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Post-seismic relaxation and earthquake triggering in the southern Adriatic region

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SUMMARY

An attempt at quantifying post-seismic relaxation triggered by decoupling earthquakes along the eastern thrusting border of the Adriatic plate (southern Dinarides) is carried out by finite element modelling, with a model constituted by an elastic lithosphere riding on a viscous asthenosphere.

In particular, it is investigated the possibility that the above phenomenon is responsible for the fact that in the last two centuries most major earthquakes in the southern Dinarides ($M_S >$ 6) have been followed, within a few years, by intense, mainly tensional, earthquakes in southern Italy, i.e. the zone lying on the opposite margin of the Adriatic plate.

This analysis has been applied to the last example of the supposed seismic interrelation, i.e. the triggering 1979 April 15 Montenegro event ($M_S = 6.7$) and the presumably induced 1980 November 23 Irpinia earthquake in the southern Apennines ($M_S = 6.9$). Results indicate that the strain induced in the southern Apennines by the triggering event has significant amplitude, since it largely exceeds the effect of earth tides, and the principal stress axes are consistent with those of southern Apenninic earthquakes. The order of magnitude of the time delay between the Montenegro and Irpinia events (1.6 yr) could be explained by assuming that earthquake triggering is most probable when the highest values of the induced strain rate reach the southern Apennines. In particular, this interpretation predicts the observed time delay when a model diffusivity of 400 m^2 s^{-1} is assumed. The constraints that this diffusivity value may pose on the structural and rheological features of the crust–upper-mantle system in the study area are discussed. It is shown that the effects of the Montenegro event on the present velocity field are comparable to, though systematically lower than, the velocities suggested by geodetic observations in the Italian region. This suggests that geodynamic interpretations of geodetic data given without taking into account possible transient effects on the kinematic pattern, as those related to post-seismic relaxation, may be incorrect. Experiments carried out by tentatively simulating the presence of subducted lithosphere along the western margin of the Adriatic plate as a lateral variation of diffusivity, have shown that this structural feature may emphasize E–W tensional strains in the southern Apennines.

Key words: Adriatic plate, earthquake triggering, mantle viscosity, post-seismic relaxation, stress diffusion.

1 INTRODUCTION

It is widely recognized that large earthquakes cause the redistribution of displacement, strain and stress in the surrounding regions (e.g. Elsasser 1969; Anderson 1975; Lehner *et al.* 1981; Rice & Gu 1983; Li & Rice 1987; Rydelek & Sacks 1990). These perturbations propagate at rates much lower than those of seismic waves, due to the coupling of the upper elastic lithospheric layer with the underlying viscous (or viscoelastic) asthenosphere, and may trigger seismic activity when they reach seismogenic zones.

Quantifications of this phenomenon by simplified diffusion models have provided possible explanations for post-seismic geodetic velocity patterns and for the time pattern of seismic activity in the circum-Pacific zones (Rydelek & Sacks 1990; Savage & Svarc 1997; Pollitz *et al.* 1998a,b, 2000; Deng *et al.* 1998), central Asia (Chéry

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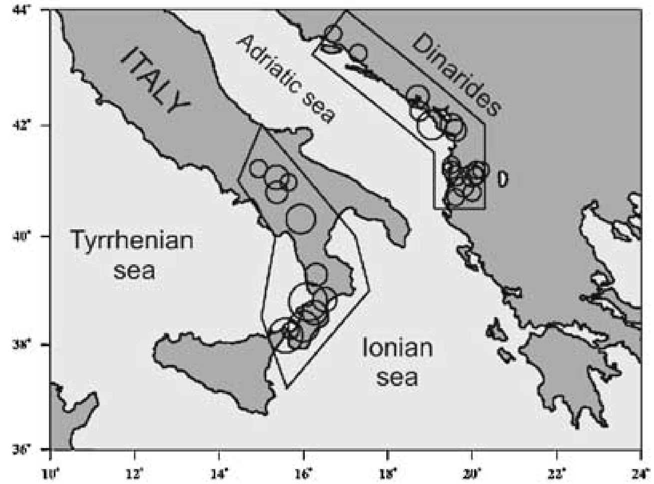
et al. 2001; Calais *et al.* 2002), Middle-East (Piersanti *et al.* 2001) and in the Aegean-Anatolian region (Mantovani *et al.* 2001a; Cenni *et al.* 2002; Papadopoulos 2002). Possible effects of post-seismic relaxation, in terms of seismic interrelations between tectonically connected zones, have also been observed in the Mediterranean area (Mantovani *et al.* 1986, 1987, 1991, 1997a; Mantovani & Albarello 1997). In particular, it has been pointed out that major earthquakes occurring in the southern Dinarides in the last two centuries have been followed within a few years by strong shocks in southern Italy (Fig. 1).

An analytical quantification of viscoelastic relaxation in the Adriatic area, with a simplified 1-D model (Albarello & Bonafede 1990), has provided possible explanations for the time delays between the triggering Dinaric events and the presumably induced Italian earthquakes. However, a more detailed knowledge of the major features of post-seismic relaxation processes in the central

(a)



(b)

Southern Dinarides	Southern Italy
1780 (6.6)	1783 (7.1,6.7,6.7,6.1)
1843 (6.2)	
1851 (6.2)	1851(6.1)
1855(6.5)	1857(7.0)
1865(6.2)	
1869(6.2)	
1870(6.4)	1870(6.6)
1898(6.3)	
1905(6.6)	1905(7.5)
1906(6.5)	
1907(6.2)	1908(7.3)
1923(6.2)	
1926(6.1)	1930(6.6)
1959(6.4)	
1962(6.1)	1962(6.2)
1979(7.1,6.3)	1980(6.5)

(c)

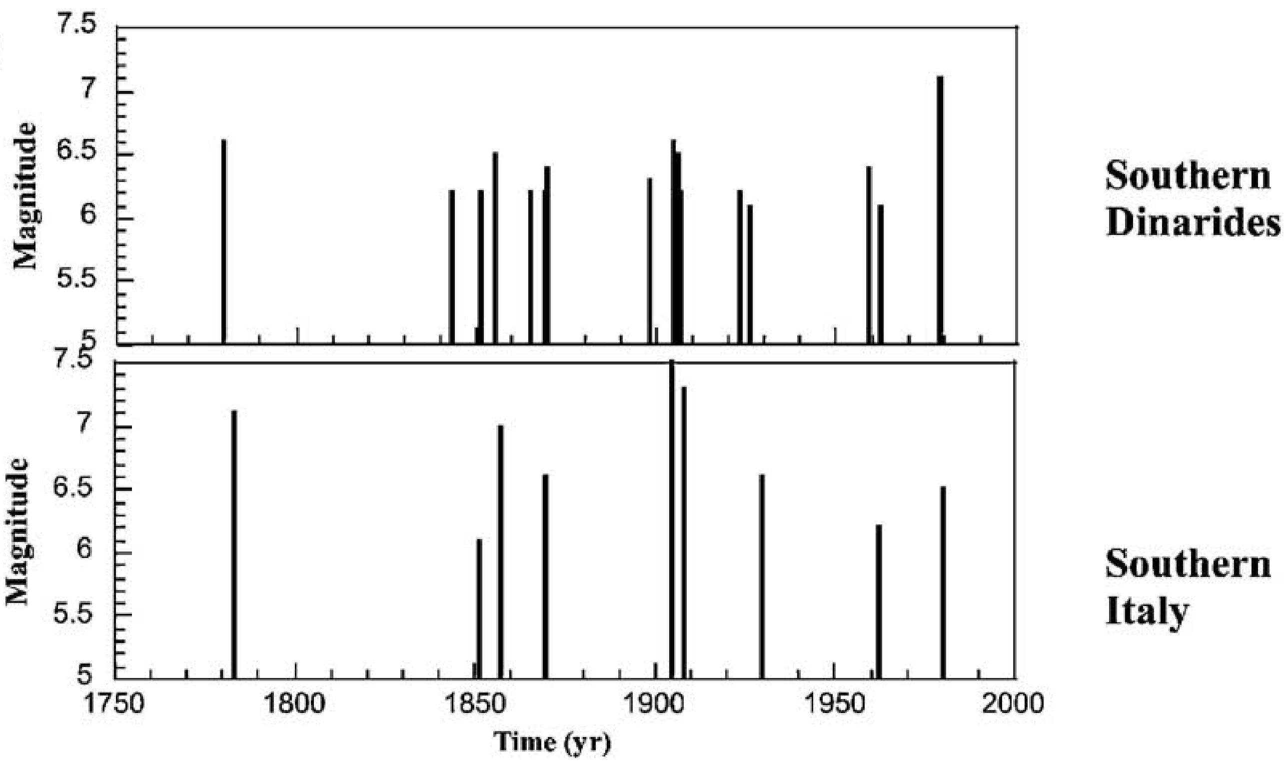


Figure 1. Seismic interrelation between southern Dinarides and southern Italy (from Mantovani & Albarello, 1997). (a) Geometry of the two zones and epicentres of major earthquakes ($M_S > 6$) that have occurred since 1770 (b) List of earthquakes with magnitude M_S given in brackets (c) Time series of earthquakes in the two zones.

Mediterranean area, as magnitude, time evolution and lateral variations of the expected perturbations, could be useful for planning suitable observations of the above effects by geodetic and geophysical techniques. This work describes an attempt to gain such information by a 2-D numerical approach, based on the classical Elsasser's (1969) diffusion model, constituted by an elastic lithosphere overriding a viscous asthenosphere.

2 TECTONIC SETTING IN THE STUDY AREA

Fig. 2 shows the main tectonic features of the central Mediterranean area. Tectonic activity in this region is mainly connected with the interaction of the Adriatic plate, a continental block of African affinity, with the surrounding orogenic zones.

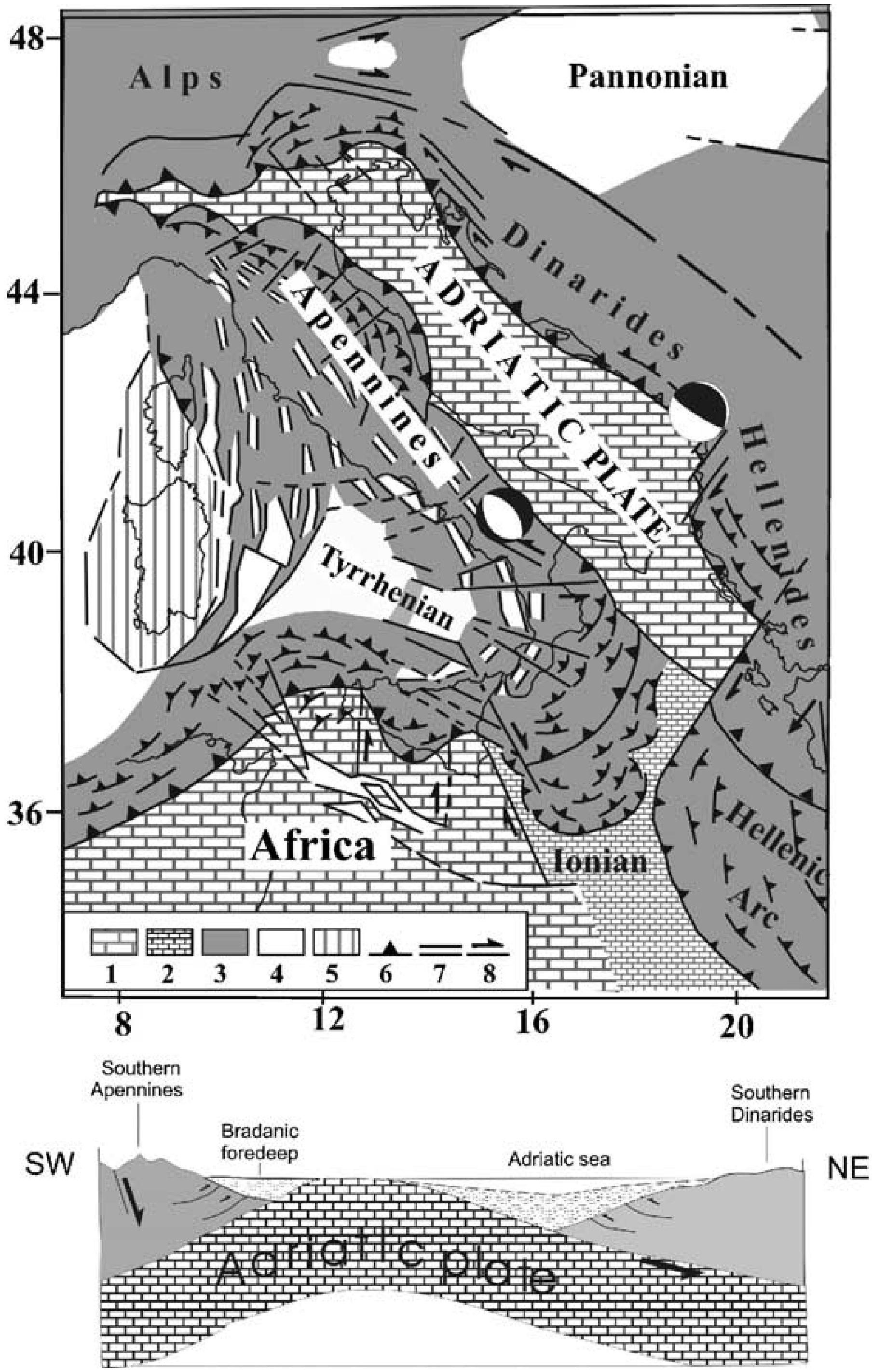


Figure 2. Tectonic sketch of the central Mediterranean region. 1) African/Adriatic continental domain; 2) Ionian oceanic domain; 3) Main deformation belts; 4) Neogenic basins; 5) Corsica-Sardinia block; 6, 7, 8) Main compressional, tensional and transcurrent features. Details on the evolution and active tectonics of the study area are given by Mantovani *et al.* (1997b, c). Focal mechanisms of the triggering Montenegro earthquake (1979 April 15) and of the presumably induced Irpinia shock (1980 November 23) are reported in the respective epicentral zones (Boore *et al.* 1981; Giardini 1991). The section schematically shows the underthrusting of the Adriatic plate beneath both the southern Dinarides and southern Apennines and the sense of fault slip for the two earthquakes considered (big arrows).

Neotectonic evidence and earthquake focal mechanisms (e.g. Anderson & Jackson 1987; Mantovani *et al.* 1997b,c; Viti *et al.* 2001) suggest that the above interaction is characterized by a compressional regime in the Dinaric border and by a dominant tensional regime in the Italian border (Apenninic belt). This strain pattern has been interpreted as an effect of the Adriatic anticlockwise rotation (Anderson & Jackson 1987; Mantovani *et al.* 1997b,c). In the framework of this interpretation, the major earthquakes in the Dinaric border are interpreted as decoupling dislocations between the underthrusting Adriatic plate and the overriding Dinaric belt. The roughly E–W tensional stress pulse triggered by these events propagates up through the Adriatic plate to reach the Apenninic belt, where it may induce seismic activity (Mantovani *et al.* 1997a). In the next chapter, an attempt at quantifying the implications of the above interpretation is described.

3 METHODOLOGY AND MODEL PARAMETERISATION

It is assumed that the mechanical behaviour of the study area may be modelled as an elastic layer (lithosphere) overlying a viscous one (asthenosphere). In the Elsasser (1969) approximation, the components of the viscous stress acting at the base of the lithosphere are:

$$\tau_{zx} \cong \frac{\eta}{H_a} u_{x,t}$$
$$\tau_{zy} \cong \frac{\eta}{H_a} u_{y,t} \qquad (1)$$

where u_x and u_y are the components of the displacement along the horizontal axes (x, y), t is time, η is the Newtonian viscosity and H_a is the thickness of asthenosphere; commas indicate partial differentiation. In the thin sheet approximation, the components of the lithospheric stress tensor acting along vertical planes vanish (England & McKenzie 1982). In this case, the equilibrium equations for the lithosphere are:

$$H_l(\sigma_{xx,x} + \sigma_{yx,y}) = \tau_{zx}$$
$$H_l(\sigma_{xy,x} + \sigma_{yy,y}) = \tau_{zy} \qquad (2)$$

where the stress components have been averaged over the lithospheric thickness H_l.

Equation (2) can be expressed in terms of horizontal displacement components by taking into account the displacement-strain relationships and the constitutive equations of linear elasticity:

$$\frac{2(1-\nu)}{(1-2\nu)} u_{x,xx} + \frac{1}{(1-2\nu)} u_{y,yx} + u_{x,yy} = \frac{2(1+\nu)}{D} u_{x,t}$$
$$u_{y,xx} + \frac{1}{(1-2\nu)} u_{x,xy} + \frac{2(1-\nu)}{(1-2\nu)} u_{y,yy} = \frac{2(1+\nu)}{D} u_{y,t} \qquad (3)$$

where ν is the Poisson's ratio of the lithosphere and $D = EH_lH_a/\eta$ is the model diffusivity ($m^2\ s^{-1}$), i.e. the parameter that controls the stress diffusion rate through the model. E is the Young's modulus.

Equation (3) is solved numerically by a finite element approach. In particular, the implicit Crank-Nicolson method (e.g. Press *et al.* 1992) is used to discretize time and to reduce the parabolic form of (3) to a non-homogeneous elliptic form, more suitable to be handled by numerical methods (see Sewell 1981, 1986, for details). By assuming that the thin sheet approximation is appropriate in our context, the zone considered (Fig. 3) has been modelled by a 2-D grid of triangular elements. The grid, which has been generated by an automatic procedure in order to achieve suitable triangular shapes and distribution (EasyMesh shareware software, Niceno 2002), contains 647 nodes and 1190 triangles. This kind of approach allows the modelling of heterogeneous structures by assigning different material properties (D, ν) to each node of the grid, which may particularly be useful when complex laterally heterogeneous zones, such as the central Mediterranean one, are involved.

Figure 3. Model adopted and location of the fault where a displacement (u) is imposed to simulate the horizontal component of the coseismic slip of the Montenegro earthquake (1979 April 15 $M_S = 6.7$). The estimated seismic moment ($M_0 = 4.6 \times 10^{19}$ N m, Boore *et al.* 1981) and the distribution of aftershocks (Console & Favali 1981) are compatible with a fault length (L) of 50–100 km and a slip (u) of 1–2 m. The black dot located in the southern Apennines is the point at which the time dependence of post-seismic displacement, velocity, strain and strain rate is computed. H_l, H_a = lithospheric and asthenospheric thickness, E = lithospheric Young's modulus, η = asthenospheric viscosity.

The model is stressed by imposing an instantaneous displacement at the long side of a thin rectangular box (Fig. 3), which simulates the rupture zone of the last strong earthquake which occurred at the eastern border of the Adriatic plate (1979 April 15 Montenegro, $M_S = 6.7$, Boore *et al.* 1981). This condition may allow for the finite size of the fault zone and for magnitude and direction of the observed coseismic slip. Fig. 3 shows the region considered and the location of the zone where the displacement is imposed (southern Dinarides). This kinematic boundary condition aims at simulating the horizontal motion of the Adriatic zone most close to the activated dip-slip thrust fault. The aftershocks distribution (Console & Favali 1981) and the estimated seismic moment (Boore *et al.* 1981) are compatible with a fault length of 50–100 km. Assuming that the coseismic slip is related to fault length by the scale relationship suggested by Wells & Coppersmith (1994), it ranges between 1 and 2 m. Along the external boundaries of the zone represented in Fig. 3 only motions parallel to the sides of the model are allowed.

In computations, we assumed for the lithosphere the values 0.25 and 10^{11} Pa for Poisson's ratio (ν) and Young's modulus (E), respectively, as suggested by current knowledge on elastic properties of crustal and mantle rocks (Dziewonski & Anderson 1981; Kirby 1983; Christensen 1996).

The choice of the other parameters defining diffusivity (H_l, H_a, η) is more problematic, since quite different hypotheses are reported in literature about this problem. Some authors (e.g. Albarello & Bonafede 1990; Piersanti *et al.* 1997; Mantovani *et al.* 2001a) assumed that the elastic and viscous layers involved in stress diffusion correspond to the classical concept of lithosphere and asthenosphere, derived from seismological techniques, as surface wave dispersion and tomographic analysis (e.g. Zhang & Tanimoto 1993; Ekström *et al.* 1997). This view implies that the base of the lithosphere corresponds to the $\sim$1300°C isotherm surface (Pollack & Chapman 1977; Artemieva & Mooney 2001), which provides lithosphere and asthenosphere thickness of the order of 100 km for cold continental domains, such as e.g. the Adriatic block (Panza & Suhadolc 1990; Martinez *et al.* 2000).

Other authors (e.g. Rydelek & Sacks 1990; Pollitz 1997) have instead suggested that the elastic thickness controlling stress diffusion is considerably less than the seismological lithosphere, being related to the long-term rheological behaviour of the crust–upper-mantle system (Ranalli & Murphy 1987; Lobkovsky & Kerchman 1991). This hypothesis implies that the elastic and viscous layers of the Elsasser's model respectively correspond to the brittle seismogenic upper crust and to the ductile lower crust. Based on the reconstruction of rheological profiles in various kinds of continental domains (e.g. Ranalli & Murphy 1987; Viti *et al.* 1997), this last definition implies a thickness of $\sim$10–20 km for both the lithosphere and asthenosphere, which is about an order of magnitude less than seismological thickness. Adopting these last values for H_l and H_a would imply a remarkable reduction of the model diffusivity and, in general, a minor role played by viscous relaxation with respect to other physical processes possibly responsible for post-seismic strain, as, e.g. fault afterslip (Pollitz *et al.* 1998a; Calais *et al.* 2002).

Estimating the viscosity of the crust–mantle system is as problematical as defining layer thickness. It is known from laboratory experiments (Kohlstedt *et al.* 1995) that ductile rock deformation is ruled by a non-linear flow law, which means that effective viscosity depends on strain rate (Ranalli 1995). Thus, linear models, such as the Elsasser's one, represent an oversimplification of reality. Furthermore, the evolution of transient phenomena, as e.g. post-seismic relaxation and postglacial rebound, involves variations of strain rate with time and is often modelled by assuming *a priori* the behaviour (elastic, viscoelastic, viscous) of model layers. In this kind of procedure, it is not clear to what extent viscosity estimates depend on the assumptions made about the model itself. As an example, postglacial rebound models usually consider the crust as an elastic body, whereas the deformation induced by deloading is considered to be taken up by the whole mantle (from the Moho to the core), and no asthenospheric layer is considered (Lambeck & Kaufmann 2000). In these models, the role of a mantle asthenosphere has been not fully explored yet, although recent works suggest that a relatively low viscosity of this layer (10^{18}–10^{19} Pa s) could account for glacial isostatic adjustments (Wieczerkowski *et al.* 1999; Wang 2001).

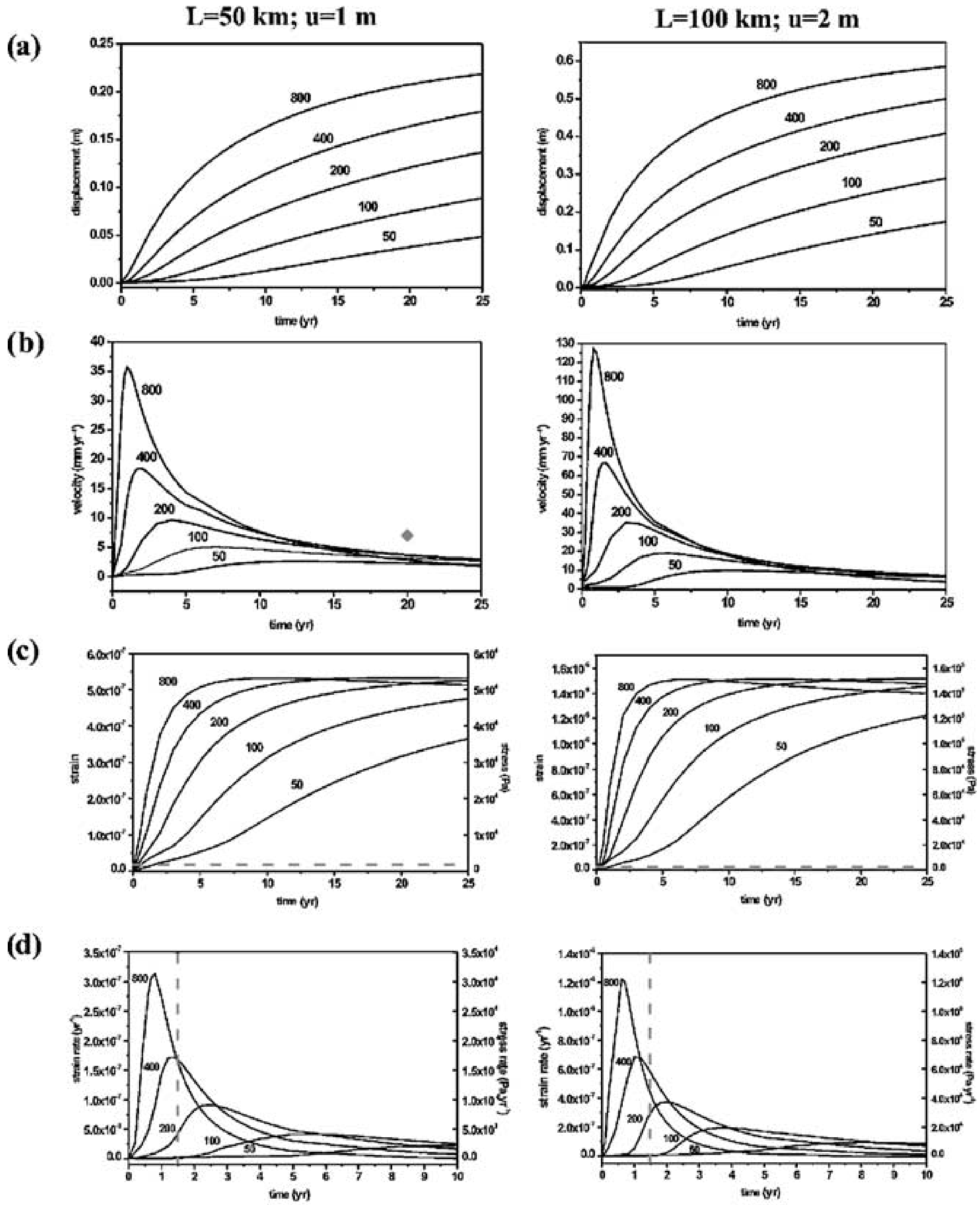
Bearing in mind that current viscosity values should be taken with caution, quantifications of post-seismic relaxation generally indicate that the lower crust is more viscous than the mantle by about one order of magnitude. In particular, viscosity values in the range 1–3 $\times 10^{19}$ Pa s and 0.5–3 $\times 10^{18}$ Pa s have been found for the lower crust and the asthenospheric mantle respectively (Pollitz & Sacks 1996; Pollitz *et al.* 1998a,b, 2000). These values fairly well agree with independent estimates, based on studies of thermal convection (Davaille & Jaupart 1994) and seismic anelasticity (Sato 1991).

Other insights about this problem are provided by the reconstruction of viscosity profiles based on knowledge of rock rheology, seismic layering and thermal state of the crust–mantle system (Meissner 1986). In particular, computations carried out for continental domains characterized by surface heat flow and crustal thickness comparable to that estimated for the Adriatic plate (Scarascia *et al.* 1994; Cataldi *et al.* 1995), show that the viscosity of the lower crust, the lithospheric mantle and the asthenospheric mantle respectively falls in the range 10^{21}–10^{22}, 10^{19}–10^{21} and 10^{18}–10^{19} Pa s, respectively (Meissner 1986). These results refer to a value of strain rate (10^{-14} s^{-1}) comparable to that involved in post-seismic relaxation, as reported in the next section (Fig. 4d). The above ranges suggest that the asthenospheric mantle is much less viscous than the lower crust, in line with the fact that very different temperatures presumably characterize these layers ($\sim$500°C and $\sim$1300°C for lower crust and asthenosphere, respectively). However, one must be aware that the reconstruction of viscosity profiles is affected by large uncertainties, mainly concerning the thermal and rheological parameters of the crust–mantle system (Meissner 1986).

Given the above difficulties in assessing reliable values of layers thickness and viscosity, one could try to obtain indirect information on these parameters by an empirical procedure, i.e. by looking for the diffusivity values which may best account for the time delay between the Montenegro triggering event and the Irpinia earthquake, under the assumption that earthquake triggering is most likely when the highest values of strain rate reach the southern Apennines, provided that the amplitude of strain perturbations significantly overcome a minimum threshold, that we identify with the earth tide effect. The above assumption about the strain rate, previously advanced by Pollitz *et al.* (1998b), is consistent with the fact that the thickness of the brittle layer, and thus the maximum size of the earthquake rupture zone, closely depends on strain rate (e.g. Sibson 1984; Smith & Bruhn 1984).

4 MAIN RESULTS

The empirical investigation mentioned above has been carried out by adopting different values of model diffusivity (D), fault length (L), and coseismic slip (u), as described in Fig. 4. A wide range of diffusivity values (from 4 to 800 m^2 s^{-1}) has been explored, in order



Table

D ($m^2\ s^{-1}$)	Hl (km)	Ha (km)	η (Pa s)
800	100	100	1.25E+18
500	100	100	2.00E+18
400	100	100	2.50E+18
200	100	100	5.00E+18
100	100	100	1.00E+19
50	50	50	5.00E+18
25	50	50	1.00E+19
4	20	20	1.00E+19

Figure 4. Time patterns of displacement (a), velocity (b), strain (c) and strain rate (d) for different values of model diffusivity (D, reported on the curves in $m^2\ s^{-1}$) and source dimensions of the 1979 Montenegro perturbing event (L and u, reported in the left and right columns). To handle scalar quantities only, the Euclidean norm of the strain and strain rate tensors is reported. For reference, an estimate of the strain induced by earth tides (Tsuruoka *et al.* 1995; Vidale *et al.* 1998) is reported in the strain pattern (c) as a horizontal dashed line. The vertical dashed line in the strain rate pattern (d) identifies the time delay between the triggering Montenegro event and the presumably induced 1980 Irpinia earthquake. The table shows a possible set of lithospheric and asthenospheric thickness and asthenospheric viscosity for each diffusivity value adopted in the investigation. In all experiments Young's modulus is $E = 10^{11}$ Pa. The time patterns for diffusivity values $<50\ m^2\ s^{-1}$ are not reported in the figures. See text for comments.

to encompass values of D compatible with the variety of layered models, from the 'seismological' to 'rheological' ones, discussed above. In order to give an idea about the implications of diffusivity on model parameters, the table in Fig. 4 reports possible values of H_l, H_a and η for each value of D. Since we are mainly interested in the time pattern of post-seismic effects in the southern Apennines, where the induced earthquake is expected to occur, we computed this kind of information (Fig. 4) for a point located in that zone (see Fig. 3).

Fig. 4(a) shows that the displacement induced by the triggering earthquake progressively increases with time, tending to an asymptotic value which corresponds to the deformation of an elastic lithosphere under the above mentioned boundary conditions. The values of displacement reached in the time interval reported in figures (25 yr) go from a few centimetres to more than 20 cm, for $L = 50$ km, or to more than 50 cm, for $L = 100$ km, depending on model diffusivity. This result indicates that, if the range here assumed for D is realistic, the effects of post-seismic relaxation on the displacement field should be detectable by geodetic observations, whose uncertainty is generally lower than 1 cm.

The time pattern of velocity (Fig. 4b) is characterized by a more or less pronounced maximum, depending on model diffusivity. This shape could give the opportunity to obtain important constraints on model diffusivity by geodetic observations. In fact, identifying the time location of the velocity maximum in one or more geodetic sites would significantly help to identify the value of D in the zone involved. This possibility, closely connected with the recognizability of the velocity peak, is more evident for the highest values of D (Fig. 4b). For diffusivity lower than 100 $m^2\ s^{-1}$ the low value of the maximum and the very slow increase of velocity over time would not allow distinguishing the expected peak from experimental uncertainties.

The time patterns of strain (Fig. 4c) show that the influence of diffusivity on this quantity is highest in the first decade and decreases considerably for longer time intervals. To evaluate tentatively the potentiality of the induced strain to trigger earthquakes, it is useful to compare its value with those expected as an effect of earth tides. Since this last phenomenon seems to be ineffective in triggering earthquakes (e.g. Vidale *et al.* 1998; Lockner & Beeler 1999), the related strain level (reported in Fig. 4c) can be taken as a lower boundary for the range of strain perturbation that may trigger seismicity. Estimates of this effect (Melchior 1983; Rice & Gu 1983; Tsuruoka *et al.* 1995; Vidale *et al.* 1998) indicate strain values of the order of 2–3 *times* 10^{-8}(for E = 10^{11} Pa), which are about one order of magnitude lower than the strain predicted for the Montenegro event (Fig. 4c). It is worth noting that for the lowest values of D (<50 $m^2\ s^{-1}$ for $L = 100$ km, and <100 $m^2\ s^{-1}$ for $L = 50$ km) the strain induced by post-seismic relaxation in the first 5 yr is comparable or slightly higher than the earth tide threshold. This could imply that such D values are not likely since they cannot account for the earthquake triggering effect implied by the proposed seismic interrelation (Fig. 1).

Similarly to velocity pattern, the shape of the strain rate time pattern (Fig. 4d) is characterized by a relative maximum. This pattern, considering that the probability of earthquake triggering seems to be higher for higher values of strain rate, could help to explain the observed time delays between Dinaric and Italian events. On the basis of the above interpretation, the value of diffusivity that could best account for the time delay of 1.6 yr between the 1979 Montenegro and 1980 Irpinia shocks roughly corresponds to 400 $m^2\ s^{-1}$. The shape of the time pattern of strain rate shows a weak dependence on source dimensions (Fig. 4d).

The table in Fig. 4 shows that assuming a seismological stratification (e.g. $H_l = H_a = 100$ km) the best-fit diffusivity value implies a viscosity of the mantle asthenosphere in the range indicated in Section 3. On the other hand, if one assumes a rheological stratification ($H_l = H_a = 20$ km), the above diffusivity implies a viscosity of 10^{17} Pa, which is out of the range estimated for the lower crust (see Section 3).

The hypothesis that stress diffusion involves seismological rather than rheological thickness of the lithosphere may also be supported by the following considerations. Assuming a Maxwell rheology, the lithosphere shows an elastic behaviour for times less than the relaxation time $\tau = \eta/\mu$, where μ is the elastic shear modulus. For characteristic times of a few years, such as those suggested by the seismic correlation considered here, the above condition is fulfilled for viscosities of the order of 10^{18}–10^{19} Pa s. For mantle rocks (e.g. wet olivine) and post-seismic strain rate of the order of $10^{-14}\ s^{-1}$, the above viscosity values correspond to the temperature interval 1200–1350°C, that defines the bottom of the thermal lithosphere (Pollack & Chapman 1977; Artemieva & Mooney 2001). Above these isotherms, the lithosphere acts as an elastic body in post-seismic relaxation.

It is worth noting that diffusivity values in the range 200–400 $m^2\ s^{-1}$ have allowed satisfactory reproduction of the geodetic velocity field in the Anatolian-Aegean zone and the time pattern of the presumably induced seismicity in the Aegean zone, which followed the strong seismic sequences that have occurred along the North Anatolian fault since 1939 (Mantovani *et al.* 2001a; Cenni *et al.* 2002).

Rydelek & Sacks (1990) suggested that, at a given distance from the source zone, earthquake triggering is most probable when the induced strain is greater than 63 per cent of its maximum value. This alternative approach would not change significantly the results we have obtained by using strain rate peaks, as the time delays predicted by the two approaches are similar (Figs 4c and d).

Fig. 5 gives an idea about the evolution of the displacement and strain fields in the zone considered for our preferred model parameterisation ($D = 400\ m^2\ s^{-1}$, $L = 50$ km, $u = 1$ m). It can be noted that the induced strain in southern Apennines has a dominant E–W principal lengthening, which is coherent with the geometry of seismic sources in this zone (Fig. 2) and that the velocities that are expected in the central Mediterranean region roughly 20 yr after the 1979 Montenegro event (Fig. 4b), i.e. at present, have amplitudes of the same order of magnitude as those derived by geodetic measurements (Fig. 6). The last evidence suggests that the effects of post-seismic relaxation may have a considerable influence on the kinematic pattern of the study area up to tens of years after the triggering event. Thus, the geodynamic interpretations of GPS velocity fields that do not take into account such possible effects in zones affected by recent strong seismicity should be taken with caution.

Seismological observations and heat flow data indicate that the structural/rheological features of the crust–upper-mantle undergo a remarkable variation passing from the Adriatic continental plate to the Tyrrhenian oceanic domain (Baldi *et al.* 1982; Mantovani *et al.* 1985; Panza & Suhadolc 1990; Finetti *et al.* 2001; Martinez *et al.* 2000). Furthermore, tomographic investigations (e.g. Spakman *et al.* 1993; Lucente *et al.* 1999; Morelli & Piromallo 2001), study of high-frequency wave propagation in the southern Tyrrhenian region (Mele 1998) and reconstructions of the Plio-Quaternary evolution of the central Mediterranean area (e.g. Mantovani *et al.* 1997a, 2001b, 2002, and references therein) suggest the presence of subducted Adriatic lithosphere under the southern Apenninic belt and Calabrian Arc. Such a buried body could behave

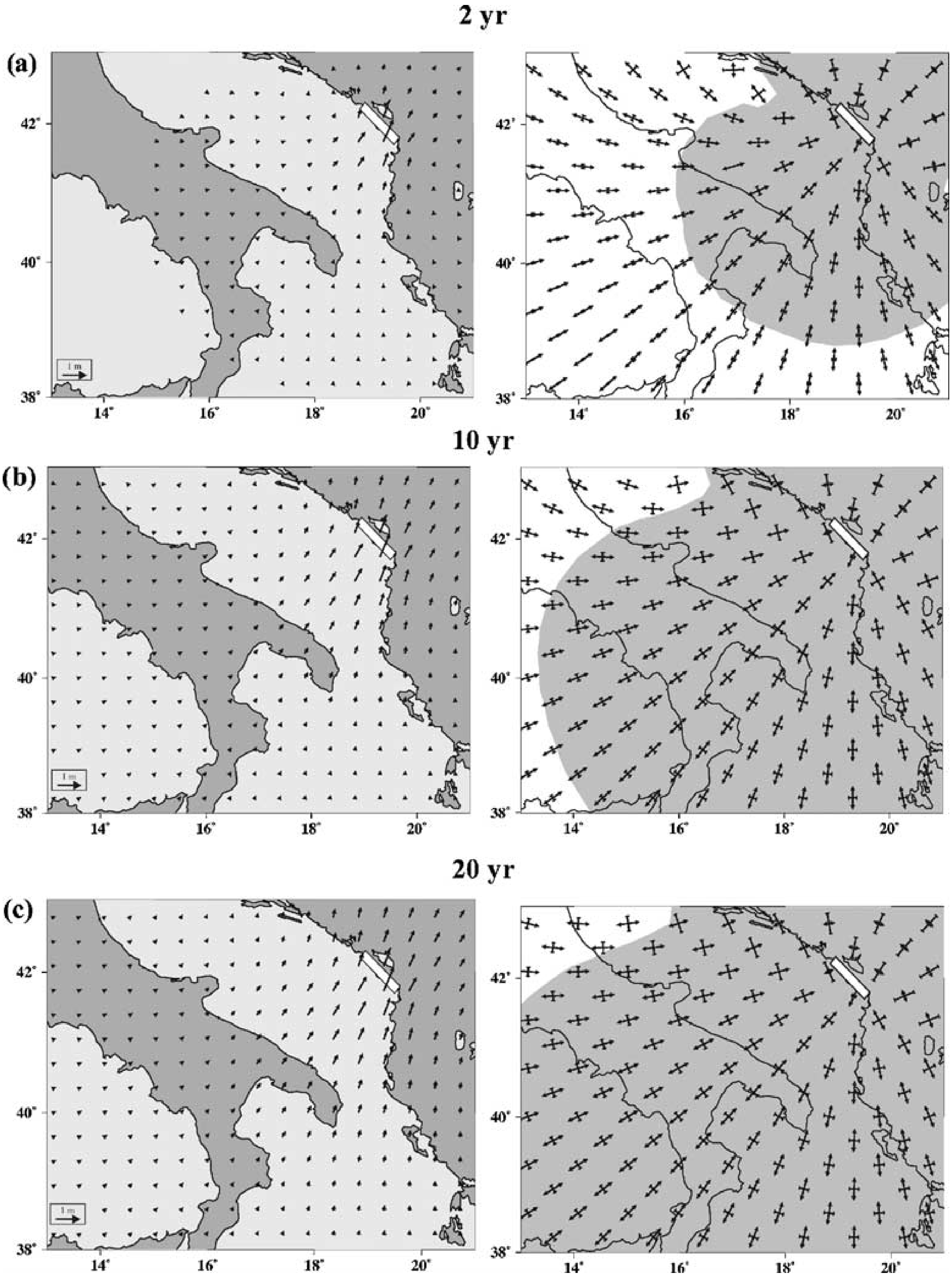


Figure 5. Displacement (left) and strain (right) fields, computed at three different times (2, 10 and 20 yr). Diffusivity, fault length and fault slip are 400 m^2 s^{-1}, 50 km and 1 m, respectively. The source area of the perturbing event (1979 Montenegro) is indicated by a white box. The principal axes of the strain tensor are reported; bars and diverging arrows respectively represent shortening and lengthening. For each point, the length of the two eigenvectors is normalized to the Euclidean norm of the local strain tensor. The shaded region encloses the points where strains overcome 10^{-7}, i.e. a value about an order magnitude greater than the earth tide strain.

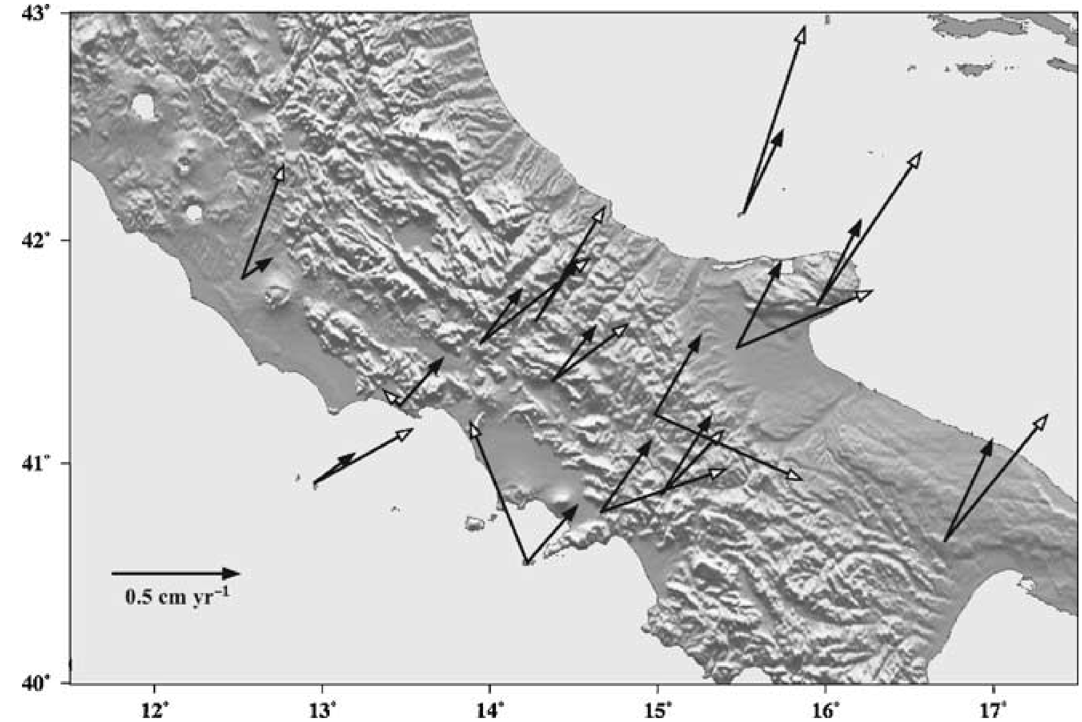


Figure 6. Comparison between geodetic (white arrow heads) and computed (black arrow heads) velocities in southern Italy in a number of GPS sites (Anzidei *et al.* 2001). Both velocities are averaged over the time interval 1994–1999. Diffusivity, fault length and fault slip adopted in computations are 400 m^2 s^{-1}, 50 km and 1 m, respectively.

as a sort of anchor, which can considerably reduce the mobility of the western Adriatic margin and thus of the belt lying above it, i.e. the southern Apennines. Since this last zone is the one where the correlation with Dinaric earthquakes is most significant, as indicated by statistical analysis (Mantovani & Albarello 1997), one could suppose that post-seismic relaxation processes are significantly influenced by the above structural heterogeneity. We have investigated this problem by numerical experiments, adopting a model where the slab's effect has been tentatively simulated by a very low value of diffusivity in the Apenninic domain (Fig. 7). This choice is based on the hypothesis that the slab could behave as an obstacle against the diffusion of stresses through the zone lying above it. The results obtained by this experiment, in which we have tentatively assumed a value of D equal to one hundredth of the best fit value (i.e. 4 m^2 s^{-1}) are displayed in Fig. 8. The comparison between the strain field shown in Fig. 8(a) and that related to the laterally homogeneous model (Fig. 5a) suggests that the influence of the Adriatic slab on relaxation processes in the Adriatic region could favour earthquake triggering in the southern Apennines. This possibility is suggested by two results of the above experiment. One is that a zone of roughly E–W pure lengthening appears in the southern Apennines (Fig. 8a), which should favour the activation of the normal fault systems observed in this zone. The other feature is that both strain and strain rate inside the low diffusivity zone (southern Apennines) are higher than those obtained with the homogeneous model (Fig. 8b).

5 DISCUSSION AND CONCLUSIONS

Post-seismic viscous relaxation induced in the Adriatic plate by Dinaric decoupling earthquakes might be responsible for the occurrence of tensional events in southern Italy, within a few years of the triggering event. Quantification, by finite element modelling, of the relaxation processes produced by a dislocation in the southern Dinarides compatible with the seismic slip of the strong 1979 Montenegro earthquake, has shown that the strain increase induced in the southern Apennines by such an event is significantly higher than that produced by earth tides and is compatible with fault geometries in the southern Apennines. The time delay (1.6 yr) between the Montenegro event and the presumably induced 1980 Irpinia earthquake could be accounted for by supposing that earthquake triggering was most likely when the strain rate peak reached the southern Apennines. The best match of this time delay, under the above condition, is obtained when a value of 400 m^2 s^{-1} is assumed for model diffusivity. This result is corroborated by the fact that comparable diffusivity values have allowed satisfactory reproduction of the geodetic velocity field and the time pattern of seismic activity in the western Anatolian-Aegean region, as effects of viscous relaxation triggered by the sequence of very strong earthquakes occurring along the North Anatolian fault system since 1939 (Mantovani *et al.* 2001a; Cenni *et al.* 2002). Another significant constraint on the actual value of *D* may derive from the fact that values lower than 50 or 100 m^2 s^{-1} (depending on the fault length) cannot account for the observed interrelation, as in the first 5 yr they imply strain amplitudes comparable to those expected from earth tides.

The best-fit diffusivity value (400 m^2 s^{-1}) implies that in the Adriatic region the viscosity of the mantle asthenosphere is about 2.5×10^{18} Pa s, if a seismological thickness is assumed for model stratification, while a crustal asthenosphere viscosity of about 10^{17} Pa s is implied if a rheological thickness is adopted. Since the viscosity related to seismological thickness is more consistent with independent estimates of η (see the earlier discussion) one may be induced to prefer such kinds of model for the quantification of stress

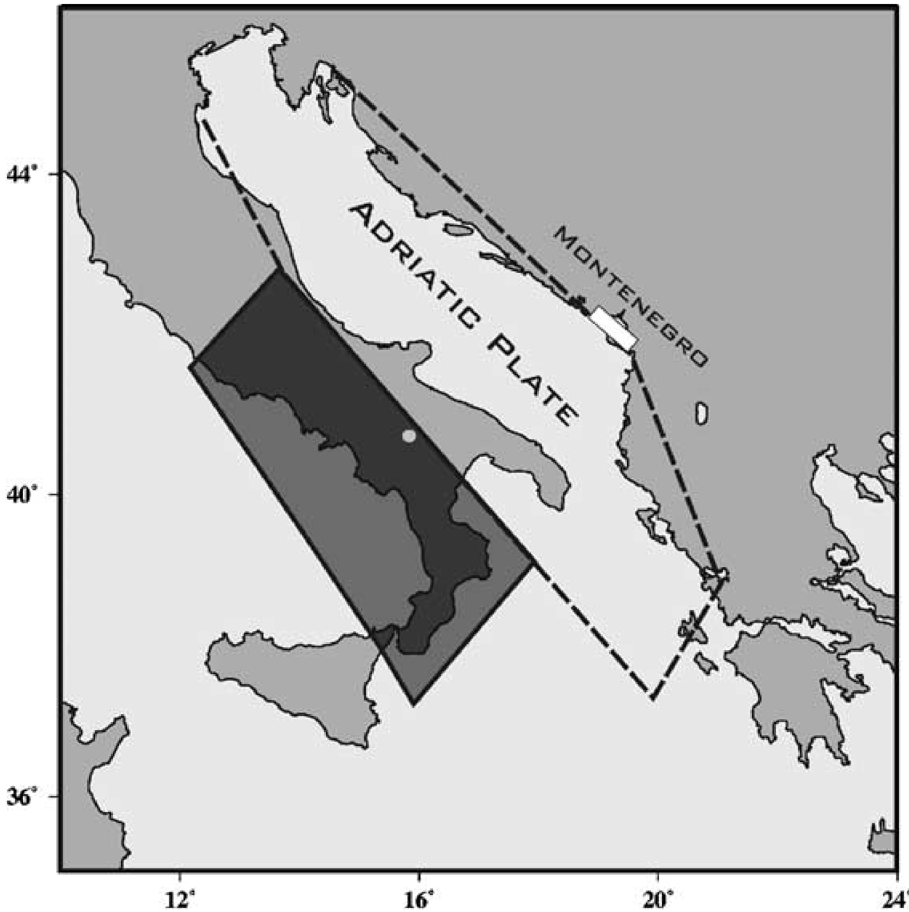


Figure 7. Model adopted for investigating the effects of lateral heterogeneity on viscous relaxation. Diffusivity is 400 $m^2\ s^{-1}$ everywhere, but in the southern Apennines (shaded area), where a very low value ($D = 4\ m^2\ s^{-1}$) is assumed to tentatively simulate the presence of subducted lithosphere under this zone (see text for further explanations). The rupture zone of the 1979 Montenegro earthquake (white box) is 50 km long. A sketch of the Adriatic plate boundaries is reported for reference (dashed line).

diffusion processes. However, several uncertainties still surround this problem.

Numerical experiments performed with a heterogeneous distribution of diffusivity in the model have shown that the presence of subducted lithosphere under the Apennines, tentatively simulated by assuming a very low value of diffusivity in that zone, can amplify the magnitude of the strain induced by viscous relaxation, with particular reference to the E–W principal lengthening. This effect could explain the nature of seismic sources of the major southern Apenninic earthquakes, which are mainly associated with dip-slip motion on normal faults parallel to the main Apenninic trend.

A major source of uncertainty in quantifying relaxation processes concerns the question of whether this phenomenon is controlled by a purely viscous asthenosphere, as we assumed for simplicity in our approach, or, instead, a viscoelastic nature must be adopted for this layer. Insights into this problem can tentatively be inferred from analytical investigations of post-seismic relaxation, even if performed with 1-D models (e.g. Savage & Prescott 1978; Albarello & Bonafede 1990). The results of these studies (see, e.g. Fig. 2 of Albarello & Bonafede 1990) show that the time pattern of strain related to viscous and viscoelastic asthenospheres are significantly different in the near field, i.e. for distances lower than 200–300 km, while they tend to assume similar shapes for larger distances, such as the one involved in the seismic interrelation considered here (Fig. 1). Thus, our earlier considerations about the possible influence of the peaked-like shape of the time pattern of strain rate on earthquake triggering would not be crucially affected by the presence of a viscoelastic asthenosphere. The main difference between viscous and viscoelastic relaxation concerns the very short-term behaviour, in that in the viscoelastic case an elastic strain immediately propagates in the whole model. However, possible effects of this phenomenon on long-range earthquake triggering, at least related to major events, are not observed in the study area.

Another basic and still unanswered question about earthquake triggering is whether the strain induced by post-seismic relaxation, in spite of being a small fraction of the earthquake stress drop (Harris 1998), may trigger large shocks. Possible explanations of this phenomenon could be found within the conceptual framework of self-organized criticality (SOC), which postulates that a deformable system, stressed by constant external forces, organizes itself in a way that there are parts of the system always near the failure threshold (e.g. Sornette *et al.* 1990; Grasso & Sornette 1998). Viewing plate tectonics and seismicity in the light of this theory permits the explanation of much of the observational evidence, such as the Gutenberg-Richter distribution of earthquakes, the power-law distribution of fault size, the fractal pattern of faulting and the sporadic occurrence of large shocks in areas far from plate boundaries (Bak & Tang 1989; Cowie *et al.* 1993; Main 1996). The SOC theory implies that even very small perturbations travelling into the system can lead to catastrophic evolution of metastable domains, as it

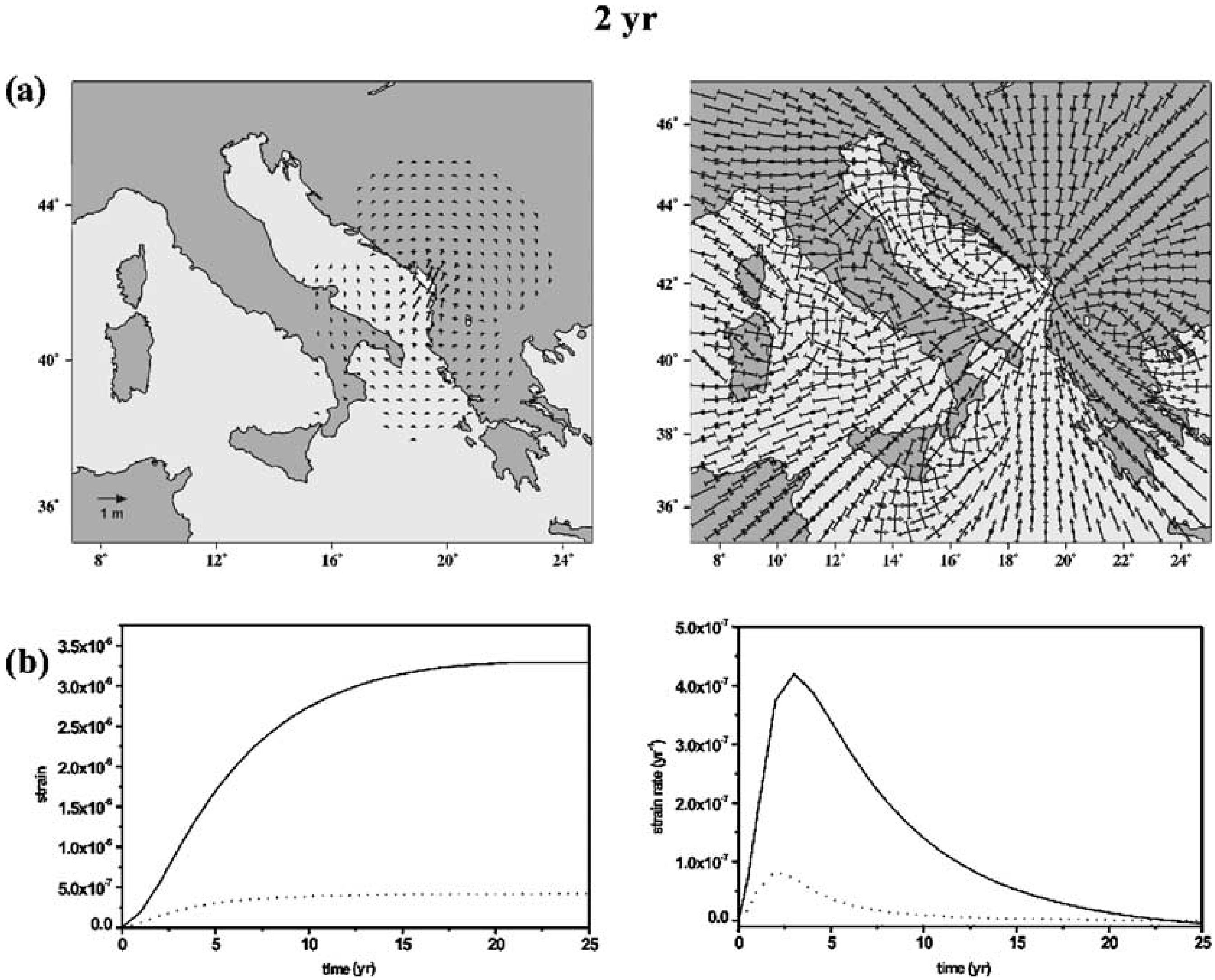


Figure 8. Effect of a lateral variation of diffusivity on the time pattern of post-seismic relaxation. (a) Displacement (left) and strain (right) fields computed at 2 yr, with the model described in Fig. 7. Symbols as in Fig. 5. With respect to the homogeneous model (Fig. 5), the tensional character of the strain regime is enhanced in the southern Apennines. (b) Comparison between the time patterns of strain (left) and strain rate (right) in the southern Apennines obtained with the model of Fig. 7 (full line) and the homogeneous model (dotted line). Both results have been computed at the white point indicated in Fig. 7.

occurs for avalanches, climatic changes and possibly earthquakes (Main 1996). Moreover, numerical modelling performed with cellular automata suggests that stress perturbations lower than 1 bar, i.e. of the same magnitude of those induced by viscous relaxation (10^4–10^5 Pa, or 0.1–1 bar for $E = 10^{11}$ Pa, see Fig. 4c), can trigger large seismic events in complex systems formed by many interacting elements, as fault-bounded blocks, loaded to the critical state by tectonic forces (Rydelek & Sacks 1999).

The possible role of small stress pulses on earthquake triggering has also been investigated by biaxial shortening tests carried out with pre-cut granite samples (Lockner & Beeler 1999). In these experiments, a periodical stress perturbation has been added to the imposed constant shortening rate, in order to simulate the behaviour of a fault subjected to a stress variation exceeding tectonic stress. The above study has shown that the occurrence of stick-slip events, representing earthquakes, is clearly influenced by the imposed perturbation when stress values overcome 10^4 Pa (0.1 bar), which is coherent with the evidence and arguments discussed above.

One should also consider that the stresses induced by post-seismic relaxation are comparable to or greater than static stress changes in the elastic earth, which are believed to be a plausible cause of the increase of seismicity rate near hypocentral zones (e.g. Okada 1992; King *et al.* 1994; Harris & Simpson 1996; Deng & Sykes 1997; Nalbant *et al.* 1998; Hardebeck *et al.* 1998). In this regard, however, one should be aware that the strain induced by viscous relaxation is higher when a model with a finite thickness of the elastic lithosphere is considered, with respect to models involving an elastic half-space, such as those adopted by the above mentioned authors.

As final remark, we would point out that the evidence and considerations reported in this work may support the hypothesis (Rydelek & Sacks 1999) that, at present, transient perturbations induced by viscous relaxation seem to be the most adequate physical mechanism for explaining delayed earthquake triggering at large distance.

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