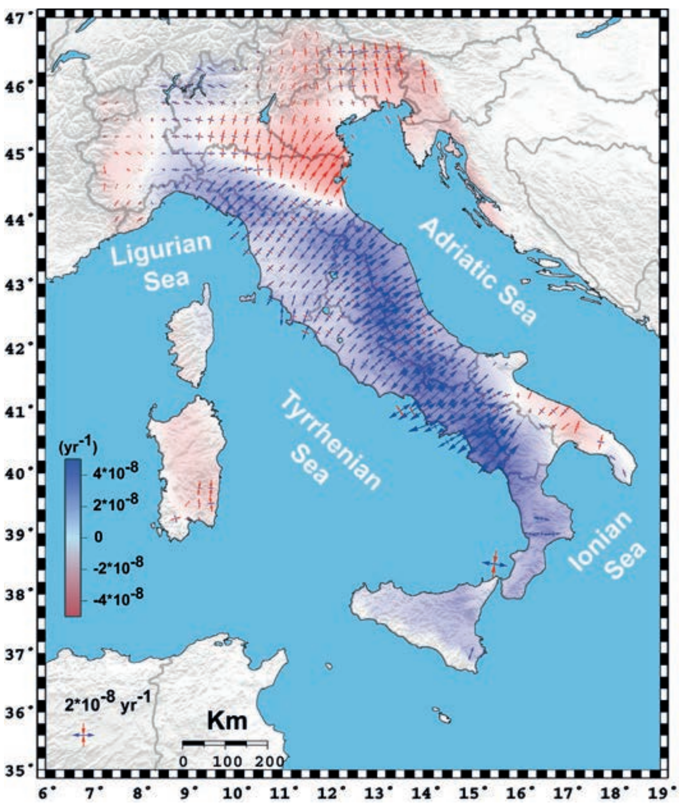

Fig. 2 - Horizontal strain rate field inferred from the velocity field shown in Fig. 1 using the weighted square method described in the text. This result was obtained by using a smoothing distance (or distance decay factor) of D=50 km, on a regular grid of 0.25° × 0.25°. Red converging and blue diverging arrows indicate principal axes of shortening and lengthening, respectively. The 2D dilatation-rate field (red compressional and blue extensional) is also shown in the figure.

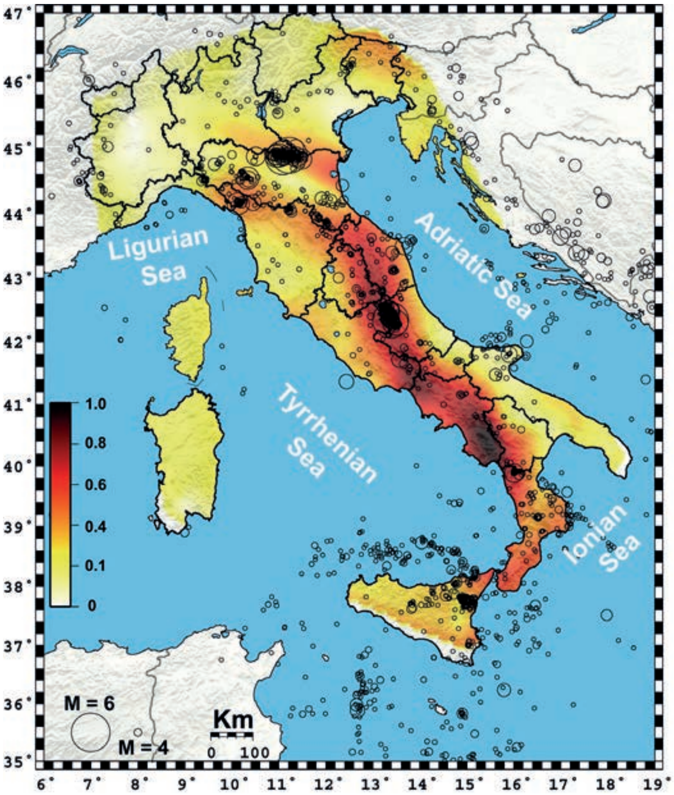

Fig. 3 - Pattern of NTSR, estimated from the 2D velocity field (Fig. 2) on a regular grid of 0.1°× 0.1°, using a smoothing distance of 50 km. The *TSR* values are normalized by using the maximum computed amplitude ($4.1 \cdot 10^{-8}$ yr^{-1}). Circles indicate the locations of the earthquakes with magnitude $M \geq 3.0$ and depth lower than 40 km, occurred since 2005/04/16 to 2013/12/11 (ISIDe Working Group, 2010).

square method with two geometric criteria, as suggested for instance by Teza *et al*. (2008, 2012) and Cenni *et al*. (2012). In particular, the strain rate values are taken as acceptable only when: 1) at least three GPS sites are located at a distance lower than *D* from the node considered and 2) such sites are uniformly distributed in the surrounding region (one in each 120° angular sector). Fig. 2 shows the results obtained on a regular grid space 0.25° × 0.25°, using a *D* value of 50 km, that is about three times the average spacing of the network.

In order to relate the strain rate to the recent seismicity, independently from the deformation style, we have estimated the total strain rate (*TSR*), which is defined by the following relation:

$$TSR = \sqrt{\left(\dot{\varepsilon}_{11}^2 + \dot{\varepsilon}_{22}^2 + 2 \cdot \dot{\varepsilon}_{12}^2\right)}$$

where $\dot{\varepsilon}_{11}$, $\dot{\varepsilon}_{22}$ and $\dot{\varepsilon}_{12}$ are the horizontal components of the strain rate tensor (e.g., Kreemer *et al*., 2003; Riguzzi *et al*., 2012).

Fig. 3 shows the normalized *TSR* distribution (*NTSR*) obtained on a denser, regular grid space 0.1° × 0.1° using a *D* value of 50 km, along with the distribution of seismicity from April 2005 to December 2013. Seismic data are taken from the Italian Seismic Instrumental and parametric Data-basE (ISIDe Working Group, 2010). Gasperini *et al*. (2013) indicate the magnitude M=1.9 as a reasonable completeness threshold for the above catalogue. Instead, Schorlemmer *et al*. (2010) suggest the value M=2.5. To avoid possible bias due to data incompleteness, we have considered the seismic events with $M \geq 3.0$ only.

The comparison between the *NTSR* pattern and the seismicity distribution (Fig. 3) suggests

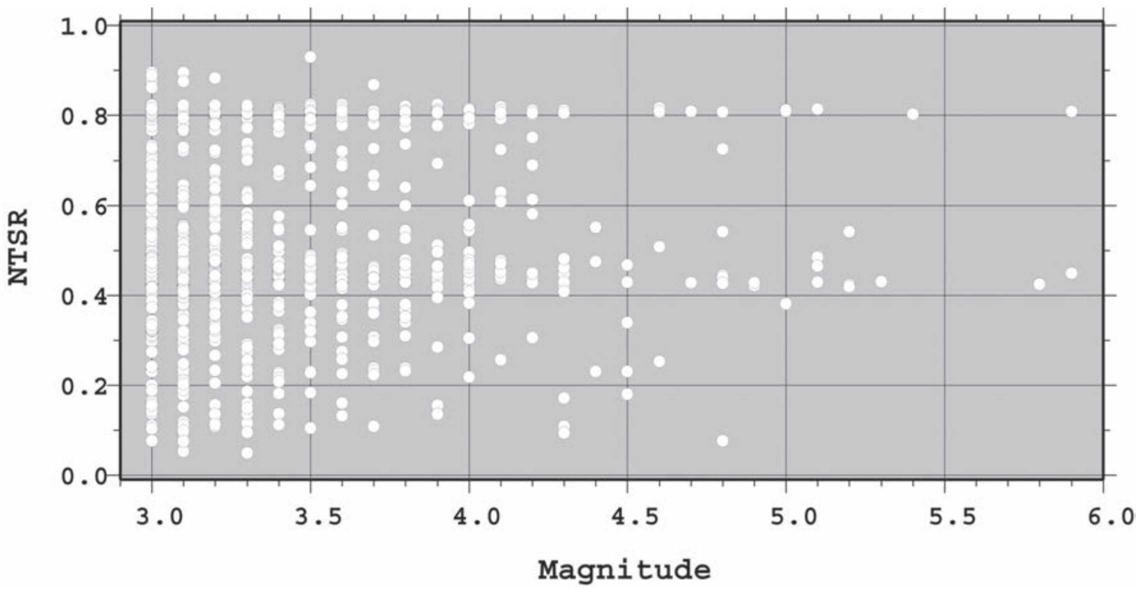

Fig. 4 - Pattern of the *NTSR* versus the magnitude of the earthquakes considered in Fig. 3. The *NTSR* values are evaluated at the epicentre coordinates for events with magnitude $M \geq 3.0$ and depth lower than 40 km. The geometrical criteria described in the text are adopted for this analysis, therefore the strain rate has not been estimated for 894 out of the 1941 earthquakes.

a weak correlation between local maxima of geodetic strain rate and the location of recent earthquakes, at variance with the conclusions drawn by Riguzzi *et al*. (2012). To better explain this result, it may be useful to plot the *NTSR* versus earthquakes magnitude (Fig. 4). To this purpose, we have estimated the *NTSR* at the epicentral coordinates of each seismic event, only considering the values for which the geometrical criteria described above are fulfilled (i.e., for 1047 out of 1941 events). Fig. 4 points out that for most magnitude values, both low and high *NTRS* values are present. Obviously, these results have been considerably influenced by the adoption of the geometrical criteria mentioned above. In particular, Fig. 5a shows that the geometrical criteria are not satisfied for seismic events with M<5.1. Since about 700 events have M<4.0 (Fig. 5b), one can observe that the most important earthquakes occurred in the considered time span have been included in the comparison depicted in Fig. 4.

Further investigation about the possible influence of the geometrical criteria on the map shown in Fig. 3 has been carried out by another test. Fig. 6 shows the distribution of the points for which such geometrical criteria are fulfilled in two cases relative to different values of the parameter *D*. In particular, this fact can explain why the results that we have obtained (adopting geometrical criteria) are not compatible with the results of other studies (Riguzzi *et al*., 2012) carried out without adopting such kind of criteria.

4. Geodetic monitoring of post-seismic relaxation and recognition of its possible effects on seismicity in the zones surrounding the triggering earthquake

Large earthquakes cause time-dependent redistribution of strain and stress in the surrounding regions, a phenomenon usually defined as post-seismic relaxation (e.g., Feigl and Thatcher,

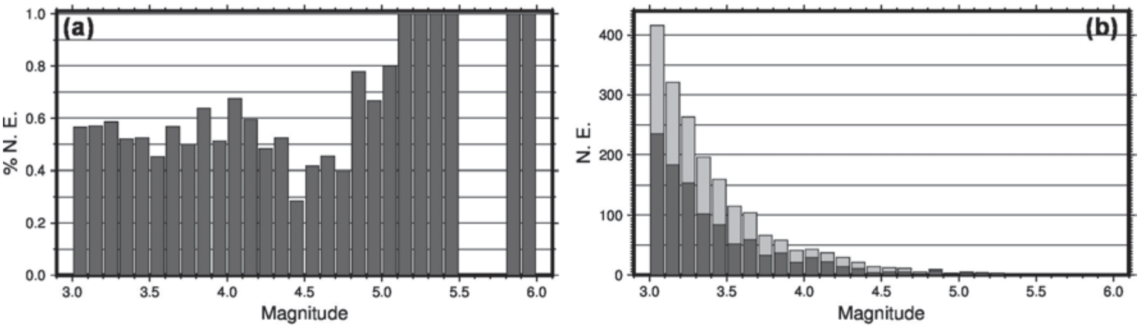

Fig. 5 - a) Frequency distribution (% NE) of the earthquakes whose epicentral location satisfies the geometrical criteria for the *NTSR* (see text for explanations). The earthquakes reported in Fig. 3 (ISIDe Working Group, 2010) have been ordered by magnitude values in 30 bins, each bin having an M=0.1 width. Then, for each bin the number of seismic events that satisfy the geometrical criteria has been compared to the total number of events in the bin. b) Distribution of earthquake with M>3.0 (Fig. 3, ISIDe Working Group, 2010). The dark and light gray bars show the number of events (NE) that satisfy and not satisfy the geometrical criteria, respectively.

2006 and references therein). When such perturbation reaches mature faults (i.e., favourably oriented and close to failure) it may induce further seismicity (e.g., Harris, 1998; Scholz and Gupta, 2000; Steacy *et al.*, 2005; Lou and Liu, 2010). The important point is that post-seismic relaxation has been repeatedly detected by means of GPS measurements (e.g., Pollitz *et al.*, 2006; Ergintav *et al.*, 2009; Hammond *et al.*, 2009, 2010; Ozawa *et al.*, 2011; Shao *et al.*, 2011; Vigny *et al.*, 2011). This phenomenon has been recognized as responsible of long-range (hundreds of km) and long-term (years) interactions of seismic sources (e.g., Anderson, 1975; Rydelek and Sacks, 1990, 2001, 2003; Pollitz and Sacks, 1995; Pollitz *et al.*, 1998, 2004; Mikumo *et al.*, 2002; Marsan and Bean, 2003; Viti *et al.*, 2003; Freed, 2005; Freed *et al.*, 2007; Lay *et al.*, 2009). In particular, it is suggested that the highest probability of induced seismicity corresponds to the arrival of the greatest values of the strain rate generated by the triggering earthquake (e.g., Pollitz *et al.*, 1998). This hypothesis is consistent with the evidence and arguments described in a number of papers concerning the presumed interaction between peri-Adriatic seismic sources (Viti *et al.*, 2003, 2012, 2013; Mantovani *et al.*, 2010, 2012).

The framework of evidence described above may delineate the deterministic basis for recognizing the zones where the probability of strong earthquakes may significantly increase in the next years. In order to better define the proposed methodology, we describe the operations that should be carried out after the occurrence of a strong earthquake in one of the peri-Adriatic zones:

- the expected time pattern of the strain and strain rate caused by the triggering earthquake is predicted by using the modelling approach described in previous works (Viti *et al.*, 2003, 2012, 2013; Mantovani *et al.*, 2012). In particular, the comparison of the estimated strain and strain rate amplitudes with the perturbation of such parameters caused by Earth tides may be used to recognize whether induced seismicity can or cannot occur (e.g., Viti *et al.*, 2003 and references therein). This evaluation is only tentative, due to the uncertainty about the structural and rheological parameters adopted in numerical modelling of post-seismic relaxation;
- the actual development in time and space of the perturbation induced by the triggering

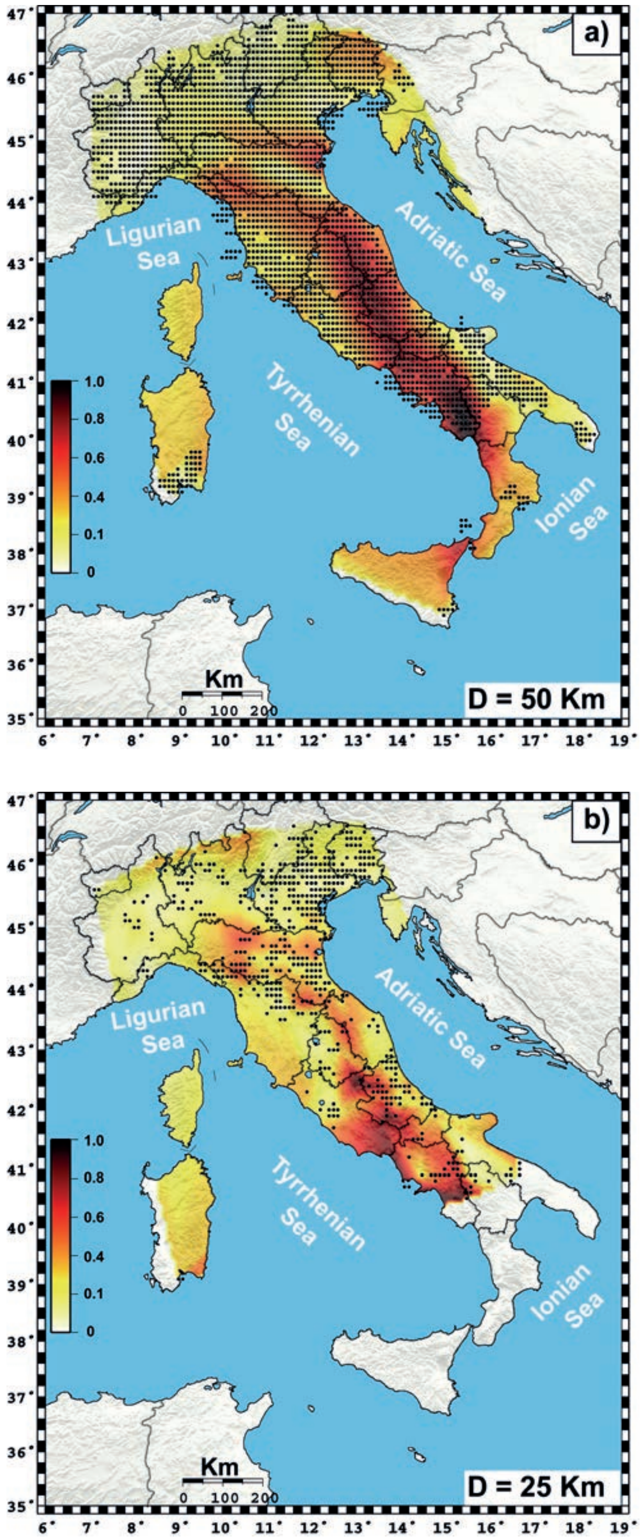

Fig. 6 - Distribution of the *NTSR* estimated on a regular grid of 0.1°× 0.1°, using different smoothing distances: 50 km (a) and 25 km (b). The black dots indicate the position of the grid nodes where the geometrical criteria described in the text are fulfilled.

earthquake can be monitored by using the GPS data acquired by the fairly dense station network now available in the Italian area (Cenni *et al.*, 2012, 2013). This operation can be carried out by analysing the series of daily positions, focusing in particular on the recognition of eventual non-linear velocity patterns. These last features may allow reconstructing the time pattern of strain and strain rate in the sites surrounding the earthquake that caused the transient strain perturbations. The identification of non linear trends in the time series may be achieved for instance by using moving window techniques (e.g., Baldi *et al.*, 2008). Using the results of the above mentioned analysis, one could try to predict when the maximum amplitude of strain perturbation will occur in the zones of interest. The above information can also be useful for improving our knowledge on the structural and rheological properties of the zones through which the studied perturbation has propagated. This result can be achieved by comparing the velocity and strain deduced by GPS data with the corresponding patterns predicted by numerical modelling of post seismic relaxation (e.g., Hammond *et al.*, 2010).

Conclusions

GPS space geodetic data have widely been used to tentatively recognize both short and middle-long-term precursors of earthquakes. As concerns short-term precursors, we note that most of the analyses so far proposed are retrospective, and no one of the destructive shocks occurred worldwide since the beginning of this century has been predicted by the help of such kind of investigation. In particular, we argue that no unambiguous, systematic and widely accepted precursor, based on space geodetic observations, has been identified yet. This is in part related to the fact that the present knowledge about earthquake nucleation processes is still very limited.

The recognition of long-term precursors is mainly based on the determination of the velocity and strain rate fields by means of space geodetic measurements. Then, these fields are tentatively used to predict the future behaviour of seismogenetic structures in the zones involved. This procedure is conceptually plausible, but the predictions performed might be affected by considerable uncertainty, since the relation between the ongoing strain rate, computed by the present geodetic velocity field and the proximity to failure of the faults involved is still poorly understood. In this regard, Riguzzi *et al.* (2012) have advanced an hypothesis that could delineate an important tool for earthquake prediction in the Italian region: the zones characterized by relative minima of the strain rate field may be associated with the highest probability of earthquake occurrence. In support to that suggestion, the above authors present the comparison between the *NTSR* pattern, computed by the present geodetic velocity field, and the distribution of recent seismic events (2005-2013) in the central Mediterranean region. The above authors claim that such comparison points out a significant agreement between strain minima and major seismic events. However, we think that some aspects of this check must be reconsidered. In particular, the accuracy of strain assessment may be scarce in a number of zones, due to the little number of GPS stations, mainly concerning the Dinaric area, where numerous seismic events are reported. The importance of this problem is enhanced by the fact that the computation of strain rate in a site, as performed in the above work, is not conditioned by the features of the local network. Furthermore, we do not share the opinion that the comparison given by Riguzzi *et al*. (2012) provides satisfactory results, since our results show that both large and low magnitude events are located in areas characterized by low and high strain rate values.

We argue that, at present, the best chance to gain information about future seismicity by the help of geodetic data is related with the monitoring of post-seismic relaxation triggered by a strong earthquake. This analysis may allow the reconstruction of the spatio-temporal development of the perturbation triggered by a strong earthquake, aimed at recognizing when the highest values of the induced strain and strain rate are expected to arrive in the zones of interest. The possibility of applying the above procedure for middle-term earthquake prediction is supported by several examples of possible interaction between seismic sources in the central Mediterranean area (Viti *et al*., 2003, 2012, 2013; Mantovani *et al*., 2008, 2010, 2012).

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