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PLATE DEFORMATION

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PLATE DEFORMATION

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Introduction

The task at hand will be to evaluate vertical and horizontal displacements and velocities by appropriate geophysical Earth models and compare them against GPS (global positioning system) data to understand plate deformation mode. At the turn of the third millennium this class of geodetic data, from internationally recognized geodetic networks, GPS campaigns in selected local areas of geophysical interest and DInSAR (differential interferometry-synthetic aperture radar) analyses, represents a powerful tool geophysicists have at their disposal to unravel the characteristics of plate deformation mode and plate physical properties by comparison to geophysical modelling predictions. In effect, given that, such a comparison can be carried out at various wavelengths, or spatial dimensions, over the globe, from continental to regional or local scales ranging from 10^3 to $1 - 10^2$ km

Vertical deformation properties of the lithosphere will be exploited by considering the Earth's global response to the redistribution of surface loads during ice-sheet disintegration in a relatively short time scale of the order to 10 yr after comparison to long lasting tectonic processes, typically of the order to 10^3 yr. This process, called post glacial rebound (PGR), is responsible for the largest vertical deformation rates over the Earth's surface, of the order of 10 mm/yr, and is thus the most interesting geophysical phenomenon to exploit in order to appreciate the planet's characteristic vertical deformation when subjected to time-varying surface loads.

Horizontal deformation properties of the lithosphere at the global scale will be analyzed by considering the response of the lithosphere to continental collision, a long lasting tectonic process that can accommodate relative plate velocities of the order of centimetres per year. Deformation properties of the lithosphere at the regional or local scale will be exploited by considering the response of the crust-lithosphere system to earthquakes or to localized surface loads, such as present-day shrinkage of mountain glaciers.

The following sections are organized from global to regional and local geophysical phenomena, or from long to short lithosphere deformation wavelengths. PGR and continental collision are thus considered first and co- and post-seismic deformation in active seismogenic zones afterwards. The former will thus be analyzed both in terms of their mathematical modelling and their signature in geodetic deformation patterns, and the latter retrieved mainly from GPS networks in Europe and the Mediterranean. This part of the globe has been chosen since it can be considered a sort of natural laboratory where all these geophysical phenomena are active today, from PGR in Fennoscandia and its periphery, including central continental Europe and the Mediteranean, to continental collision due to the Africa- Eurasia convergence and co- and post-seismic deformation in active seismogenic zones. The latter are in fact found throughout the plate boundary separating the African and Eurasian domains, running from Gibraltar in the west to Anatolia in the east through North Africa, the Italian peninsula and the Aegean.

Only for co-seismic deformation are DInSAR results considered since they release a continuous deformation pattern and provide extra information with respect to GPS data, which are sparse over sites where GPS antennae are located. In fact, the pixels in DInSAR analysis are distributed with a spatial resolution of tens of meters, wherein the differential interferometric phase is recorded from two or more images so as to yield a sort of continuous deformation pattern of the Earth.

Focusing on plate-deformation properties will thus provide a close look at the lithosphere and crust. Indeed, it should be noted that, when studying PGR, the Earth's outermost elastic layer cannot be isolated from the mantle since a great deal of the latter's material is required to fill mass deficiency in deglaciated areas during the process of isostatic readjustment. This meas that equations of motion must necessarily be applied to the whole spherical volume, including the lithosphere and underlying mantle. When studying continental collision, which involves only the outermost part of the planet, the mantle can be considered as totally decoupled from the lithosphere. In this case, to a good approximation, the spherical volume considered in PGR can thus be shrunk to the outermost lithosphere, where differential equations of motion are solved within a thin outer layer. This concept is the basis of the mathematical developments for PGR and continental collision, as it will be shown.

In modelling geophysical phenomena such as plate deformation, the choice of the rheology, or the way the Earth's material behaves under an applied deviatoric stress, depends essentially on the characteristic time scale over which such deformation occurs. In what follows, the main questions at issue in attempting to study transient and long time-scale geodynamic phenomena are addressed in a wide arc of time scales. These range from seconds, characteristic of co-seismic deformation, to hundreds of millions of years, as in the case of subduction, making use of results based on the analytical normal mode theory in viscoelasticity or finite element schemes.

Figure 1.1 sketches out the entire geodynamic spectrum spanning the whole range of phenomenological time scales. One of the key questions is whether one can implement a constitutive law that can satisfactorily model all these phenomena, from the anelastic transient regime to the steady-state domain. Analysing plate deformation thus requires the implementation of the appropriate constitutive relation between stress and strain, which is capable of reproducing the mechanical behaviour of the crust-lithosphere system over the widest range of time scales characterizing the geodynamical process under study.

The appropriate constitutive relation which is to be employed in analyzing transient geodynamic phenomena like PGR, is currently a matter of controversy in geophysics.

There is evidence that at the stress levels in PGR (less than 0(10 bar)) the creep mechanism may be linear for polycrystalline aggregates (Relandeau, 1981) since grain boundary processes such as Coble creep may become dominant.

Since there is no unambiguous evidence in either the postglacial rebound event or in other types of geodynamic data which absolutely requires a nonlinear viscoelastic rheology, geophysicists tend to prefer the simple linear models in viscoelasticity, which allow for a considerably simpler mathematical treatment of the dynamics.

The simplest viscoelastic model which can describe the Earth as an elastic body for short time scales and as a viscous fluid for time scales characteristic of plate deformation during continental collision or subduction, is that of the linear Maxwell solid, portrayed in Figure 1.2. This figure shows a standard one-dimensional spring and dashpot analog of the Maxwell rheology. The speed for shear wave propagation depends on the square root of the instantaneous rigidity μ_1, whereas the strength of the lithosphere and mantle, over time scales of $10^3 - 10^6$ yr, depends on the viscosity ν_1, characterizing the ability of the planet to propagate deformation over those long time-scales.

In order to gain a deep insight into the physics of plate deformation processes, the mathematical equations which describe the mechanical behaviour of the whole set of geophysical phenomena are implemented, from PGR to plate behaviour in continental collision and subduction.

POST GLACIAL REBOUND

Introduction

From oxygen isotope analysis isotopes of ocean sediments , Shackleton and Opdyke (1976) have shown that great ice ages repeatidily occurred over the Earth, characterized by a growth period of about 90,000 years and a decay period of about 10,000 years, leading to a total period of about 100,000 years for one complete cycle. These findings are based on the relative concentration of the oxigen isotopes ^{16}O and ^{18}O contained in water. Depending on the amount of ice that has accumulated on land, the relative abundance of the two isotopes, incorporated in ocean sediments, vary due to their specific mass difference. Since the lighter isotope evaporates more easily from the oceans, giving an over-abundance of this isotope in ice with respect to ocean water, an over-abundance of the heavier isotope is created in the oceans and recorded in the sediments during glacial times.

The mathematical simulation, by means of the appropriate geophysical models, of the response of the Earth to this huge redistribution of ice and water masses, occurring over time scales of 10^3 - 10^4 yr, makes it possible to retrieve the mechanical and deformation properties of the lithosphere, the outermost portion of our planet, including the underlying mantle. In modelling PGR, the plate under consideration is thus the entire outermost, elastic shell of the Earth, rather than a part of it as in continental collision or subduction, as it will become clear in the following. PGR, being such a huge, global readjustment of the Earth to surface loads, involves mainly the deformation properties of the mantle, rather than the outermost plate, but the latter certainly plays the role of fine-tuning the surface, visible effects of the rebound on GPS data, as shown in the following for PGR modelling.

PGR is responsible for the largest vertical motions occurring over the Earth, such as those occurring in Hudson Bay or in the Gulf of Bothnia, and the largest global gravity changes, as those nowadays seen from satellite data.

PGR mathematical modelling

For long time scale processes, as PGR, the inertial forces vanish, and conservation of linear momentum requires that the body forces $\mathbf{F}$ per unit mass acting on the element of the body are balanced by the stresses that act on the surface of the element. The stress tensor σ acting on the infinitesimal block with density ρ, satisfies the following momentum equation

$$\nabla \cdot \sigma + \rho \mathbf{F} = 0. \tag{1}$$

at each instant of time. As a start, the Earth is assumed compressible, laterally homogeneous, radially stratified, hydrostatically pre-stressed and not rotating. PGR models shown afterwards will be based on the assumption that Earth's material is incompressible, since in this case solutions to equation (1) can be treated analytically, as detailed in Sabadini and Vermeersen (2004).

Equation (1) holds in each volume element of the Earth shown in Figure 3. Elastic equation of motion are considered, since any linear viscoelastic problem, which is of interest to us, is equivalent to an elastic problem in the Laplace domain (see Sabadini and Vermeersen, 2004 for details). Momentum and gravity equations are thus solved for an elastic medium, and only at the end, once the elastic solution has been obtained, the solution is derived in the time domain by anti-transformation from Laplace to time domain..

The stress tensor σ is the sum of the initial pressure, due to the hydrostatically prestressed conditions, plus a perturbation σ_1, so that σ reads

$$\sigma = \sigma_1 - p_0 \mathbf{I}. \tag{2}$$

where σ_1 denotes a tensor which describes the acquired, non-hydrostatic stress, related to the strain by means of the appropriate constitutive equations, describing from the mathematical point of view, the mechanical analog depicted in Figure 2. The hydrostatic pressure p_0, with $\mathbf{I}$ the identity matrix, enters the equation above with the minus sign since it denotes a compressive stress, which is negative according to the convention that stresses are positive when they act in the same direction as the outward normal to the surface embedding the Earth's volume under study. On the elementary surface enclosing the elementary volume in which the equation of equilibrium holds, the pressure, or stress due to the load of the overlying material, is negative. According to this convention. The equation of conservation of linear momentum thus becomes

$$\nabla \cdot \sigma_1 - \nabla p_0 + \rho \mathbf{F} = 0. \tag{3}$$

If the body is subject to an elastic displacement $\mathbf{u}$ in t_0, the pressure in $t_0 + \delta t$ at a fixed point in space is given by

$$p_0(t_0 + \delta t) = p_0(t_0) - \mathbf{u} \cdot \nabla p_0. \tag{4}$$

The minus sign accounts for the fact that the pressure increases at a fixed point in space if the elastic displacement occurs in the opposite direction with respect to the pressure gradient.

The equation of conservation of linear momentum after the elastic displacement becomes, with $p_0(t_0 + \delta t)$ instead of $p_0(t_0)$,

$$\nabla \cdot \sigma_1 - \nabla p_0(t_0) + \nabla(\mathbf{u} \cdot \nabla p_0) + \rho \mathbf{F} = 0. \tag{5}$$

The gradient of the initial pressure is given by

$$\nabla p_0 = -\rho_0 g \hat{\mathbf{e}}_r, \tag{6}$$

where $\hat{\mathbf{e}}_r$ denotes the unit vector, positive outward from the Earth center. Subsituting equation (6) into equation (5), the equation of equilibrium becomes

$$\nabla \cdot \sigma_1 - \nabla p_0(t_0) - \nabla(\rho_0 g \mathbf{u} \cdot \hat{\mathbf{e}}_r) + \rho \mathbf{F} = 0. \tag{7}$$

The force $\mathbf{F}$ can generally be split into gravity and all kinds of other terms that represent the forcing due to a variety of geophysical phenomena, such as tidal forces, loads due to ice-sheet disintegration and ice-water redistribution, earthquake and so on.

For the moment, it is assumed that the force $\mathbf{F}$ is gravity, thus simulating the condition of a free, self-gravitating Earth with no other forcings or loads acting on its surface or interior, and that, being a conservative force, it can be expressed as the negative gradient of the potential field ϕ

$$\mathbf{F} = -\nabla\phi. \tag{8}$$

The potential field ϕ can be written as the sum of two terms

$$\phi = \phi_0 + \phi_1, \tag{9}$$

with ϕ_0 as the field in the initial state and ϕ_1 the infinitesimal perturbation.

The linearized equation of momentum becomes, with ρ_1 the perturbation in the density and g the gravity,

$$\nabla \cdot \sigma_1 - \nabla(\rho_0 g \mathbf{u} \cdot \hat{\mathbf{e}}_r) - \rho_0 \nabla\phi_1 - \rho_1 g \hat{\mathbf{e}}_r = 0, \tag{10}$$

the second term of equation (1.7) being canceled by the term $-\rho_0\nabla\phi_0$.

The first term of equation (1.10) describes the stress contribution, the second term the advection of the hydrostatic pre-stress, the third term the changed gravity (self-gravitation) and the fourth term the changed density due to compressibility. In cases where self-gravitation is neglected, the third term will be zero, while in the case of incompressibility, as considered herefter, the fourth term will be zero.

The perturbed gravitational potential ϕ_1 satisfies the Poisson equation

$$\nabla^2\phi_1 = 4\pi G\rho_1, \tag{11}$$

with G as the universal gravitational constant. In the case of incompressibility the right-hand term will be zero since $\rho_1 = 0$, and equation (1.11) reduces to the Laplace equation

$$\nabla^2 \phi_1 = 0. \tag{12}$$

The equations above need to be supplemented by a constitutive equation describing how stress and strain (or strain-rate) are related to each other. For PGR modelling, the Maxwell rheological model depicted in Figure 1.2 is considered. The momentum and Poisson equations for an elastic solid will first be expanded in spherical harmonics and only afterwards transformation from the Laplace to the time domain is performed in order to retrieve the viscoelastic solution, specialized for an incompressible material in PGR results shown afterwards.

The equilibrium and Poisson equations can be written in spherical coordinates, as shown in Schubert *et al.* (2001), page 281.

By denoting derivatives with respect to r and θ by means of ∂_r and ∂_θ, and assuming that there is no longitudinal component in the fields or in their derivatives, with symmetric deformation around the polar axis as appropriate for a point load loading the north pole of the Earth's model of Figure 2, and taking account of the continuity equation written as follows

$$\rho_1 = -\nabla \cdot (\rho_0 \mathbf{u}) = -\mathbf{u} \cdot \hat{\mathbf{e}}_r \, \partial_r \rho_0 - \rho_0 \nabla \cdot \mathbf{u}, \tag{13}$$

with $\Delta = \nabla \cdot \mathbf{u}$ and ρ_1 denoting the perturbed density, the two r and θ components of the momentum and Poisson equations become

$$\begin{aligned} &-\rho_0 \partial_r \phi_1 + \rho_0 g_0 \Delta - \rho_0 \partial_r (u g_0) + \partial_r \sigma_{rr} \\ &+ r^{-1} \partial_\theta \sigma_{r\theta} + r^{-1} (2\sigma_{rr} - \sigma_{\theta\theta} - \sigma_{\phi\phi} + \sigma_{r\theta} \cot\theta) = 0 \end{aligned} \tag{14}$$

$$\begin{aligned} &-\rho_0 r^{-1} \partial_\theta \phi_1 - \rho_0 g_0 r^{-1} \partial_\theta u \\ &+ \partial_r \sigma_{r\theta} + r^{-1} \partial_\theta \sigma_{\theta\theta} + r^{-1} ((\sigma_{\theta\theta} - \sigma_{\phi\phi}) \cot\theta + 3\sigma_{r\theta}) = 0 \end{aligned} \tag{15}$$

$$r^{-2}\partial_r(r^2\partial_r\phi_1) + (r^2\sin\theta)^{-1}\partial_\theta(\sin\theta\partial_\theta\phi_1) = -4\pi G(\rho_0\Delta + u\partial_r\rho_0), \tag{16}$$

where σ_{rr}, $\sigma_{\theta\theta}$, $\sigma_{\phi\phi}$ and $\sigma_{r\theta}$ denote the stress components in spherical coordinates.

In principle, deformation, stress field and gravity field can be solved by means of numerical integration techniques. Sabadini and Vermeersen (2004) show how it is possible to solve these equations analytically by means of normal model modelling in the Laplace-transformed domain, as stated by the Correspondence Principle, providing deep insights into the physics of relaxation processes.

Momentum and Poisson equations are expanded in spherical harmonics, defined by

$$Y_l^m(\theta,\phi) = (-1)^m P_l^m(cos\theta)\exp(im\phi), \tag{17}$$

with $l = 0, 1, 2, ...$ and $m = -l, -l+1, ...l$ and $P_l^m(cos\theta)$ denoting the associated Legendre function

$$P_l^m(cos\theta) = \frac{(1-cos^2\theta)^{m/2}}{2^l l!}\frac{d^{l+m}}{d(cos\theta)^{l+m}}(cos^2\theta - 1)^l. \tag{18}$$

For the spheroidal part, of importance in PGR problems, the radial displacement u, the tangential (colatitudinal) component v, and the perturbation in the gravitational potential ϕ_1 as functions of the scalars U_l, V_l and ϕ_l, which depend solely on the harmonic degree l and on the radial distance r from the center of the Earth

$$u = \sum_{l=0}^{\infty} U_l(r)P_l^m(\cos\theta), \tag{19}$$

$$v = \sum_{l=0}^{\infty} V_l(r)\partial_\theta P_l(\cos\theta), \tag{20}$$

$$\phi_1 = -\sum_{l=0}^{\infty} \phi_l(r)P_l(\cos\theta), \tag{21}$$

where the Legendre polynomial $P_l(cos\theta)$ is obtained from the Rodrigues' formula

$$P_l(cos\theta) = \frac{1}{2^l l!}\frac{d^l}{d(cos\theta)^l}(cos^2\theta - 1)^l, \tag{22}$$

or from $m = 0$ in the previous definition (1.29) of the associated Legendre function. Note that the spheroidal solution does not carry any longitudinal displacement, as anticipated above.

The toroidal displacement components v' along colatitude and w along longitude, entering plate deformation processes activated by earthquakes, are defined by the following expansion in spherical harmonics

$$v' = \sum_{l=0}^{\infty} W_l(r) \sum_{m=-l}^{l} \nabla_\phi Y_l^m(\theta, \phi) \tag{23}$$

$$w = -\sum_{l=0}^{\infty} W_l(r) \sum_{m=-l}^{m=l} \nabla_\theta Y_l^m(\theta, \phi), \tag{24}$$

which provide the toroidal components of the latitudinal and longitudinal displacements v' and w as a function of the W_l scalar harmonic coefficients. Collectively, U_l, V_l, W_l and ϕ_l are named scalar eigenfunctions and satisfy linear systems of homogeneous ordinary differential equations in the radial variable r.

Spheroidal deformations belong to the same cetegory of those that the reader can produce in a football baloon by pressing it with his finger, while toroidal deformation are those produced by turning around, in opposite direction, the two hemispheres of the baloon.

The Correspondence Principle is based on the result that in the Laplace domain, viscoelastic equations are formaly identical to elastic ones, due to relation between strain-rate and stress in time-domain, that for and incomrressible Maxwell viscoelastic solid reads

$$\frac{d\epsilon}{dt} = \frac{\sigma}{2\nu} + \frac{1}{2\mu}\frac{d\sigma}{dt}, \tag{25}$$

with ν being the viscosity of the dashpot of the mantle or lower crust, for the case of the Earth. Laplace transformation of the equation above leads to

$$\tilde{\sigma}_{ij}(s) = 2\tilde{\mu}(s)\tilde{\epsilon}_{ij}(s), \tag{26}$$

with the Laplace-transformed digidity $\mu(s)$ being

$$\tilde{\mu}(s) = \frac{\mu s}{s + \mu/\nu}. \tag{27}$$

formally identical to the elastic stress-strain relationship.

This result indicats that solution to the time-dependent viscoelastic problem is obtained first in the Laplace transform domain as an equivalent elastic solution.

This elastic solution in the Laplace transform domain is then transformed back into time domain. It should be reminded that this approach is valid only for linear viscoelastic problems, for which the linear relation providd by equation (xx), in which μ do not depend on stress or strain, can be established.

In the Laplace transform domain, viscoelastic normal mode theory, detailed in Sabadini and Vermeersen (2004), provides the tools to obtain the time-dependent, viscoelastic solution.

The core of that theory stands on the result that in the Laplace transform domain, the solution for the stress-free boundary conditions at the Earth's surface, is singular at points of the Laplace variable s-axis which correspond to real, negative numbers. For an incompressible Earth's model, these numbers represent a finite ensamble of M real inverse relaxtin times s_j, with M dpending on the numbers of viscoelastic layers in which the Earth's model itself is stratified. For compressible modela, this ensamble is denuberably infinite (Vermeersen et al.,1997).

These singularities of the soltion in the s-domain are first-order poles, so that each singularity s_j provides a relaxing exponential $e^{s_j t}$, once anti-transformed in time-domain.

These s_j numbers are named viscoelastic relaxation modes, M being the total numbers of modes; the viscoelastic solution in time-domain becomes, for the the radial displacement pertaining to this chapter,

$$U_l(t) = U_{le}\delta(t) + \sum_{j=1}^{M} U_{lj}e^{s_j t}, \tag{28}$$

where l denotes the harmonic degree l, U_{le} the elastic conribution from the spring of Figure 2 and $sum_{j=1}^{M} U_{lj}e^{s_j t}$ the summation over the modes the (viscoelastic) conribution from the dashpot, coupled with the spring.

Similar expression hold for the horizintal displacement and perturbation in the gravitational potential V_l, ϕ_l, also of interest in theis chapter, .

For the simple four layer Earth's model, as sketched in Figure 3, it can be seen, as explained in Sabadini andVermeersen (2004), that eight modes are excited by density and rigidity contrasts at the various interfaces separating the lithosphere from the atmosphere, the mantle from the core, the lithosphere from the mantle and the upper mantle from the lower mantle.

These relaxation modes, which ultimately constrain the characteristic deformation time-scales of the Earth's model forced by any kind of geophysical processes, are plotted as a function of the harmonic degree l, entering the series given by equations (x),

They span time-scales from hundred to million years, with the labelling referring to the interface trigering that mode, with M denoting the mode excited by the density contrast between the Earth and the atmosphere, L that due to the rheloical contrastbetween teh elastic ouer lithosphere and underying viscolastic mantle, C that due to the density contrast betwen the core and the mantle, M1 that due to the density contrast between the upper and lower mantle, and T denoting fast relaxing mode sud to discontinuities in viscosity and rigidities at the various interfaces. These mdodes are those of the Earth's model used in Barletta and Sabadini (2006) to simulate upwelling of superplumes from the mantle, hitting the bottom of the lithosphere.

Global vertical and horizontal displacements from PGR

PGR responsible not only for vertical and horizontal displacements at the Earth's surface, but also for the secular component of the change of sea level and Earth's gravity field. These variations can also be affected by present-day mass instability in Antarctica and Greenland.

For PGR analyses, the multilayered,spherically stratified self-gravitating relaxation model, outlined in *Vermeersen and Sabadini (1997)* and based on the normal mode relaxation theory described above, is used. The redistribution of the glacial melt water on the viscoelastic Earth is solved within the ICE-3G model, by using the spectral analysis first implemented by *Mitrovica and Peltier (1991)*, appropriate for sea-level change calculations, as shown afterwards.

The results shown in Figure 4, portraying the modelled global deformation pattern induced by PGR, show that vertical and horizontal velocities associated with PGR are sensitive to the rheological or viscosiy stratification of the mantle, where the viscosity denotes the flow properties of the mantle, as depicted by the ν parameter in the dashpot of Figure 2.

The two panels of Figure 3, panel a, corresponds in fact to an upper and lower mantle viscosity of $\mu_{UM} = 0.5 \times 10^{21}$ Pa s and a lower mantle viscosity $\mu_{LM} = 1.0 \times 10^{21}$ Pa s, defining the PGR-21 model; panel b coresponds to $\mu_{UM} = 0.5 \times 10^{21}$ Pa s and $\mu_{LM} = 1.0 \times 10^{22}$ Pa s, defining the PGR-22 model. The elastic lithosphere is 80 km thick and both lithosphre an mantle are assumed incompressible, as stated above, which means that density remains constant within each layer of the radially stratified Earth's model and that, at a fixed position in space, density changes can occur only via displacements of interfaces separating material with different density. Both PGR-21 and PGR-22 are built as the Earth sketched in Figure 3, but each layer representing the lithosphere, the upper mantle and the lower mantle is in turn subdivided into thinner layers, in such a way that, taken collectively, these parts on the planet contains 31 layers, whose physical properties are volumetricall averaged from realistic, seismologically retrieved Earth's stratification (Anderson and Dziewonski, 1979).

The vertical deformation, represented by the colours, is characterized by uplifting centers over deglaciated areas in North America, Northern Europe and Antarctica, where ice-sheet complexes where located, in agreement with ICE-3G, and by subsidence in the periphery of these deglaciation centers. It should be reminded that these global deformation patterns represent, in terms of vertical and horizontal velocities in millimeter per year, the mathematical simulation of present-day deformation of the Earth, forced by ice-sheet disintegration during the Pleistocene, which ended abour 7000 yr ago, and that is going on today because of the viscous memory of the Earth, as scketched by the dashpot of Figure 2.

The horizontal velocity field is characterized by two different components. A global one, directed northward or southward with respect to the equatorial region of the Earth and due to the suction effect of the mantle material toward the deglaciated regions of the northern and southern hemispheres, and a regional one, directed radially and outward from the center of the different deglaciation zones. The relative strengths of these components, as the intensity of vertical motions, depend both on the viscosity ratio between the upper and the lower mantle, as comparison between panel b and panel a shows.

Increasing the viscosity in the lower mantle to 10^{22} Pa s, with respect to 10^{21} Pa s, makes mantle material more difficult to relax after deglaciation, which maintains larger horizontal and vertical velocities for present-day situation, as portrayed by Figure 4.

For lower mantle viscosity 10^{21} Pa s, it is remarkable that the outward horizontal velocity from deglacition centers is larger than the global velocity due to material flow from the equatorial region of the mantle, particularly visible in north America. Increasing the lower mantle viscosity makes the global flow from the equator larger, as anticipatd above, which in turn dampens, at deglaciation centers, the outward velocity directed toward the equator. This effect can be better visualized in the following Figure 4, focussing on PGR velocity patterns in Europe.

Figure 4 is an enlargment of the Fennoscandia region from Figure 4, showing vertical and horizontal velocity fields in Europe for the two Earth's models introduced previously in Figure 4, namely PGR-21 in panel a and PGR-22 in panel b. Two different components of the horizontal velocity field can be distinguished, a global north trending one, due to the suction effect of mantle material from the equator towards the deglaciated regions of the northern hemisphere, as anticipate above, and a regional one, directed radially and outward from the centre of deglaciation in the Gulf of Bothnia. Each component prevails over the other depending on the viscosity ratio between the upper and lower mantle. For PGR-21, panel (a), the two deformation styles in the deglaciated region and in the far field are well separated, as indicated by the outward horizontal velocities of at most 0.6 mm/yr in proximity of the deglaciated region, and by the north trending lower values, less than 0.2 mm/yr, in the far field.

The vertical deformation is characterized by two uplifting centres in the north eastern region where ice-sheet complexes were located. The Mediterranen region is affected by a subsidence of -1 mm/yr, with the adjacent European continental region essentially unaffected by vertical motions. The deformation pattern portrayed in panel (a) agrees with the findings of *Mitrovica et al.* [1994] (their Figure 3), except for the slightly smaller rates due to the overall reduction in the upper and lower mantle viscosity by a factor two in this analysis.

For PGR-22, panel (b), the high north trending component of the horizontal velocity, exceeding the local outward velocity in the southern part of the deglaciation region, is due to the global larger isostatic disequilibrium of the planet with respect to

PGR-21, caused by viscosity increase. Due to the latter, a substantial increase in the horizontal velocity is obtained, up to 2.4 mm/yr in the north and 1.2-1.6 mm/yr in the far field. In comparison with panel (a), a substantial intensification of the uplift is noticed, affecting a wider region. Subsidence increases in the Mediterranean region to 2 mm/yr is obtained, extending its influence also in Central Europe, at a rate of 1 mm/yr.

In general, these PGR patterns of 3-D deformation predicted by models have shown to be relatively consistent among the various analyses[*James and Lambert*, 1993; *Mitrovica et al.*, 1993, 1994b; *Peltier*, 1998].

PGR in GPS analyses

In order to help the reader to appreciate the quality of PGR geophysical modelling, and thus of its ability to correctly reproduce the physical properties of the lithosphere-mantle system and the characteristics of the deformation pattern at the Earth's surface, comparison between PGR predictions and two classes of geodetic observables retrieved from GPS data analyses, rates of change of baselines and strain-rates, is carried out. The first observable, shortened into baseline rate, denotes the rate of change of the distance, or baseline, between to GPS sites, along the direction connecting the two sites; in the following, only the style of this change, extension or equivalently lenghtening of the baseline, and compression or equivalently shortening of the baseline, will be considered, leaving to the analysis of strain-rate patterns, within triangular domains connecting three GPS sites, a detailed analysis of the deformation.

Baseline analyses thus provides a one dimensional picture of the deformation, between two GPS sites, while strain-rate analysis portrays a two dimensional picture, within the area surrouded by the three segments connecting the three GPS sites. The latter analysis thus provides a precise picture of the deformation style of the area under study.

The region has been monitored by surveying using permanent GPS receivers of the ITRF2000 network, established by the International Earth Rotation Service [IERS; *Altamini et al.*, 2002]. BIFROST data provide additional stations not included in the ITRF network [*Johansson et al.*, 2002; *Milne et al.*, 2001]. The reader should be aware that sites quoted as BIFROST sites in the following figures, include, in addition to sites in the actual BIFROST network, a set of 5 other sites, namely HERS, MADR, BRUS, KOSG, POTS, WETT, RIGA, not officially belonging to this network..

Consistency between model predictions and geodetic deformation patterns can thus be viewed as indicator that plate deformation properties have been properly accounted for, in terms of those parameters that are particularly relevant for the present analyses, such as the thickness of plate or its elasticity or viscosity, or its coupling with the underlying mantle

Baseline rates obtained from geodeticaly rerieved data are first shown with respect to the reference site VAAS. Figs. 6a-b show, in fact, the location of baselines associated with ITRF2000 and BIFROST data sets, respectively, where the observed dominant extension is denoted by blue and the observed shortening by red. Both panels indicate a well defined extensional pattern, for all baselines, independently on their length, except for a few of them, indicating compression.

Fig. 7a shows PGR predictions, based on the PGR compressible model detailed in Milne et al. (2001), and used also in Marotta et al. (2004), providing the style of changes in baseline rate for all VAAS-referenced baselines in Fig. 6, blue indicating extension while red indicating shortening. Except for some inconsistencies with a few southerly directed baselines, PGR modelling well reproduces the major feature of Fig. 6, which is the dominant extension in the ITRF2000 and BIFROST data.

Since post-glacial adjustment in Fennoscandia is characterized by horizontal motions directed outward from the center of the ancient ice complex, in proximity of VAAS, as shown previously in Figure 4, widespread lenghtening of baselines or extension along the short BIFROST baselines Fig. 6b is expected. On the other hand, the origin of the widespread extension for the longer ITRF2000 or BIFROST baselines, extending to central and southern Europe, visible in Figs. 6a-b, is not as clear as lengthening along short baselines. Explanation for this behaviour can be found in Marota et al. 2004 and is repeated herein. For the PGR-21 model, for example, carrying a viscosty profile similar to that defining the Milne et al 2001 model considered in Figure 7, horizontal velocity field is characterized by outward-directed motions within Fennoscandia, changing to motions toward Fennoscandia at the periphery, as shown in Figure 4a in detail. On this basis, it could be expected that PGR would induce shortening in the longer baselines, VAAS to central or southern Europe, in Fig. 6a. It should now be reminded, on the other hand, the both horizontal and radial motions contribute to changes in baseline length. From Fig. 4a VAAS is predicted in uplift at a rate close to 1 cm/yr, while central and southern European sites, which lie within the peripheral bulge of Fennoscandia, are subsiding at lower rates. The net contribution of this uplift and moderate subsidence is to extend the baselines, thus counteracting the baseline shortening associated with the PGR-induced horizontal velocity field. The net effect of baseline lengthening is consistent with the pattern of widespread extension evident in the longer baselines in Fig. 6a.

For ITRF2000 and BIFROST data sets, Fig. 9 shows the pattern of baseline rates referenced to Potsdam (POTS) in northern Europe. Short BIFROST baselines defined by sites between $55° - 60°$ N are primarily in compression, while baselines extending to more northerly located sites are in extension, as visible in Fig. 9b. The same pattern is evident in the ITRF2000 baselines extending into Fennoscandia, as in Fig. 9a. ITRF2000 baselines are suject to changes in deformation style from northern, central and southern Europe. In fact, baselines south of POTS are predominantly in compression, with some exceptions showing extension, as for the cluster of baselines defined by sites in the southeast portion of Fig. 9a. North of POTS, extension is the dominant mechanism, as shown by baselines connecting POTS with sites north of it, except for a few cases, essentially NE directed.

Fig. 10a shows predictions for the same set of baselines generated using the PGR model described above. This model reconciles the pattern evident in the northern baselines, in particular a transition from shortening to extension as one considers more northerly sites.

In the following figures, the impact of PGR is shown on strain-rate tensor in the horizontal plane, as done in Marotta nd Sabadini (2004), rather than on baselines connecting pairs of sites, as done in Matotta and Sabadini (2002) and Marotta et al. (2004).

In the following, results on strain-rate eigenvalues and eigendirections, computed for a set of triangular domains covering the study area and compared with the corresponding ones obtained from the ITRF2000 velocity solutions, are considered. Strain-rates, within triangles with vertices corresponding to permanent GPS sites in Europe, can be retrieved from the gradients of the horizontal velocities at GPS sites, from standard strain-rate mathematical expressions as defined in continuum mechanics.

Strain-rate eigenvalues are in units of nanostrain per year, meaning, for example, shortening of 1 millimeter per year over a distance of 1000 km along the eigendirection corresponding to that eigenvaue, for compression, or lengthening of the same relative amount for extension, strain means relative deformation, so the same strainrate would be obtained for shortening (or lengthenin) of 1 cm over 10,000 km, and so on.

In our approach, this is accomplished assuming that the horizontal velocity components vary linearly with distance within each triangle. This constant space gradient provides a first order approximation of the strain-rate in each tectonic region embedded within the vertices of the triangles. This approximation can be improved by increasing the spatial resolution of the geodetic network by introducing new geodetic sites. The intersection point of the bisectors of each triangles has been elected as the reference point where strain-rate tensor is evaluated. Once strain-rate tensor is obtained, standard procedudes of matrix algebra is used to obtain eigenvalues and eigendirections. Within each triangular domain, strain-rate eigenvalues $\dot{\epsilon_1}$ and $\dot{\epsilon_2}$, for both data and model predictions, are computed following the procedure described in Devoti et al. (2002). The latter requires the inversion of a system of linear equations in six unknowns; four tensor components plus two velocity components at the reference point. The velocity gradient tensor can then be decomposed into its symmetric part and antisymmetric part, the first one providing the strain-rate eigenvectors and eigenvalues, after diagonalization procedure. The second one provides a rigid rotation rate. The errors associated with the geodetic strain-rate tensor are obtained by means of the covariance matrix associated with the velocity components at each site.

Models shown in Figure 5 are similar to PGR-21 and PGR-2 of Figure 3 and 4, in terms of mantle viscosity, except for the 120 km-thick lithosphere; blue and red bars referring to a lower mantle viscosity of 10^{21} Pa s, similarly to PGR-21, and cyan and yellow bars to a lower mantle viscoity of 10^{22} Pa s, simlarly to PGR-22..

Particular care is devoted to define the final set of triangular domains. Although several criteria could be followed for choosing the triangulation, a combination of geometric and reliability criteria has been adopted. The sub-set of sites, uniformly distributed through the study area, is selected first, in such a way that velocity is known with the lowest variance, by selecting the best geodetic GPS sites from ITRF2000. This most representative geodetic triangulation must be compliant with representative homogeneous tectonic units; this requirement is particularly important when Africa-Eurasia convergence, or active tectonics, is considered, as done in following sections.

In the selected triangulation, the geodetic ITRF2000 solution pattern obtained from GPS data is shown in Figure 6 in terms of strain-rate eigenvalues and eigendirections and relative errors bounds, that for the eigenvalues are provided by the radii of the circular sectors while for the eigendirections are provided by the azimuths bounding the colored areas. The strain-rate pattern is characterized by global SE-NW directed extension and SW-NE directed compression both at low and high latitude. Another characteristic of this pattern is the ratio between extension and compression, being significantly higher in Fennoscandi then in central Europe. Uncertainties in strain-rate predictions are generaly high, both in magnitude and in direction. Tectonic contribution, due to Africa-Eurasia convergence, to strain-rate is be considered in a following section, where PGR and tectonic contributions will be integrated, within the framework of a final comparison between modelled and geodetic strain-rate patterns. In this section, attention is devoted to PGR strain-rate patterns.

The 10^{21} Pa s lower mantle viscosity model, shown in Figure 4, predicts SE-NW extension in Fennoscandia, blue bars, in agreement with the geodetic data, and negligible SE-NW compression in the south, red bars. A peculiar situation is visible south of VIS0 and north of POTS, where PGR modelling predicts a SSE-NNW compression comparable in magnitude with that induced by the tectonic model characterized by the stiff Baltic Shield, with no ridge push, as it will be shown afterwards in Figure XXX.

Extension in the north and small compression between 52^o and 58^o north latitudes, are due to the outward motion from deglaciation center and from the small northerly oriented velocities from equatorial regions due to PGR, as shown previously in Figure 4.

Stiffening of the lower mantle to 10^{22} Pa s Pa s, cyan and yellow, is responsible for an increase in extension in all the triangles, as shown by the cyan bars, that improves, north of VISO, the fit with geodetic strain-rates.

South of the line connecting ONSA and VIS0 and north of POTS, compression, yellow bars, is increased with respect to the 10^{21} Pa s lower mantle viscosity model. South of POTS, the contribution of this PGR model to both compression and extension is larger than the 10^{21} Pa s lower mantle viscosity model, but compression points in the SE-NW direction, in contrast with observation.

After having analyzed the effects of active tectonics on strain-rate in section XXX, the combined effects of PGR and Africa-Eurasia convergence will be analyzed, showing that both geophysical processes contribute to observed geodetic strain-rates.

Sea-Level Changes

If the ice sheets of Antarctica and Greenland melts, sea level is expected to rise. This is in fact a major concern, that global warming of the Earth might induce melt of the present-day large ice sheets, thus inducing a global sea-level rise, leading to floodings of lowlands: it is thus relevant to study the effects of ice-sheet melting on sea-level changes.

The link between ice melt and sea-level rise might seem obvious, but it is not, once it is realized that ice melt could actually induce a sea-level drop at some places on the Earth's surface, rather than sea-level rise.

In fact, an ice sheet on a continent attracts the water of the ocean, due to its own mass: the water near the ice sheet will be elevated with respect to a situation in which this gravitational interaction between the masses of the ice an of the water does not occur, because of the absence of the ice sheet. If the ice sheet melts, this gravitational attraction effect, or self-gravitation, would disappear. Woodward (1888) showed that this effect is not negligible and derived the following formula for the ratio of the change in sea level with self-gravitation taken into account and the sea-level change without the self-gravitation of the ice sheet, treated as s point-source

$$\left(\frac{1}{2\sin(\theta/2)} - 1 - \frac{\rho_E}{3\rho_w}\right) / \frac{\rho_E}{3\rho_w}. \tag{29}$$

where θ is the angular distance from the ice sheet, ρ_E the mean density of the Earth, and ρ_w the density of the water. Equation (6.1) shows that the ratio is not dependent on the total mass of the ice sheet. A derivation of formula (6.1) can also be found in Farrell and Clark (1976). From equation (6.1) it can be easily shown that sea level will drop in the oceans as a result of ice melt within a distance of about 20°, that on the

Earth's surface, corresponds to about 2200 km from the former ice sheet. Even within 60°, equals to about 6700 km, sea level will rise less than if the amount of water from the ice sheet had been distributed uniformly, or eustatically, over the oceans. At distances exceeding 60° sea level will rise more than the eustatic value.

The relation between ice melt and sea-level change is clearly not so simple as the reader might think. In fact, eventhough the mass of ice that melts equals that of the total amount of water that is added to the oceans, the redistribution of the melt water is not uniform.

The situation becomes even more complicated, once it is realized that the solid Earth is not rigid, as shown by the picture portrayed by Figure 4 of this chapter.

The relation between ice melt, sea-level rise, modified by equation 6.1 because of self-gravitation, must so be expanded to a three-component relation involving continental ice, sea level and solid Earth. Ice melt will cause a sea-level change that will both induce solid-Earth deformation. But solid-Earth deformation in turn will induce a sea-level change again. This sea-level change will induce solid-Earth deformation, and so on. It is thus clear that this relationship is a non-linear one: the formulation for this relationship has been derived in the 1970s and have become known as the sea-level equation, which is an integral equation. This equation for sea-level change reads (Farrell and Clark, 1976)

$$S = \rho_I \frac{\phi}{g} * L + \rho_w \frac{\phi}{g} * S + C, \tag{30}$$

in which S denotes the change in sea level, L the change in (continental) ice thickness, ϕ the Green function for the variation in the gravitational potential, ρ_I the density of the ice and g surface gravity. C is a constant which is to be determined by invoking the condition that the total amount of ice change is equal to the total amount of sea-level change. The asterisk denotes time convolution. Sea-level change S is both on the left-hand side and on the right-hand side of equation (6.2), which make the latter an integral equation.

Thanks to this sea-level equation, by which the sea-level change and solid-Earth deformation interrelationship as function of ice mass variations and solid-earth models can be taken into account, present-day sea-level variations can be modeled adequately, both for sea-level changes due to Pleistocene deglaciation and for those induced by recent continental ice mass changes (e.g., Farrell and Clark, 1976; Peltier and Andrews, 1976; Wu and Peltier, 1983; Nakada and Lambeck, 1987; Lambeck *et al.*, 1990; Mitrovica and Peltier, 1991; Johnston, 1993; Di Donato *et al.*, 2000; Mitrovica and Vermeersen, 2002).

The importance of PGR on the interpretation of sea-level variabilities has been addressed by several authors. Lambeck and Nakiboglu (1984) find that post-glacial rebound contributes between 30 and 50 per cent of the present-day secular rise in sea level of 1.8 mm/yr (Douglas, 1995). Although there are many uncertainties in estimating such a percentage, these findings show the potential importance of including solid-Earth deformation processes in studying sea-level changes.

The following Figure 6 shows a global picture of sea-level changes due to PGR, based on the same kind of modelling that has been used to produce previous figures of verticsl and horizonytal PGR induced deformation, in terms of surface vertical and horizontal velocities, baseline rates and strain-rates. These results show that sea-level changes are highly correlated with vertical deformation rates of the lithosphere, as indicated for example by the result that the largest values of sea-level drop coincide with the deglaciation centres of Figure 4, where the lithosphere is subject to uplift. In fact, since the surface of the sea has to remain an equipotential surface, and vertical displacement of this surface is about one order of magnitude lower that the vertical displacement of the lithosphere, or sea bottom, sea-level change S, depicted in this figure by $S = S_1 - S_0$, decreases, since the distance between the equipotential and the sea bottom is reduced in uplifting areas. The following figure deals with the vertical displacement of the surface of the sea that moves up or down, with respect to the Earth's centre, in order to remain an equitotenial, eventhough ice, water and solid Earth material is redistributed, thus modifying the gravity field.

The redistribution of mass caused by PGR, in fact, is responsible not only for vertical and horizontal displacements of sites over the Earth's surface, but also for perturbation of the gravitational potential, as on the other hand indicated by the equations of previous sections. Uplifting of the Earth's crust, for example, over North America, Fennoscandia and Antarctica, causes dense crustal material replacing the air or the water, thus infilling the mass deficit left by ice-sheet disintegration. This growing positive mass over such wide continental areas is responsible for increasing the gravity over these areas, by Newton law. PGR has thus the effect of changing the position of surfaces of constant gravity potentials, or equipotentials, with respect to the Earth centre, In particular, one of these equipotentials, named geoid, representing the surface of mean sea level, excluding dynamical phenomena such as currents, waves due to winds and tides, is displaced due to PGR, as shown in Figure XX. This figure shows the vertical displacement of such a surface, or geoid, in millimeter per year, as predicted by the same kind of modelling used in previous figures. As said above, the geoid is increasing, by a few millimeter per year, over the deglaciation centres, while is subject to a small decline at far distances from these centres. As anticipated, geoid changes are smaller than sea-level changes, since the latter is mainly controlled by the vertical diaplacement of the Earth's surface, or sea bottom.

Of course, such an increase of gravity cannot be felt by human beings, but moder satellite do sense such an increase of gravity. Dedicated gravity space missions, such as GRACE (Gravity Recovery And Climate Experiment) mission from NASA, for example, show that gravity, and thus the geoid, is presently growing over Hudson Bay, in agreement with previous picture, as portrayed by the following figure obtained from GRACE gravity data.

This figure portrays the displacement rate, in millimeter per year, of the geoid, obtained from GRACE data, that represents the real changes of the Earth'gravity, obtained by extracting the secular trend or drift from GRACE data, pertaining to the years XXXX-XXXX, which can be directly compared with Figure XXX.

Comparison between Figures xx and x shows on the other hand that such a correlation of increasing geoid is not so obvious when the other two deglaciation centres, Fennoscandia and Antarctica, are considered and the reason for that can be found in other phenomena, such as present-day melting in Antarctica, that mask the effects of PGR. In any case, it is the first time after Newton, that scientists have the possibility to monitor gravity changes of the Earth, thanks to gravity space missions.

Similarly to what it has been done for displacement patterns in Figures 4 and 5, relative to global and European perspectives, Figures 6.5a and 6.5b now focus on modelled present-day rates of sea-level change, in northern Europe and in the Mediterranean region. The lower mantle viscosity considered for these sea-level changes calculation is consistent with inferences based on post-glacial relative sea-level variations in northern Europe (Lambeck *et al.*, 1990) and numerical predictions of post-glacial sea-level change in southern Europe (Mitrovica and Davis, 1995). The sea-level equation (6.16) has been solved by using the pseudo-spectral algorithm by Mitrovica and Peltier (1991) with a truncation at degree and order 256, so the spatial resolution is sufficient to model sea-level increase in small regions, such as those considered herein.

As expected, the highest rates are obtained in the center of deglaciation, the Gulf of Bothnia, wher uplift of the land causes a sea-level fall of -11 mm/yr. In the periphery of the uplifting region, the land is subsiding due to the collapse of the peripheral bulge, and sea-level thus increases along the coastal areas of northern Europe, by as much as 1 - 2 mm/yr. This sea-level increase along the coasts of northern continental Europe is not constant but is subject to large variabilities, from 0.4 mm/yr near the French coast to 1 - 1.5 mm/yr along the coasts of The Netherlands and Germany. It should be noted that collapse of the peripheral bulge decreases in central Europe and, further to the south, sea level is actually increasing, with rates in the order of 0.6 - 0.8 mm/yr in the central Mediterranean, due to the subsidence of sea bottom caused by the water load of the Mediterranean itself.

The periphery of the Mediterranean is characterized by a weak sea-level rise, with rates from 0.2 to 0.4 mm/yr in the Adriatic Sea and about 0.4 mm/yr along the Mediterranean coast of France. Here the rates of sea-level change have small values due to the levering effect, as discussed in Nakada and Lambeck, 1989, for example. Such an effect implies that subsidence of sea basins is accompanied by uplift of surrounding continents. This sea-basin subsidence is due to mantle material presently flowing from the Mediterranean region towards the Fennoscandian region in northern Europe, concordant with the main stream from the Earth's equtorial regions to deglaciatin centres, as anticipated in previous sections. These values are comparable with the sea-level changes due to active tectonics in the central Mediterranean. Finite-element calculations of active tectonics in peninsular Italy have indicated in fact that overthrusting of the Apennines onto the Adriatic plate is responsible for sea-level increases of about 0.4 mm/yr, as detailed in Negredo *et al.*, 1997). Sea-level change values in central Mediterranean due to post-glacial rebound is thus comparable with that due to active tectonics, as first enlightened by Di Donato *et al.*, 1999.

CONTINENTAL PLATES

Introduction

Crustal deformation in Europe are influenced not only by PGR, as shown in previous sections, but also by plate tectonic forces, including boundary forces associated with Africa-Eurasia convergence, spreading at the mid-Atlantic ridge and subduction-related ones.

The reader can thus wonder whether there are regions in Europe where PGR or tectonic effects on baseline rates or strain-rates can be ignored, or whether there are regions where both phenomena act in concert, in order to investigate the nature and origin of intraplate deformation in continental Europe. More generally, it is relevant to analyze the complex geometric interplay between tectonics and PGR in European continental deformation.

These issues are now considered in this section, in order to compare predictions generated from thin-sheet tectonic models and PGR simulations.

Modelling intraplate deformation

Mathematical model of continent-continent collision, to simulate the effects of active tectonics in Europe due to Africa-Eurasia convergence, is based on the thin sheet scheme [*England and McKenzie*, 1983; *Marotta et al.*, 2001]. This mathematical scheme allows to analyze the sensitivity of modelling results on changes in velocity forcing along the Atlantic Ridge and variations in the lithospheric strength, or rheology, of various European sub-domains. The analysis finally highlights a combined tectonic model pluds PGR which best-fits ITRF2000 data.

An incompressible, viscous model is adopted to investigate tectonic deformation in the Mediterranean-Fennoscandian region driven by Africa-Eurasia convergence and mid Atlantic ridge opening (Fig. 1a). This figure shows the finite element mesh, used to solve the differential equations of motion by means of the finite element scheme first presented in Marotta et al. (2004). Once compared with Figure 2, this purely viscous model includes only the dashpot, as appropriate for long lasting tectonic processes of million years. The treatment of the lithosphere as an incompressible, viscous fluid is in fact widely adopted in models of long time-scale geological processes [*Turcotte and Schubert*, 2002].

The deformation field, sampled at the nodes of the finite element mesh, is expressed in terms of crustal velocities and baseline rates obtained from the thin-sheet approximation implemented by *Marotta et al.* [2001] and modified in Marotta et al. (2004) in such a way to include spherical geometry. The lithosphere is treated as a stratified viscous sheet with constant total thickness, overlying an inviscid asthenosphere, which assures a stress-free condition at the base of the plate. The thin sheet approximation assumes that wavelengths characterizing plate deformation are larger that plate thickness. Other thin-plate aproximations are based on the assumption that the lithosphere is isostatically compensated and on vertical averaging of lithospheric rheology. Three-dimensional problems are thus reduced to two-dimensional ones, where horizontal velocity components do not depend on the depth: horizontal components of the momentum equation are vertically integrated through the plate, and are then solved using a 2-D finite element mesh, as shown in Figure 1, thus estimating only velocity horizontal components.

The western and southern borders of the model domain coincide with the mid Atlantic Ridge and Africa-Eurasia plate contact, respectively, where velocity boundary conditions are applied. The eastern border of the model ends at 45° E meridian, within the intracratonic East European Platform, where stress transmission from boundary forcing is expected to be negligible. The domain is discretized using planar finite triangular elements sufficiently small in size so that the surface of each individual grid element can be considered as flat.

In the tectonic modelling under study, the differential equations describing the mechanical behaviou of the plates are equations (1) in spherical geometry and concide with those of PGR only for the radial r and colatitudinal θ components of the divergence of the stress tensor $\nabla \cdot \sigma$, since plate deformation modelling under applied horizontal tectonic forces includes the longitudinal component ϕ of the momentum equations. Differently from PGR where due to the type of forcing and symmetry of the problem only the radial and colatitudinal equations are relevant, in the tectonic modelling the longitutidal equation must be included, since horizontal tectonic forces have, in general, both θ and ϕ components. In the same coordinate system the θ, ϕ and r, components of the momentum equations are then [*Schubert et al.*, 2001]

$$\frac{1}{r}\frac{\partial}{\partial\theta}\sigma_{\theta\theta} + \frac{1}{rsin\theta}\frac{\partial}{\partial\phi}\sigma_{\theta\phi} + \frac{\partial}{\partial r}\sigma_{\theta r} + \frac{1}{r}[(\sigma_{\theta\theta} - \sigma_{\phi\phi})\cot\theta + 3\sigma_{\theta r}] = 0 \tag{31}$$

$$\frac{1}{r}\frac{\partial}{\partial\theta}\sigma_{\phi\theta} + \frac{1}{rsin\theta}\frac{\partial}{\partial\phi}\sigma_{\phi\phi} + \frac{\partial}{\partial r}\sigma_{\phi r} + \frac{1}{r}(3\sigma_{\phi r} + 2\sigma_{\phi\theta}\cot\theta) = 0 \tag{32}$$

$$\frac{1}{r}\frac{\partial}{\partial\theta}\sigma_{r\theta} + \frac{1}{rsin\theta}\frac{\partial}{\partial\phi}\sigma_{r\phi} + \frac{\partial}{\partial r}\sigma_{rr} + \frac{1}{r}(2\sigma_{rr} - \sigma_{\theta\theta} - \sigma_{\phi\phi} + \sigma_{r\theta}\cot\theta) + f_r = 0 \tag{33}$$

where f_r denotes the gravitational body force term. As usual, the stress can be written as

$$\sigma_{ij} = \tau_{ij} - p_0\delta_{ij} \tag{34}$$

where p_0 is the hydrostatic pressure.

Under the assumption that only horizontal tectonic forces are active and are those associated with Africa-Eurasia convergence and that there is no shear stress or basal drag at the base of the lithosphere, the components $\sigma_{r\theta}$ and $\sigma_{r\phi}$ within these general equations are neglected. By applying the constitutive equation for an incompressible, viscous material and the conditions for isostatic balance, the momentum equations become after integration over the thickness of the lithosphere

$$\begin{aligned}&\frac{\partial}{\partial\theta}[2\bar{\mu}(\frac{\partial}{\partial\theta}u_\theta + u_r)] + \frac{1}{sin\theta}\frac{\partial}{\partial\phi}[\bar{\mu}(\frac{1}{sin\theta}\frac{\partial}{\partial\phi}u_\theta + \frac{\partial}{\partial\theta}u_\phi - u_\phi cot\theta)]\\ &+[2\bar{\mu}(\frac{\partial}{\partial\theta}u_\theta - \frac{1}{sin\theta}\frac{\partial}{\partial\phi}u_\phi - u_\theta cot\theta)]cot\theta = \frac{g\rho_c R}{2L}(1-\frac{\rho_c}{\rho_m})\frac{\partial}{\partial\theta}S^2\end{aligned} \tag{35}$$

$$\begin{aligned}&\frac{\partial}{\partial\theta}[\bar{\mu}(\frac{1}{sin\theta}\frac{\partial}{\partial\phi}u_\theta + \frac{\partial}{\partial\theta}u_\phi - u_\phi cot\theta)] + \frac{1}{sin\theta}\frac{\partial}{\partial\phi}[2\bar{\mu}(\frac{1}{sin\theta}\frac{\partial}{\partial\phi}u_\phi + u_\theta cot\theta + u_r)]\\ &+[2\bar{\mu}(\frac{\partial}{\partial\theta}u_\phi + \frac{1}{sin\theta}\frac{\partial}{\partial\phi}u_\theta - u_\phi cot\theta)]cot\theta = \frac{g\rho_c R}{2L}(1-\frac{\rho_c}{\rho_m})\frac{1}{sin\theta}\frac{\partial}{\partial\phi}S^2\end{aligned} \tag{36}$$

where $\bar{\mu}$ denotes the vertically averaged viscosity of the lithosphere. S denotes the crustal thickness, L is the lithospheric thickness, ρ_c and ρ_m denote the densities of the crust and lithosphere, g is the gravity and R is the radius of the Earth.

The third unknown u_r, is eliminated from these equations by invoking incompressibility and by assuming that the radial strain-rate $\frac{\partial}{\partial r}u_r$ vanishes. As detailed in Marotta et al. (2004), u_r can thus be expressed as

$$u_r = -\frac{1}{2}\left\{\frac{\partial u_\theta}{\partial \theta} + \frac{1}{sin\theta}\frac{\partial u_\phi}{\partial \phi} + u_\theta cot\theta\right\} \tag{37}$$

The thin sheet model is appropriate for estimating horizontal components of velocity field u_θ, u_ϕ only: the radial component of the differential equation guarantees the condition of isostatic equilibrium, requiring that mass excess is supported in the radial direction by radial stress.

By specifying crustal thickness S and boundary conditions, numerical integration of previous equations yields the stationary tectonic deformation field. Velocity is approximated by linear polynomial interpolating functions within each finite element and numerical integration is performed by Gaussian quadrature with 7 integration points (Marotta et al., 2004).

A series of numerical tectonic deformation models are summarized in Table 1 as Models 1-7, distinguished in terms of lithospheric viscosity and velocity boundary condition along the north Atlantic Ridge. Model inputs are discussed next.

Lateral variations in lithospheric strength can be taken into account in the modelling by specifying distinct viscosity values at each element of the grid. Lateral viscosity variations are referred to the European lithosphere, treated as the reference domain where a prescribed reference viscosity is fixed.

In the homogeneous viscosity model, velocity pattern is shaped by velocity boundary conditions and is unaffected by changes in the lithospheric viscosity in the range 10^{23} Pa s t0 10^{25} Pa s; the value of 10^{25} Pa s is the reference viscosity that guarantees numerical stability when lateral viscosity variations are introduced.

Two isoviscous lithospheric domains are considered, the first corresponding to the Mediterranean domain, extending from the Tyrrhenian Sea to the eastern limit of the Pannonian Basin through the Adriatic Plate, while the second is the East European Platform, as shown in Fig. 1a.

This modelling has some similarities with earlier work by *Grunthal and Stromeyer* [1992], who modeled the stress field in central Europe by means of elastic rheology with laterally varying rigidities that simulate different tectonic units. A viscous fluid with laterally varying viscosiy is considered herein, to compare model predictions with geodetic observations.

Velocity boundary conditions are relative to the Eurasian plate. The velocity of Africa relative to Eurasia is prescribed by NUVEL-1A, red arrows at the bottom border of the finite element mesh, Fig. 1a, the pattern reflecting Africa-Eurasia continental convergence of the order 1 cm/yr. These velocities impose a counterclockwise rotation of the Africa plate with respect to Eurasia. Relative to a fixed Eurasia, ridge push forces acting along the north Atlantic Ridge are also modelled. In the simulations, these forces are parameterized in terms of velocity boundary conditions applied along the ridge, to simulate the effects of the line forces acting along the plate boundary, as described in *Richardson et al.* [1979]. These ridge velocities are represented by thick yellow arrows, to make it clear that these velocity are derived in a different manner with respect to those related to Africa-Eurasia convergence.

Line forces normal to the ridge, resulting from applied velocity boundary conditions, can be estimated from the eigenvalues of the stress tensor within those elements whose left sides define the ridge, in order to verify that boundary conditions induce realistic forces within the lithospheric model. Along the westernmost part of the Atlantic Ridge, predicted ridge push forces range from $\sim 10^{12}$ N/m, for an imposed velocity boundary condition of about 1 mm/yr, to $\sim 10^{13}$ N/m for 5 mm/yr, the latter representing an upper bound for ridge push forces, as shown by Richardson and Reding (1991). Ridge velocities are not constant along the ridge, but are scaled with respect to the spreading velocities deduced from NUVEL-1A, so that imposed ridge velocities of 1 mm/yr and 5 mm/yr are of the order of 1/20th and 1/4th of the full spreading rate ($\sim$ 2 mm/yr) according to NUVEL-1A.

Along the Aegean trench, geodetically determined velocities by *McClusky et al.* [2000] are applied as velocity boundary conditions to an equal number of nodes, as indicated by the blue arrows in Fig. 1a, from west to east. These sites are LOGO (25 mm/yr), LEON (33 mm/yr), OMAL (30 mm/yr), ROML (32 mm/yr), KAPT (33 mm/yr) and KATV (30 mm/yr). These velocities reflect trench subduction forces along this boundary and represent the velocity of these geodetic sites with respect to Eurasia.

The eastern boundary of the model domain is held fixed, and to avoid large effects from artificial stress accumulation, due to this boundary condition, a shear stress free boundary condition is imposed at this border, as indicated by the red dots along the right boundary of the model. The imposed conditions at the eastern boundary would be consistent with a possible decoupling between the western and eastern parts of the Eurasia plate, as suggested by Molnar et al. (1973). This condition clearly imples that intraplate deformation of Eurasia due to Africa-Eurasia convergence and Atlantic Ridge push takes place within the model domain.

The contact between the East European Platform and Arabian Plate is fixed in the simulations carried out in this section, as indicated by the pink triangles in the south-eastern part of Fig. 1a. This condition will be released in following sections, focused on strain-rate in the Mediterranen.

Crustal thickness variations have been obtained from linear interpolation onto the adopted grid of model CRUST 2.0 Bassin et al. (2000).

Monitoring intraplate deformation via GPS analyses

The goal is now to explore the effects of lateral viscosity variations and varying velocity boundary condition along the Atlantic Ridge on intraplate deformation in Europe from active tectonics.

The first three models in Table 1 correspond to different imposed velocity boundary condition along the north Atlantic Ridge, while the other boundary conditions are as described above. The horizontal velocities predicted within the modelled domain for these three models are shown in Figs. 2a-c, respectively.

Fig. 2a focuses on the efects of Africa-Eurasia convergence on intraplate velocity pattern within the model domain, where crustal velocity gradually diminishes from $\sim$ 2 mm/yr at latitudes of 45° along the Alpine front to 0.2 mm/yr in central Fennoscandia. The velocity field driven by the African indenter extends through whole central Europe, approximatively in the north-western direction, with the velocity iso-contours roughly parallel to the collision front.

For 1 mm/yr along the north Atlantic Ridge, the case of Fig. 2b, a rotation from NW to NE in the global velocity direction is obtained in central and northern Europe. With respect to Fig. 2a, the velocity is increased throughout the western part of the study domain and an increase from 0.2 mm/yr to 0.5 mm/yr is obtained at the latitude of Fennoscandia.

When the velocity along the Atlantic ridge is increased to 5 mm/yr, an upper bound in terms of ridge push forces as discussed above, the case of Fig. 2c, tectonic velocity in England and Fennoscandia reach magnitudes of $\sim$ 3 to $\sim$ 2 mm/yr, respectively. The imprint of western and southern boundary forcings are evident in this tectonic velocity pattern. In the Alpine Front, north-directed motions up to $\sim$ 4 mm/yr are obtained in Fig. 2c and, to the north of this region, sites in central Europe are now characterized by an eastern velocity component.

Figure 2 velocity patterns shown represent intraplate deformation for homogeneous viscosity models, with Figure 2-c portraying unrealistic deformation magnitudes for large ridge push forces.

The effect of incorporating lateral variations in lithospheric stiffness into the tectonic model is exploited in Figures 3a-b, corresponding to viscosity increase of two orders of magnitude in the East European domain with respect to the reference viscosity of 10^{25} Pa s. The two simulations differs in terms of boundary condition along the north Atlantic Ridge, either 0.0 mm/yr in Fig. 3a, Model 4, or 5 mm/yr in Fig. 3b, Model 5.

Stiffening of the lithosphere within the East European Platform has a prominent effect on tectonic velocities within that region. From north to south across the platform, velocity gradients are reduced considerably in Fig. 3a with respect to Fig 2a. A nearly constant crustal velocity of $\sim$ 0.6 mm/yr thus characterizes a large portion of the stiffened craton, including Fennoscandia.

The stiffened lithosphere shields the Baltic region and Fennoscandia from the eastward-directed velocity driven by the ridge and dampens the velocity gradients within the East European Platform, as shown by Fig. 2c, Model 3, compared to Fig. 3b, Model 5. Increasing the viscosity of the East European Platform results into a sustantial reduction, in northern Europe and Fennoscandia from north Atlantic Ridge-induced velocity.

Model 6 and 7 are characterized by one order of magnitude viscosity reduction within the Mediterranean lithosphere, Figs. 3c and d, respectively. Comparison of Fig. 3c with Fig. 2a, for example, shows that a large amount of the deformation driven by push of Africa is accomodated in the weakened lithosphere, with velocity gradients concentrated in the Mediterranean. The relatively small velocities within Fennoscandia in Fig. 2a now extend well in the south into central Europe in Fig. 3c. As expected, intraplate deformation in Europe due to Africa-Eurasia convergence is sensitive to the amount of deformation which takes place within the Mediterranean lithosphere.

A velocity along the north Atlantic Ridge is introduced in Model 7, characterized by the weakened Mediterranean lithosphere, and the result of Fig. 3d can be compared with Fig. 2c. Weakening of the Mediterranean sub-domain favours the diffusion into Europe of the eastward-directed velocity driven by the Atlantic spreading. It is in fact remarkable the pronounced eastward migration of the 4 mm/yr contour in Fig. 3d relative to Fig. 2c.

Table 1. Tectonic models

Model	*Rheological heterogeneities*	*Ridge velocity b.c (mm/yr)*
1	*No rheological heterogeneities*	0.0
2	*No rheological heterogeneities*	1.0
3	*No rheological heterogeneities*	5.0
4	*Stiff East European Platform*	0.0
5	*Stiff East European Platform*	5.0
6	*Soft Mediterranean Domain*	0.0
7	*Soft Mediterranean Domain*	5.0

Deformation style in Europe from tectonics is first analyzed in terms of baseline rates, as done for PGR in previous section.

Baseline rate modelling is first compared with ITRF2000 data predictions, with respect to VAAS, as in Figure 6a. Figs. 7b-c show predictions of baseline rates generated from a tectonic model subset. Fig. 7b is based on Model 1 of Table 1, characterized by an homogeneous lithosphere, by Africa-Eurasia convergence and no Atlantic Ridge forcing. For this homogeneous viscosty VAAS-referenced baselines show a general pattern of shortening, except for a limited extension for short baselines connecting three sites east of VAAS. It is thus remarkale that, except for this limited extension, the style of baseline rates is opposite to the observed pattern of Figure 6.

In Fig. 7c, results are shown for lateral variations in plate strength, Models 5, characterized by stiffening of the East European Platform. With respect to homogeneous model results, Model 1, Fig. 7b, Model 5 in Fig. 7c improves the fit to observed baseline rate pattern by yielding extension for baselines directed from VAAS to south-east to south-south-west. These findings are also unsatisfactory, since modelling predicts shortening for other baselines, in contrast to geodetically retreieved baseline rates. These results from Models 1 and 5 show that tectonic modelling has the tendency to predict too much compression, even at high latitudes; this modelling performs better at intermediate latitudes in central Europe, as shown in the following.

Figs. 10b-d show POTS-referenced baseline rate results generated by using the same three tectonic models of Figs. 7b-d.

The uniform lithosphere Model 1 of Fig. 10b is driven by forcing along the southern Africa-Eurasia boundary and the resulting northward decrease in velocity, shown in Fig. 2a, yields a shortening of all baselines, similarly to what already observed for VAAS-referencd baslines in Fig 7b, thus failing to reproduce the extension of the baselines connecting sites north of POTS.

Fig. 10c shows the impact on POTS-referenced baselines of increasing the viscosity in the East European Platform, Model 5. The combined effect of viscosity increase in the Baltic Shield and push from the Atlantic Ridge reproduces the observed pattern, characterized by dominant shortening between POTS and the Mediterranean and extension between POTS and Fennoscandia. The effect of the high viscosity in the shield is to maintain into southern and central Europe, south of POTS, the north-directed motion driven by the Africa indenter. The stronger platform acts to significantly reduce the modelled tectonic deformation across central and northern Europe, including Potsdam. Sites clustered near the Africa-Europe plate boundary in the southwest, for latitudes $43° - 48°$ N, are thus expected to move toward a relatively more stationary Potsdam, which results into baseline shortening. Fig. 10c thus shows that realistic tectonic modelling, such that characterizing Model 5 with lithospheric stiffening in the Baltic Shield and with velocity applied along the Atlantic Ridge to simulate ridge push forces, is able to reproduce the dominant shortening of baselines from POTS south and to contribute to extension north of it.

For the final tectonic model of this sequence, Model 7, Fig. 10d, the weakened Mediterranean domain, in contrast to the strong Baltic Shield case, is responsible for a decrease in horizontal motions north of POTS through the Fennoscandian region, which leads to shortening of baselines ending at BIFROST sites.

In analogy with PGR analyses, strain-rate patterns are now considered for tectonic modelling.

Fig. 2 shows strain-rate results from tectonic models 3 and 1. Except for POTS-PENC-BOGO and POTS-GRAZ-PENC where extension dominates, as indicated by cyan bars, model 3 predicts SE-NW compression, yellow bars, due to the combined effects of Africa-Eurasia convergence and Atlantic Ridge spreading. The dominant extension in the triangle POTS-GRAZ-PENC is concordant with the extension along the baseline POTS-PENC first obtained by Marotta and Sabadini (2002), where it has been attributed to the lateral extrusion induced by the Alpine front. The striking result of model 3 is that modeled compression acts at right angles with respect to that derived from geodetic data. The two triangles quoted above are the only ones in which complete agreement is obtained, in terms of extensional eigendirections and eigenvalues, between modelling results and observations. Another major problem of this model is that it predicts compression north of VIS0, while ITRF2000 data indicate extension. This inconsistency suggests that the effects of ridge push are too large and that by reducing them, misfit should be also reduced. This hypothesis is tested by considering the end-member model 1, in which no ridge push is considered (see Table 1). This extreme case implicitly assumes that deformation due to ridge push is completely adjusted in the part of the domain between the ridge and the area under study. As expected, in model 1 the previous disagreement between data and prediction is significantly reduced, with compression becoming negligible at high latitudes, red bars, and subject to a clockwise rotationin central Europe, in agreement with geodetic data analysis.

The introduction of rheological heterogeneities in the tectonic model is crucial for obtaining the best fit model, as shown by Fig. 3, for the stiff Baltic Shield. Even in the presence of a velocity boundary condition of 5.0 mm/yr along the Atlantic Ridge, a substantial modification is obtained in the strain-rate patterns north of the line connecting DELF to BOGO, where compression is reduced with respect to Fig. 2, yellow bars, and SSW-NNE extension appears. The reduction of compression in the stiff Baltic Shield and south of it is easily understood in terms of the reduced flow within the high viscosity region, which acts as a barrier that annihilates the propagation of the velocity driven by Africa-Eurasia convergence and Atlantic Ridge push within the Shield. When zero velocity boundary condictions are assumed along the western boundary, as in the case of model 4, the ridge push effects disappear everywhere, leaving space to the Africa indenter and to an increase in compression with respect to Fig. 2, as indicatd by the red bars. Rotation from NNW to NNE of compressive eigendirections occurs now south of Potsdam, in good agreement with ITRF2000 analyses. Due to the stiffening of the Baltic Shield, the compressive effects of Africa-Eurasia push almost disappear north of VIS0, supporting the hypothesis that another geophysical process must be invoked to explain the large SE-NW extension, which is PGR as shown in fact in the previous section.

When comparison of Fig. 4 is drawn with Fig. 3, relative to PGR ffects on strain-rates, it is noticeable that the tectonic model with the stiff Baltic Shield influences the compression south of ONSA-VIS0, with largest contributions south of DELF-POTS-BOGO, while PGR contributes to the compression in the central European region between ONSA-VIS0 and DELF-POTS-BOGO and to the extension north of ONSA-VIS0. These findings show that tectonics dominates the strain-rate pattern south of POTS, while PGR is the dominant mechanism north of VIS0 in Fennoscandia and that both tectonics and PGR contribute to the deformation in the European continental region between the latitudes of POTS and VIS0. That this is the case is better shown in Fig. 5, in which the combined effects of the best performing tectonic model 3 and PGR are portrayed, with panels a and b referring to the northern and southern domains. Model predictions denoted by the bars are superimposed to strain rates resulting from ITRF2000, represented with their errors in the magnitude and azimuth, with light blue denoting extension and light red compression. It should be noted that data are portrayed with only one side of the admissible area of variability of the magnitude, to leave room for depicting the already enlightened possible change from compression into extension or vice-versa, caused by some errors being larger in absolute value than the corresponding datum. Model results reproduce well the general pattern of NW extension in panel a, while, in panel b, only the NE compression is reproduced. Agreement in the azimuth is obtained by combining the tectonic model 3 with the PGR model characterized by the lower mantle viscosity of 10^{22} Pa s for high latitudes, as shown in panel a), and with the 10^{21} Pa s lower mantle viscosity PGR model for intermediate latitudes, as in panel b, which supports the existence of mantle lateral variations in mantle viscosity with a stiffer mantle underneath the Baltic shield, coherently with the laterally heterogeneous tectonic model.

It is worthwhile to remark that modelling has the tendency to reproduce lower bounds of observed strain-rates.

Findings from this section indicate that Africa-Eurasia convergence has the dominant role in determining plate defrmation patterns south of 52° north latitude, corresponding to the latitude of Potsdam (POTS). This tectonic mechanism has no effect north of about 58° north latitude, corresponding to the latitude of Onsala (ONSA), in Fennoscandia.

PGR is the only mechanism which has a signature north of Onsala. Strain-rate in central Europe, between Onsala and Potsdam, is affected by both mechanisms. Strain-rate analysis thus reinforces the results of the previous section based on baseline rates, that plate deformation in Europe is the result of an interplay between forces induced by tectonics and PGR. Lateral variations in lithospheric strength in the eastern European platform play a crucial role in separating the two domains where tectonics and PGR have their major influence, south of Potsdam and north of Onsala. The combination of tectonic modelling, characterized by a stiff eastern European platform and Atlantic Ridge push effects completely absorbed within the western oceanic portion of the model domain, and PGR modelling are the best performing ones, once compared with geodetic strain-rates, well reproducing the major features of compression south of Onsala and SE-NW extension north of Onsala.

PLATE DEFORMATION AT SUBDUCTION ZONES

Introduction

The tectonic setting of the Mediterranean is controlled by collision between the African and Arabian plates with Eurasia and by subduction in the Hellenic and Calabrian Arc [e.g., *McKenzie*, 1970; *Jackson and McKenzie*, 1988]. Figure 1 portrays the major tectonic structures; the solid line indicates the Africa-Eurasia plate boundary and major faults, like the North Anatolian Fault (NAF), East Anatolian Fault (EAF) and South Anatolian Fault (SAF). In spite of the Africa/Arabia vs Eurasia convergence, several regions exhibit extension in the Earth's crust, such as the Alboran Sea, the Algero-Provençal basin, the Tyrrhenian and the Aegean Seas; topography in kilometers, is given by contour lines..

This section focuses on geophysical modelling of continental collision between Africa and Eurasia, on the basis of thin plate theory, as anticipated in previous sections, generalized in such a way to include the effects of subduction in the Hellenic and Calabria Arcs and on strain-rates from permanent GPS receivers in the Mediterranean. Such velocity fields are based on GPS measurements from *Clarke et al.*, 1998, *McClusky et al.*, 2000] and from the Matera Geodesy Center (CGS) of the Italian Space Agency

(ASI).,

The Mediterranean: a natutal laboratory for understanding plate behaviour at subduction zones

Earthquakes in the Mediterranean are not confined to a single fault, which implies that deformation in this region cannot be simply described by relative motions of rigid blocks. Within the broad deforming belts in the continents some large, flat, aseismic regions such as central Turkey or Adriatic sea, behave like rigid blocks which can usefully be considered as microplates [e.g. *Jackson and McKenzie*, 1984; *Ward*, 1994].

This section focuses on the numerical modelling of the major tectonic processes which are active in the Mediterranean and on the comparison between model predictions and geodetic data, as done previously. In the present analysis, the modelling includes convergence between Africa/Arabia and Eurasia plates and the additional forces acting at plate boundaries due to subduction. Differently from PGR modelling, vertical motions cannot be reliably modeled by thin plate approximation, so they are not considered in the modelling or in geodetic data.

Geodetic data set includes 190 vector velocities, as shown in Figure 4, with respect to fixed Eurasia. 33 of these geodetic velocities, grey arrows in Figures 4a and 4b, have been obtained by the Matera Geodesy Center of the Italian Space Agency (ASI-CGS) using Global Positioning System (GPS), Satellite Laser Ranging (SLR) and Very Long Baseline Interferometry (VLBI) data [*Devoti et al.*, 2001]. These data have been completed in the eastern Mediterranean with the GPS measurements for the period 1988-1997 carried out by *McClusky et al.* [2000] (black arrows, also referred to Eurasia). A description on how geodetic data of different nature, such as GPS, SLR and VLBI techniques, have been combined in order to obtain reliable velocities at each site, is given in *Devoti et al.* [2001]. The ASI-CGS solution in Figure 4a and 4b represents the residual velocity with respect to Eurasia obtained by subtracting the rigid motion of Eurasia itself expressed within NUVEL-1A reference frame. The large error ellipses in the western and central Mediterranean, especially in the Iberian peninsula and in northern sector of the Adriatic plate, is an indication that geodetic data are sparse and variable in these areas. In the eastern Mediterranean, error ellipses are provided only for the ASI-CGS solution.

Figures 4a and 4b, from west to east, highlights three different styles for the direction of the horizontal velocity field: a generally south trending direction in the Iberian peninsula and Ligurian coast of Italy, a north trending direction for southern and peninsular Italy, with a rotation from NW to NNE from the Lampedusa island (LAMP), between Africa and Sicily, to Matera (MATE) through Calabria (COSE), as shown in Figure 4a, and a rotation from NNW to SSW from eastern Anatolia to southern Aegean.

The NE direction in southern Italy agrees with the suggestion of the counterclockwise rotation of the Adriatic plate, as suggested by *Ward*, 1994 and *Anderson and Jackson*, 1987. Another major feature, such as the magnitude of the velocity, characterizes the geodetic pattern, which shows a substantial increase from the west to the east and from the north to the south.

In central and north-east Italy the motion has a large north component except in the Po plain, with the site MEDI showing a large eastward component, probably due to local tectonic effects, for example thrusts associated with the buried Apenninic chain, or to the water table [*Zerbini et al.*, 2001].

Lampedusa, Sicily and peninsular Italy, except its westernmost coastal area, show a dominant north-trending component, in agreement with the major NUVEL-1A velocity component at these longitudes.

Moving to the eastern Mediterranean, Figure 4b highlights high velocities, of the order of $\sim$ 30 mm/year, with SSW direction in the Aegean region and with west directed velocities in the Anatolian Peninsula. Figure 4b highlights the northward motion of the

Arabian plate and counterclockwise rotation of central/western Anatolia and southern Aegean, bounded to the north by the NAF and its extension into the Aegean Sea.

Active convergence between Africa/Arabia and Eurasia is modelled as shown in Figure 6 within the Mediterranean. The global kinematics of these plates is governed by the counterclockwise rotation of Africa and Arabia relative to Eurasia, inferred from the global plate motion model NUVEL-1A [*DeMets et al.*, 1994. Eurasia is fixed, as indicated by the bold triangles along the northern, northwestern and northeastern boundaries of the domain under consideration and the velocity boundary conditions are taken relative to this plate, as shown by the black arrows. The southern boundary from 10° W to 35° E moves according to the Africa/Eurasia pole (located at 20.6° W and 21° N, with velocity of 0.12 Deg/Ma [*DeMets et al.*, 1994]); velocity increases from west to east, with 3.3 mm/year in the direction of 35° W on the western most part and around 10 mm/year in the north direction in proximity of the Arabian boundary. Figure 6 shows that the Arabian plate, located between the southern boundary from 35° E to 40° E and eastern boundary south of 40° N, moves with velocities ranging between from 20 to 24 mm/year, with direction varying from south to north, from 10° W to 26° W.

Only majors faults are considered and treated aa continuous, along plate boundaries, as represented in Figure 6 by solid line, along north Africa, Calabrian arc, Malta Escarpment [*Catalanos et al.*, 2001], Apennines, Alps, Dinarides, Hellenic Arc and Anatolian faults.

The fingerprint of subduction in GPS data

The model of continental convergence shown in Figure 6 must be supplemented with the effects of subduction in the Aegean (Hellenic Arc) and southern Italy (Calabrian Arc). Within a thin plate formulation, this can be achieved by applying the appropriate horizontal velocities at the plate boundaries in such a way to simulate the effects of tectonic forces due to trench suction acting on the overriding plate and slab pull on the subducting plate [*Bassi et al.*, 1997; *Meijer and Wortel*, 1996]. This implies that the plate boundary must coincide with a boundary of the finite element mesh, which can be achieved by subdividing the model between the Eurasian and African domains, as shown in Figure 8 a and 8b. The appropriate velocity boundary conditions, simulating subduction forces, can then be applied where subduction is active, in the Hellenic and Calabrian Arcs.

The reader would find the detailed description of the forces that are active at subduction zones in another chapter of this volume, but it may be relevant to provide here a brief description of the physical process that is responsible for the appearance of the velocities along the Hellenic and Calabrian Arcs, depicted in Figures 8a and 8b by the arrows along the arcs.

Subduction denotes the penetration of oceanic lithosphere into the mantle and under another plate, named overriding plate. Specifically, in Figure 8a and 8b, along the arcs, two pieces of lithosphere, pushed by Africa, enter the mantle, with approximatively northwest and northeast directions, in the Calabrian and Hellenic Arcs, respectively. By falling down in the mantle, the two subducted plates would leave an empty space underneath the overriding plates, corresponding to the Tyrrhenian and Aegean basins, north of the two arcs. Since cavitation, or empty volume, cannot exist, the overriding plates are sucked towards the arcs, to fill the volume left by the subducting plates. In the modelling, these suction forces towards the arcs are simulated by the black arrows directed to south-east in the Calabrian Arc and to south-west in the Hellenic Arc. This process is also responsible for extra velocities applied at the subducting plates, representing the downpull of the plates toward the arcs, as shown in Figure 8b, that supplement the velocities induced by Africa in the same regions. Velocities plotted in Figure 8 are thus taken from another family of subduction models, not considered now, in vertical cross sections, as in *Giunchi et al.*, 1996 or *Negredo et al.*, 1997, 1999, where the effects of trench suction and slab pull are taken self-consistently into account.

To evaluate the slab pull effects, positive density contrasts, based on the petrological studies of *Irifune and Ringwood* [1987], have been considered. These models are purely gravitational, driven solely by the negatively buoyant subducted portion of the slab. In short, these velocities, parameterizing subduction forces, are portrayed by the black arrows perpendicular to the arcs, for the Eurasian plate, as in Figure 8a, and Africa plate, as in Figure 8b. These velocity boundary conditions yield suction and slab pull forces that agree, in magnitude, with those applied by *Meijer and Wortel* [1992] to simulate the effects of subduction in the Andes.

The remaining plate boundaries (north Africa and eastern Anatolia), where subduction is not presently occurring, are subject to free boundary conditions, where the only effects are those to due lithostatic stresses.

For the Calabrian arc, the boundary condition velocities applied at the edge of the overiding plate are those obtained by means of previous 2D dynamic models in the Tyrrhenian sea (see Fig. 4b of *Giunchi et al.* [1996]). A velocity of 10 mm/year is applied at the edge of the overriding plate (Eurasia) along the arc, and 5 mm/year at the edge of the subducting plate (Africa). These velocities are applied perpendicularly to the arc, as in Figure 8a and 8b, for the overriding (Eurasia domain) and underthrusting (Africa domain) plates, respectively.

For the Hellenic subduction, a new series of two-dimensional models, in vertical cross section, has been implemented in order to take into account the effects of the geometry of the subducted slab on velocity boundary conditions to be applied at the arcs.

This class of subduction models are based on the distribution of earthquakes [*Kiratzi and Papazachos*, 1995; *Papazachos et al.*, 2000] and on seismic tomography, the latter indicative of a deep slab below the Aegean [*Jonge et al.*, 1994; *Bijwaard et al.*, 1998; *Wortel and Spakman*, 2000]. It has been verified by Jimenez-Munt et al. (2003), that roll-back, the motion of the edge of the arc towards the subducting plate, is sensitive to to the geometry of the slab and to the depth reached by the subduction plate.

Once the relative velocities due to subduction have been evaluated along the arcs, they can be applied to the Eurasian and African plates, as shown in Figure 8a and 8b, in order to retrieve the velocity field induced by subduction that must be added to the velocity due to convergence.

Figure 15 portrays the eigenvectors and eigenvalues of the modeled and geodetic strain-rate tensor. The western, central and eastern Mediterranean, panels a-d, have been subdivided into triangles with vertices connecting the sites where the horizontal velocity components are available, as described in previous sections, in order to estimate the strain-rate, from the numerical and geodetic standpoint, and to characterize the style of deformation in the area within each triangle.

In the following panels, the eigendirections are given by two perpendicular arrows, oriented with respect to the meridian; the length of the arrow is scaled to provide the eigenvalue in units of 10^{-9} yr^{-1}, indicating the unit of 1 nanostrain/yr. Red stands for compression and black for extension, with arrows in bold representing the geodetic strain-rates and the empty arrows the numerically retrieved values. From west to east, geodetically retrieved strain-rate is compared with numerical one.

In the western Mediterranean (Figure 15a), SFER-ALAC-CAGL-LAMP, compression predominates in the NNW direction. The eigendirection relative to this compression is well reproduced by the model, between 5° W and 10° E longitude, as indicated by the triangle SFER-ALAC-LAMP, where the geodetic and modeled eigendirections agree within the standard deviation σ, represented by the grey surface surrounding the eigenvectors of the geodetic strain tensor. The modeled eigenvalue is a factor two lower than the geodetic one, indicating that the model underestimates the compression. This eigenvalue is well reproduced in the central Mediterranean, within the triangle ALAC-CAGL-LAMP. NNW compression in the western Mediterranean well agrees with observed stress data as indicated, in the stress map of Figure 5, by the thrust events (blue bars) in north Africa and in the western part of Sicily. This compression is clearly induced by the relative motion of Africa versus Eurasia [*DeMets et al.*, 1994]. South of LAMP, the geodetic compression rotates by 90° with respect to the compression in the western Mediterranean and is not reproduced by the model, which severely underestimates the accompanying extension.

In the region from LAMP to MATE, the ENE compression is now well reproduced by the model. North trending extension is underestimated in the modelling, except for the triangle with vertices in LAMP and NOTO, where best fit is obtained, in terms of magnitude and direction of both compression and extension. Changes in strain style from LAMP to the north-east are evident also in the stress data of Figure 5, where, from the eastern part of Sicily to the Calabrian Arc, thrust-compression changes into normal fault-extension: this change is well reproduced by geodetic strain-rate and, to a lesser extent, by modelling. A wide zone of strike slip events in WSM2000 well correlates with the 90° rotation in the eigendirection from west to east with respect to LAMP.

In the Iberian peninsula, modelling and observation are in complete disagreement, which sugsests that some major tectonic features are not properly modeled in the westernmost part of the studied domain or that the quality of geodetic data is poor. Several competing models have been proposed to explain the geodynamic evolution of the region, including escape tectonics, subduction and slab retreat, lithosphere-mantle delamination, orogenic collapse, etc (e.g. Platt and Vissers, 1989; Royden, 1993; Zeck, 1996; Seber et al., 1996; Marotta et al., 1999). Until now, however, there is no consensous on possible active mechanisms, since their numerical modelling has produced unsatisfactory results.

Eigendirections of modeled and geodetic strain-rate are best reproduced within the triangle SFER-CAGL-LAMP, eventhough the dominant compression is underestimated. In the eastern part of the studied area the nature of the fit improves, with generally well reproduced eigenvalues but with some deviation in eigendirections. In the triangles located eastward with respect to the Calabrian Arc, compression rotates by 90° with respect to the western Mediterranean, becoming roughly perpendicular to the Arc, and appears as a surface fingerprint of subduction.

In the center of the Tyrrhenian sea (Figure 15b), CAGL-UNPG-NOTO and NOTO-UNPG-COSE, geodetic and model strain-rates are in close agreement, in terms of both eigendirections and eigenvalues. In proximity to this region, within UNPG-MATE-COSE, modeled compression is aligned with geodetic extension, due to the mismatch between modeled and geodetic velocity direction of MATE. Geodetic EW extension of UNPG-NOTO-MATE, not reproduced by the model, agrees well with the extensional tectonics perpendicular to the Apenninic chain, indicated by the normal fault events, yellow bars, appearing in the WSM2000 map in Figure 5. Observed extension, perpendicular to the Apenninic chain, indicates that subduction is also active underneath the central Apennines, a process that has not been parameterized in the modelling. This would explain the failure of the model to reproduce geodetic and WSM2000 extension.

In proximity to the Calabrian arc, the principal strain rate is extensional (UNPG-COSE-NOTO), with complete coherence between geodetic and modeled eigenvalues and eigendirections, which are roughly perpendicular to the arc, probably indicating roll back of the arc itself. This pattern is in agreement with the radial extension stress regime proposed by *Rebaï et al.* [1992] which appears also in the WSM2000 map in the Calabrian Arc region, indicated by the yellow bars parallel to the arc. Geodetic strain-rate within the pentagon GRAS-TORI-UNPG-NOTO-CAGL portrays NW compression as dominant tectonic mechanism, which changes into dominant ENE extension in the triangles CAGL-NOTO-LAMP and UNPG-COSE-NOTO. The model reproduces well all these features, except the high ENE extension in the triangle CAGL-NOTO-LAMP. The dominant NW compression in the pentagon above is also evident in the WSM2000 map, as indicated by thrust events in western Sicily and in the Ligurian coast of Italy, blue bars.

Moving to the north, in northern Italy and in the Alpine front, a deterioration in the quality of the geodetic strain is observed, as shown by the large errors. Geodetic and modeled strain-rate values are subject to a substantial reduction with respect to the southern values, in agreement with the reduction of deformation from south to north, a consequence of the increase in the distance from the Africa-Eurasia collision front and subduction zones. Eigendirections are generally well reproduced; the fit is poor for both TORI-VENE-UNPG and VENE-GRAZ-UNPG, with a 90° mismatch in this second triangle in eigendirections, and modelling predicting negligible strain-rates in the first triangle. This mismatch may be due to limitations of the model, or to the quality of the geodetic strain, both possibly related to the small size of strain-rates in the area. There may also be effects associated with the hydrological cycle of the crust, not considered in the modelling. In the triangle TORI-VENE-UNPG, geodetic compression well agrees with thrusts events, blue bars, of the WSM2000 stress map. Within the pentagon ZIMM-WTZR-GRAZ-VENE-BZRG the style of the compressive strain-rates is well reproduced by the modelling, both in eigendirections and eigenvalues. Geodetic data and geophysical model agree in fact with the WSM2000 map that portrays a cluster of thrust events in the region corresponding to the triangle WZTR-GRAZ-VENE. A mismatch between geodetic and modelled strain-rates in noted within the two triangles VENE-GRAZ-UNPG and GRAZ-MATE-UNPG: in the first triangle west directed extension is predicted rather than compression and, in the second one, EW extension is not satisfactorily reproduced.

In the Aegean region. Figure 15c, a general improvement in the coherence between the eigendirections and eigenvalues is obtained from geodetic data and from numerical modelling in comparison to Figure 15a and 15b. The dominant mechanism is NNE extension, compatible with north-trending normal faulting deformation, which is in fact the major feature of stress pattern in the Aegean region, Figure 5. This extension can be viewed as the surface expression of suction forces induced by the negatively sinking slab in the Aegean. Details of such a widespread extensional pattern, show that the largest deviation in the eigendirection occurs in the triangles LEON-7515-TWR in Greece and CAMK-BURD-7512 in western Anatolia, with a mismatch of about 90°, and in SOXO-7510-7515, with a mismatch of about 30°. Except for these three triangles, geodetic and modeled strain-rate eigendirections are in good agreement, both in Greece and western Anatolia. The eigenvalues are also in fairly good agreement, but it should be highlighted that when mismatch occurs, numerical models have the tendency to underestimate the geodetic extension, as for example in the triangle 7520-NEZA-LOGO in eastern Greece or in the two triangles in western Anatolia, D7DU-CAMK-CEIL and D7DU-BURD-CAMK. Difficulties are also encountered in reproducing the isotropic extension for SOXO-7510-7515. In general, a good agreement is obtained between eigendirections and largest eigenvalues.

Moving to the east, the NNE extension in Greece and Aegean regions shows the tendency to rotate to NE in Anatolia; in concert with this rotation in the eigendirection of the extension, compression at right angles with respect to the previous WNW direction, becomes the dominant mechanism in eastern Anatolia. Drawing a parallelism with the major driving tectonic mechanism of extension in the Aegean region, the increase in compression in the east of Anatolia well agrees with the concept that push from the Arabian plate is a major controlling mechanism in the easternmost part of Anatolia. The peculiar pattern of extension in the Aegean and compression in Anatolia is largely due to the combined effects of suction induced by deep Aegean subduction and push of Arabia. Predominant extension in the Aegean Sea and increasing compression to the east explain the tectonic regime observed in the WSM2000 map, with normal faulting in the Aegean and strike slip in the east.

Close to the Hellenic Arc, modelled extension parallel to the arc overestimates the geodetic one and, on the contrary, considerable geodetic compression perpendicular to the arc, especially in its western part, is underestimated by the model. North of Crete, compression in the geodetic strain perpendicular to the arc and extension parallel to it is noticeable, in agreement with the numerically modeled eigendirections (LEON-OMAL-TWR; TWR-ZAKR-7512). Modelling reproduces the compression perpendicular to the arc south of Greece (LOGO-LEON-OMAL), but not north of Crete. Extensional strain-rate regime parallel to the arc resulting from the modelling is visible also in the WSM2000 map. This case is mentioned as a situation in which there is a better agreement between modelling and stress data, than between modelling and geodetic data or between geodetic and stress data. In this arc region, the worst fit occurs in the westernmost part of Crete (LEON-OMAL-TWR), where predominantly WNW extension is obtained from modelling rather than NNE geodetic compression. East of Crete, in the triangle TWR-7512-ZAKR, modeled extension is consistent with observations.

Deviations between eigendirections are observed west of the Peloponnesus and east of Crete, certainly due to edge effects of the Hellenic subduction and thus to difficulties encountered by the two dimensional modelling in reproducing the real three dimensional tectonic structures at the edges of the subducting plate. In spite of this limitation, it is remarkable that the intensity of the geodetic strain-rate is well reproduced, denoting compression directed outward from the subduction zone and extension along the hinge line, where the subducting plate encounters the Earth's surface.

The largest geodetic and modeled strain-rate eigenvalues of about 10^2 nanostrain yr^{-1}, in north-western Anatolia close to the NAF, agree well with the geodetic strain rate obtained by *Ward* (1988), in the Aegean - Anatolian region. The pattern of geodetic and modeled strain release by *Ward* (1998) is not comparable in detail with ours, due to the larger set of geodetic data considered and to the higher spatial resolution herein with respect to that study.

Figure 15, panel (d), portrays geodetic and modeled strain-rates for Anatolia. Modeled eigendirections are best reproduced in the western part, with a rotation from NNE (CAMK-D7DU-BURD) to NE (AGOK-SIVR-7587), from a longitude of 28° to 32°, in good agreement with the largest geodetic eigenvalues denoting extension. Geodetic and modeled strain-rates are in complete agreement with the cluster of normal fault events in western Anatolia, denoting extension in the NNE direction, as shown by the WSM2000 map, Figure 5. From west to east, it is remarkable that, at least for the largest eigenvalue, dominant extension changes into dominant compression, both in modelling and in geodetic data.

An intermediate zone of reduced strain-rate is visible, centered approximately at 33° E, in which the dominant NE extension in the west, although reduced, changes into NE compression in the east. Modelling reproduces well the reduction in geodetic strain-rate eigenvalues, but in the triangle SIVR-7585-7580, in the center of Anatolia, the model eigenvectors are rotated by 90° with respect to the geodetic ones, a negative result which is not unexpected, due to the low strain-rate level. In the center of Anatolia the model thus reproduces well the reduction in strain-rate, and the transition from extension in the west to compression in the east, but fails to reproduce the eigendirections in the center of this zone. This strain-rate reduction is well imaged by the WSM2000 map. In the eastern part, NNW compression is the dominant mechanism, in agreement with the important push of Arabia. The magnitude of the largest eigenvalue is generally overestimated by the modelling, clearly controlled by the push of the Arabian plate. The model fails to simulate the extension observed in the data, since it has the capability to reproduce the largest eigenvalues but is has difficulties to reproduce the smallest ones.

In the eastern part of Anatolia, the geodetic eigendirection is rotated counterclockwise with respect to the modeled one. This discrepancy in the eigendirections could be due to edge effects, since this region is close to the boundary where the velocity conditions are applied, as shown in Figure 6. Comparison with WSM2000 map is not as robust as in western Anatolia, since it is based on a single stress datum in the easternmost region denoting thrusting, with an eigendirection roughly in agreement with the geodetic observation and modeled NE compression.

Stress pattern in the Mediterranean

From geophysical and geological techniques, it has been determind the approximate nature, being compressional of extensional, and orientation of tectonic stress acting in Europe and Mediterranean: compressive horizontal principal stress orientations and local tectonic regimes have been in fact compiled from the World Stress Map (WSM2000, *Mueller et al.* [2000]; Figure 5). Four different types of stress indicators, such as earthquake focal mechanisms, well bore breakouts and drilling induced fractures, in-situ stress measurements and young geological data, have been considered. These data generally show that one of the principal axes of the stress tensor is approximately vertical. Stress orientation is then defined by the azimuth of one of the horizontal principal stress axes. 1384 principal stress directions have been considered to draw a comparison with modelling and geodetic strain-rate eigendirections.

In western and central Mediterranean and northern Europe, west of 15° E longitude, the maximum compressive horizontal stress is NNW directed, parallel to the red bars and to the relative displacement between the European and African plates, as in Figure xx, except on arc structures such as western Alps and Gibraltar arcs, where small stress deviations can be appreciated. This pattern differs from what observed in central and eastern Mediterranean, where stress field presents several deviations at collision zones associated with large-scale faults and mountain belts as well as at active subduction zones, beneath the Calabrian and Hellenic arcs.

The tectonic regime along the plate boundary in north Africa and in a large part of Sicily reflects the convergence between Africa and Eurasia, showing a dominantly NW compressive trend. In the Calabrian arc, stress regime is complex, being characterized by large variability in orientation and depth as well as in the deformation style. According to *Rebaï et al.* [1992], stress pattern is characterized by radial extension.

In the southern Apennines normal and strike-slip faulting prevail, with extension perpendicular to the chain, as indicated by the yellow bars parellel to the chain [*Frepoli and Amato*, 2000]. In terms of tectonic regime, the northern Apenninic belt shows a clear distinction between an area of extension in the inner portion of the belt and an area under horizontal compression or transpression along the Adriatic margin [*Ward*, 1994; *Frepoli and Amato*, 2000]. Northern Italy coincides with the Alpine orogenic belt and is subject mostly to compression. [*Rebaï et al.*, 1992].

In the whole Europen and Mediteranean domains, the general style of stress is thus characterized by general compression and extension, the latter perpendiculat to the Apennines and spread in the Aegean.

Within Anatolia, stress direction undergoes a progressive counterclockwise rotation from a NE trending compression in eastern Anatolia to a northerly directed extension in western Anatolia. This stress pattern is consistent with the westward movement of the Anatolian Peninsula, pushed away from the collision zone along the north Anatolian right-lateral strike-slip fault (NAF) to the north, and east Anatolian left lateral fault (EAF) to the east [*Kahle et al.*, 2000]. It is also being pulled towards the Aegean by suction forces associated with subduction, well in agreement with previous strain-rate geodetic and modlelling results. Extensional tectonics along the Aegean back-arc basin and western Anatolia indicates that coupling between the African plate and the Anatolian block is weak [*Rebaï et al.*, 1992].

Blending seismic, GPS and stress data

Seismic strain provides a major constraint for plate deformation properties, since it gives the amount of strain released by the sequence of elastic shots associated with the earthquakes. The estimate of seismic strain is based on the scalar moment M_0

of the earthquakes, related to the seismic part of the strain [*Kostrov*, 1974]. Figure 3b shows the geographical distribution of earthquaks with magnitudes M_s ranging between 2,8 and 8, superimposed on the seismic strain-rates, in units of s^{-1}, obtained from the methodology described by *Kostrov* [1974] which gives a measure of the brittle deformation according to

$$\dot{\epsilon} = \frac{1}{2\mu V \Delta t} \sum_{n=1}^{N} M_0^n \tag{38}$$

where $\dot{\epsilon}$ is the strain-rate, V is the deforming volume, μ is the shear modulus and M_0^n the seismic moment of the n-th earthquake from the N total earthquakes occurring during the time interval Δt. Earthquake data have been compiled from the NEIC for the period between 1903-1999. The seismic moment has been calculated according to *Ekström and Dziewonski* [1988], using the surface magnitude, M_s

$$logM_0 = \begin{cases} 19.24 + M_s & \text{for } M_s < 5.3 \\ 30.2 - (92.45 - 11.4M_s)^{1/2} & \text{for } 5.3 \leq M_s \leq 6.8 \\ 16.14 + \frac{3}{2}M_s & \text{for } M_s > 6.8 \end{cases} \tag{39}$$

The study area includes a total of 1112 seismic events with M_s between 2.8 and 8 (Figure 3). Each earthquake involves a strain-rate effect that follows a Gaussian function, as shown in [*Jiménez-Munt et al.*, 2001a]. The most appropriate value for the Gaussian width in this area yields a value of 150 km, representing the diameter of the circular area around the earthquake where the elastic effects of the earthquake rupture are felt. Figure 3b shows the seismic strain-rate and the epicenters of the earthquakes included in these calculations. The largest seismic strain-rate release is of the order of $10^{-16} - 10^{-15}$ s^{-1}, occurring, from west to east, along the plate boundaries in north Africa, southern and north-eastern Italy, along the Alps and Dinarides, in the Aegean region, western and eastern Anatolia. Except for a localized maximum in Algeria, this seismic strain-rate pattern is characterized, due to the combined effects of the real earthquake distribution and of the Gaussian distribution, by a peculiar pattern portraying a region of $10^{-16} - 10^{-15}$ s^{-1}, embedding north-eastern Italy, Dinarides, Aegean and western Anatolia, surrounded by a region of lower strain-rate release, of 10^{-17} s^{-1}. The pattern of Figure 3b is consistent with that of the seismic moment rate density, in units of N m yr^{-1} m^{-2}, depicted in Figure 3 of *Ward* [1998], also based on the NEIC catalog.

Figure 15 shows that, once converted into strain-rate in time units of seconds rather than years, as in Figure 3, obtained by dividing teh results of Figure 15 by 3.153×10^7, denoting the number of seconds in one year, a level of about 10^{-15} s^{-1} is obtained in both modeled and geodetic strain-rates. Modelling is thus coherent with geodetic findings, but model and geodetic strain-rates are higher than seismic ones of Figure 3. This general finding should not be surprising, since geodetic strain is the sum of the deformation released by earthquakes and of that due to ductile creep of the crust and lithosphere, which may result to be larger than seismic deformation alone. Another cause for seismic strain being generally lower than geodetic and modeled ones could stand on the short time interval of 100 yr spanned by the seismic catalogue, once compared with the characteristic time scale of rheological modelling which is apropriate for 10^3 - 10^6 yr.

Some general features in modelling and geodetic strain-rate patterns, Figure 15, are well correlated with seismic strain-rate, Figure 3, such as the high values in southern Italy, decreasing to the north, the high strain-rate in the Aegean, decreasing in central Anatolia and increasing again near the EAF.

Coherence between modeled, geodetic and seismological strain-rates, and stress patterns, reveals the capability of geophysical modelling to reproduce plate deformation characteristics in terms of the mechanical properties of the crust and lithosphere, and to provide estimates of the release of seismic deformation.

CO-SEISMIC AND POST-SEISMIC DEFORMATION

Introduction

Faulting in the lithosphere causes an instantaneous deformation of the Earth's surface, called co-seismic deformation. In response to co-seismically induced elastic stress in the mantle, slow creep occurs in this deep parts of the planet, or in the viscous portion of the crust, via viscoelastic stress relaxation, as required by the dashpot sketched in Figure 1. Just after the earthquake, stress release due to viscous flow in the viscous parts of the Earth starts to operate, leading to a mechanism called post-seismic deformation. The latter is thus associated with the delayed deformation of the lithosphere caused by stress relaxation in the mantle or in the low viscosity layers of the crust.

This section deals with the effects of a realistic PREM-based (Dziewonski and Anderson, 1981) lithosphere and mantle stratification on post-seismic deformation. Relaxation is here considered to occur within the mantle, while in the following effects of stress relaxation in the viscous portion of the crust will be considered.

Modelling co- and post-seismic plate deformation

The analysis caried out in this section is key to the understanding of plate-mantle interaction, especially for the correct interpretation of GPS and DInSAR data of plate deformation, when earthquakes occur within the plate or at its boundaries. Improvements in satellite differential radar interferometry (Massonnet *et al.*, 1993) or in precise GPS monitoring of crustal motions necessitate in fact realistic and precise models of co-seismic and post-seismic effects, which must include sphericity, self-gravitation and stratification of the crust, lithosphere and mantle, as outlined in the present analysis based on a completely analytical approach, based on the mathematical formulation outlined in section XX, once the appropriate expression of the body force equivalents normal mode solution. These body force equivalents, whose explicit expression have been first given by Smilie and Manshina XXXX, can be demonstrated to induce in the Earth's material the same dynamic effects of the earthquake fault, as prescribed by dislocation theory.

With respect oto PGR, normal mode summation, in co-seismic and post-seismic modelling, must go up to harmonic degrees of the order of 10^3, one order of magnitude higher than in post-glacial rebound, as the peaks of the total strength of the modes as a function of zonal degree can be situated at zonal degrees of several thousands for shallow earthquakes at plate boundaries between the African and Eurasia domains, as considered herein.

With respect to post-glacial rebound, the toroidal solution must also be considered, as described in Sabadini and Vermeersen (2004). The procedure for solving post-seismic problems in viscoelasticity, as shown herein, was first described by Sabadini *et al.* (1984a).

Figure 8.1 deals with the relaxation times as a function of harmonic degree and shows that these relaxation modes form nice continuous-like patterns with increasing harmonic degree. The model used in this calculation is a 10-layer model, averaged from PREM, as detailed in Table 8.1, in which the uppermost mantle is stratified into four viscoelastic layers, with varying density and rigidity but uniform viscosity of 10^{21} Pa s; the transition zone and lower mantle are homogeneous.

The fastest transient viscoelastic modes superimpose at the bottom of the scale. It should be noted that not all the modes are important since, for example, the top four modes for each harmonic degree in Figure 8.1, which are the internal-mantle buoyancy modes, have extremely long relaxation times, some even exceed the age of the Earth, so that they do not play any role in post-seismic deformation.

In order to elucidate the influence of lithospheric and mantle stratification on post-seismic deformation, the following figures portray displacement patterns for a vertical, point-like, dip-slip source embedded at 100-km depth at the base of the lithosphere; the seismic moment of the source is fixed at 10^{22} Nm, characteristic of a large earthquake. Radial displacement fields are sampled at an azimuth of 90° with respect to the strike of the fault, in the subsiding portion of the Earth's surface; tangential displacements are sampled at 45°. At 180° from these directions, the same displacement patterns would be obtained, but with the reversed sign. Figure 8.2 shows, for a fixed distance of 200 km from the epicenter of the fault and varying time after the occurrence of the earthquake, the radial and tangential displacement components, (a) and (b) respectively. Remarkable differences can be observed among the 4- and 10-layer volume-averaged models and model P1 as used in previous analyses by Sabadini *et al.* (1995), in which the rigidity and density parameters were taken by imposing specific density contrasts at internal interfaces rather than averaging the density and elastic parameters from PREM, as done here. The parameters for the 4-layer volume-averaged model are given in Table 8.2.

Table 2. Parameters for the 10-layer volume-averaged earth model. r is the distance with respect to the center of the Earth, ρ the density of the layer, and μ the rigidity. The viscosity ν is given for the convex profile models, as shown in Sabadini and Vermeersen (2004).

Layer	r (km)	ρ (kg/m^3)	μ (N/m^2)	ν $(Pa\ s)$	
1	6371 – 6356	2283	2.66×10^{10}		
2	6356 – 6331	3194	5.91×10^{10}		elastic lithosphere
3	6331 – 6251	3372	6.77×10^{10}		
4	6251 – 6221	3372	6.69×10^{10}	1.5×10^{20}	
5	6221 – 6151	3372	6.61×10^{10}	2.5×10^{20}	uppermost mantle
6	6151 – 6061	3462	7.56×10^{10}	4.5×10^{20}	
7	6061 – 5971	3515	7.89×10^{10}	7.0×10^{20}	
8	5971 – 5701	3857	1.06×10^{11}	3.0×10^{21}	transition zone
9	5701 – 3480	4878	2.19×10^{11}	2.4×10^{22}	lower mantle
10	3480 – 0	10932			inviscid fluid core

Table 3. Parameters for the 4-layer volume-averaged earth model. r is the distance with respect to the center of the Earth; ρ the density of the layer, and μ the rigidity. The viscosity ν is given for the convex profile models.

layer	$r\ (km)$	$\rho\ (kg/m^3)$	$\mu\ (N/m^2)$	$\nu\ (Pa\ s)$	
1	$6371-6250$	3234	5.99×10^{10}		elastic lithosphere
2	$6250-5701$	3631	8.60×10^{10}	1.6×10^{21}	upper mantle
3	$5701-3480$	4878	2.17×10^{11}	2.4×10^{22}	lower mantle
4	$3480-0$	10932			inviscid fluid core

For radial displacement, panel (a), the volume-averaged models predict an increase in post-seismic deformation, while model P1, short-dashed curve, predicts the opposite behavior, namely reduction in post-seismic signals. Although similar to the 10-layer model in the amplification of the post-seismic deformation, the 4-layer model, portrayed by the solid curve, predicts, both in the elastic and long time-scale limit, a radial displacement which is higher by a factor two in comparison to the 10-layer models, as shown by the dashed curve. The behavior of the old P1 model is due to the absence of elastic stratification between the lithosphere and the mantle, responsible for an overestimated lithospheric rigidity. Deviations in the 4-layer model with respect to the 10-layer one are due to the absence of stratification in the elastic lithosphere in the 4-layer model. These findings indicates the importance in correctly taking into account the realistic stratification of the lithosphere in dealing with global post-seismic deformation.

The reader should not be surprised of these results, since the final state of the Erth's model, after relaxation has taken place in the mantle, depends solely on the elastic properties of the lithosphere. It has been verified that further refinement in lithospheric layering does not modify these findings with respect to the 10-layer model. The largest rigidity and density contrasts within the lithosphere are located at the base of the upper and lower crust, with the remaining portion of the lithosphere essentially homogeneous, which guarantees that a three-layered lithosphere, as done here, is sufficient to provide realistic estimates of the effects of earthquakes on global co-seismic and post-seismic deformation.

In panel (b) of Figure 8.2, the ame kind of analysis is carried out for the tangential displacement, in the outward direction with respect to the fault, at 45° from the strike of the fault. On analogy with the findings from panel (a), the tangential displacement of P1 model is also subject to a substantial reduction during post-seismic deformation. This is in distinct contrast with the results of volume-averaged models, which show a small reduction in the signal with respect to the co-seismic deformation. Comparison between the 4- and 10-layer models reinforces the conclusion drawn from panel (a) on the importance of stratifying the outer elastic layer of the Earth for realistic estimates of post-seismic displacements.

Figure 8.3 shows the displacement as a function of the distance from the source for the same Eart's models portrayed in Figure 8.2, but for fixed time intervals after the earthquake. 4- and 10-layer viscosity models are averaged from the convex viscosity profile of Figure 2.3, and viscosity values can be found in Tables 8.1 and 8.2. The panels depict three snapshots of the deformation, with panel (a) depicting the elastic limit, panel (b) an intermediate time when post-seismic deformation is operative and the long time-scale limit. The most striking result visible in this figure is that the 10-layer model behaves in a completely different fashion from the simplified P1 and 4-layer models.

In the elastic limit, in fact, both P1 and 4-layer models overestimate the radial displacement by more than a factor of two. During the transient at $t = 10^3$ yr, the P1 model predicts a substantial reduction rather than a smooth amplification of the post-seismic signal, as predicted by the volume-averaged 4- and 10-layer models. In the far field, an upwarping of the lithosphere for the 10-layer model is obtained, not reproduced by the 4-layered model, due to the overestimated rigidity of the outer elastic layer. In the long time-scale configuration, the 10-layer model does not show a visible increase in the deformation in the near field, while in the far field this model has rebounded with an annihilation of the upwarping already noted in the transient regime. This result indicates, as expected, that mantle relaxation has a strong control of post-seismic deformation in the far field from the epicenter of large earthquakes, and thus ultimately controls plate-mantle interaction. In this final state, the old P1 model predicts a pattern which does not show any resemblance to the more realistically stratified models, providing a smooth deformation of a few centimeters, in comparison with the 50 cm at 50 km from the epicenter as predicted by the 10-layer model. Except for the far field, the drastically different behavior of the three models is due to whether the lithosphere has been stratified or not.

Figure 8.4 shows the effects of mantle viscosity stratification based on the comparison between two 10-layer models, with the same lithospheric stratification, but different viscosities, such as a uniform mantle viscosity of 10^{21} Pa s and varying one, as the convex viscosity profile of Figure xxx. The elastic and long time-scale limits of the homogeneous and convex viscosity mantle are the same, as shown by the short-dashed and dotted curves. The elastic profiles of the mantle and lithosphere being idenstical, the initial and final configurations of the Earth's models are the same, since, as already noted, also the long time state of the model is determind by its elastic structure. In proximity to the source, at distances in the order of $10 - 100$ km, the two models behave similarly, except for a 10 per cent smaller signal at 50 km for the homogeneous model caused by a higher viscosity in the upper mantle, while in the far field the homogeneous and convex mantle viscosity models produce a quite different response at intermediate time-scales of $1 - 1000$ yr. As already noted in panel (b) in Figure 8.3, displacement in the far field occurs in the opposite direction with respect to that experienced in the near field. The convex viscosity profile model predicts, in fact, positive vertical displacements of about 3 cm, while the homogeneous viscosity model does not show any visible evidence of relaxation. This result demonstrates the importance of mantle rheological stratification in controlling the deformation of the plate in the far field during post-seismic deformation. The upward deflection of the lithosphere in the far field caused by the convex viscosity mantle is due to the upper mantle, softer than the homogeneous model.

The elastic stratification of the lithosphere has thus a major influence on post-seismic deformation. Three layers in the lithosphere, at the least, averaged from PREM, are found to be necessary to obtain correct estimates of the deformation following large earthquakes, both in the near field and in the far field. Further increase in the eleven lithospheric layers of PREM produces a minor refinement in post-seismic deformation estimates in comparison with the major improvement that is gained from one to three lithospheric layers. Modifications in deformation patterns during the time evolution from the initial elastic state to the long time-scale one, is mainly controlled by mantle rheology. Viscoelastic stratification of the mantle has in fact major influences on the rebound of the lithospheric plate in the far field following the earthquake. In comparison with the required stratification of the lithosphere, the same fineness of layering is not needed for the mantle, in agreement with the result that mantle rheological stratification controls the global pattern of post-seismic deformation in the far field.

Co-seimic deformation in DInSAR analyses

Post-seismic Deformation

In the previous section large and deep earthquakes were considered in order to elucidate the global properties of post-seismic deformation by considering, in particular, stress relaxation in the mantle. Normal mode theory is now applied to analyze stress relaxation in the lower crust. By dealing with shallow sources and rheologically stratified crusts, as done herein, normal mode expansion must be performed to high harmonic degrees to correcly simulate post-seismic displacements for earthquakes whose faults cut a thin upper elastic crust overlying viscoelastic layers. Even higher resolution is necessary for co-seismic calculations for shallow sources, so that summation over 4×10^4 spherical harmonics must be performed, while for estimating rates of deformation, due to stress relaxation in the lower crust, summation over spherical harmonics of a few times 10^3 is adequate. The reader can find details for implementing improvements in normal mode theory for dealing with such high degrees of 10^4 in Riva and Vermeersen (2002), who first developed the necessary mathematical tools. Modelling results shown in this sections represent a novelty, in terms of both the large number of layers dealt with, first made possible by the grid-spacing procedures to detect the normal modes in the Laplace doman described in Vermeersen and Sabadini (1997), representing the real heart of this new family of post-seismic deformation models, and by the high spatial resolution of viscoelastic solution described in Riva and Vermeersen (2002).

The example of the Umbria-Marche (1997) earthquake

Stress relaxation process in the lower crust, is here studied via the Umbria-Marche 1997 moderate-size earthquake, characteristic of slowly deforming plate boundaries in central Mediterranean, such as the one in central Italy, between the Tyrrhenian and Adriatic domains. Central Italy is undergoing continental extension, as anticipated in section xx when dealing with tectonic forces applied at plate boundaries, and experienced the moderate Umbria-Marche 1997 normal faulting earthquake sequence. Deep seismic reflection studies (CROP03) and the 1997 earthquake sequence clearly show a seismogenic layer decoupled from the lower crust by a sizeable transition zone. In accord with these observations, normal mode modelling is based on a three-layer crust divided into an elastic upper crust, a transition zone and a low-viscosity lower crust, as shown in Figure 8.5, in agreement with nearly all viscoelastic relaxation studies

(Ma and Kusznir, 1995; Pollitz, 1996), assuming the brittle upper crust (UC), the lower crust (LC) and the mantle.

Seismological and geological observations from regions of active faulting suggest that the seismogenic UC could be separated from the LC by an extra layer, named transition zone (TZ), representing the place of occurrence of both brittle and ductile processes (Sibson, 1982; Meissner and Strehlau, 1982; Chen and Molnar, 1983, Sibson, 1989; Scholz, 1990). These observations are frequent in regions of active continental extension (Jackson and White, 1989). The structure of the crust of Figure 8.5 agrees with seismic reflection profiles (Pialli *et al.*, 1998), purposely designed to study the deep crust in the northern Apennines. Below the Umbria-Marche extensional belt, the top of the LC is placed at a depth of 20 km, while the Moho is at 35 km (Coli, 1998). Cattaneo *et al.* (2000)

An abrupt cutoff of the 1997 aftershock sequence is placed at about 9 km of depth, which seems to indicat that seismic activity is largely restricted to the uppermost 10 km of the deforming continental crust and leaves room for a 10 km-thick TZ above 15 km of LC as imaged by the seismic reflection profiles (Figure 8.5). The TZ acts in the normal mode modelling as a layer which decouples the elastic UC from the low-viscosity LC. Modelling is thus based on the five layers shown in Figure 8.5. This stratification is also consistent with the range of regional velocity models estimated via surface-wave dispersion analyses, as done in Calcagnile and Panza (1981) and regional gravity and heat flow anomalies (Della Vedova *et al.*, 1991; Marson *et al.*, 1995) and other relevant geological and geophysical data (Du *et al.*, 1998).

The Umbria-Marche earthquake sequence started on September 26, 1997, takng place in a complex deforming zone along a normal fault system in central Apennines. Only the strongest earthquake of the Umbria-Marche sequence that took place on the 26 September 1997 at 9:40 (Ms = 6.0) is considered in the modelling, with seismic moment estimat based on the solution in Saraó *et al.* (1998) retrieved from broad-band waveform inversion, in good agreement with the CMT solution (Ekstrom *et al.*, 1998). The application below provides a first-order estimate of the expected vertical velocities caused by post-seismic relaxation in the area.

To illustrate first the sensitivity of the results on source's depth, the finiteness of the source in the vertical direction is neglected, and besides the source depth of 7 km, in agreement with the seismological solution (Saraó *et al.*, 1998, Aoudia *et al.*, 2003), a case in which the source is embedded at a depth of 3 km is considered, in order to simulate the effects of a fault almost reaching the Earth's surface. Seismic moment is dstributed along a line of dislocations whose length along strike is 10 km. The source has thus the infinitesimal extension of a point in the down dip direction. The dip of the fault is 37°, in agreement with the seismlgical solution by Saraó *et al.* (1998). In order to resolve the deformation produced by this shallow fault, summation of 4000 normal modes is performed, in order to ensures convergence of the normal mode solution. Vertical deformation is considered, being most relevant for normal faulting simulations and results are computed at the Earth's surface along a line perpendicular to the strike of the fault and crossing its center. The fault is dipping to the left, and the distance is measured from the intersection of the upward prolongation of the fault with the Earth's surface. In Figure 8.6, the depth is fixed at 7 km.

Figure 8.6a shows the modelled co-seismic and post-seismic vertical deformation, measured at the surface. The co-seismic component, indicate by solid line, shows the deformation pattern characterizing normal faulting, with uplifted, localized footwall and broad subsidence in the hanging wall. The post-seismic component, indicated by the dashed line, accounting for both co-seismic deformation and post-seismic viscoelastic relaxation, is responsible for a broadening of the subsidence, affecting the footwall. Post-seismic displacements are evaluated after complete viscoelastic relaxation. The largest co-seismic displacement is -8 cm 10 km from the fault. Due to visoelastic relaxation, the maximum displacement increases to -12 cm, and a broad area, which remained at rest during the co-seismic deformation, is now subject to subsidence at distances of tens of kilometers, due to the decoupling of the uppermost part of the crust with respect to the lower layers. Normal faulting occurrence and extension at the bottom of the UC, below the neutral plane of the uppemost elastic layer, are responsible for down flexure of this elastic layer, imaged by the broad subsidence in Figure 8.6a.

Figure 8.6b portrayes modelled vertical velocities at different times after the earthquake. A wide region of negative velocity marks the broad down-flexure, or subsidence, of the UC, while two regions of low velocity represent the peripheral flexural response of this plate to subsidence. The short wavelength feature at the center of the subsiding area can be recognized as a relative maximum corresponding to the footwall and the subsidence of the hanging wall, within a generalized subsidence.

The highest velocity is -0.85 mm/yr in the hanging wall at $t = 0$ and, due to the relatively high viscosity of the TZ (10^{19} Pa s), is subject to a minor reduction after 10 yr. After 50 yr, there is still a visible velocity signal of about -0.25 mm/yr. A viscosity of 10^{18} Pa s is considered for the transitioz zone TZ, thus coupled with the LC, velocity is subject to an increase of one order of magnitude, attaining the value of -8.5 mm/yr. Since the TZ and LC viscosities are largely uncertain, ranging between 10^{18} Pa s and 10^{19} Pa s, it is likely that the largest subsidence rates in the highly deforming area of the hanging wall vary in the range obtained in this figure. These findings show that comparison between ongoing GPS campaigns and model predictions is crucial for estimating the effective viscosity of the TZ and LC layers.

Figure 8.6c deals with stress relaxation in the mantle and sampling times increased with respect to previous cases. Deformation pattern is broadened by stress relaxation involving larger portions of the Earth, with substantially reduced vertical velocities, once compared with Figure 8.6b. From Figures 8.6b and 8.6c, it is remarkable that for shallow normal faulting TZ and LC play a major role in comparison to the mantle, whose role is insignificant in present analyses.

In Figure 8.7 the source is embedded at the shallow depth of 3 km, in the top half of the upper crust UC. The pattern of both co-seismic and post-seismic displacement is now sharper than in Figures 8.6, and the largest co-seismic subsidence in the hanging wall is increased by a factor of two with respect to the case of 7 km depth considered in Figure 8.6a. The uplift in the footwall, with respect to Figure 8.6a, is three.times larger. Such an increase in the amplitude of the displacement and sharpening of its pattern is due to the shallow depth of the source. With respect to Figure 8.6, it is noticeable the upward migration of post-seismic displacement pattern with respect to co-seismic displacement. Such an effect is caused by the extensional source being located above the neutral plane of the upper crust UC, which induces a bending moment opposite to the one induced by the deeper source of Figure 8.6 and thus a general uplift, rather than subsidence. From Figures 8.6a and 8.7a, depending on the location of the source beneath or above the neutral plane, post-seismic curves lie completely above or below the co-seismic one. These findings differ from those reported by Ma and Kusznir (1995) where, due to the finiteness of the source in the vertical direction, the fault cuts the whole UC through its neutral plane and post-seismic curves thus intermingle with co-seismic ones. These findings represent a clear example on the sensitivity of plate deformation properties, in particular of the Earth's crust, in seismogenic areas.

Velocity pattern in Figure 8.7b is also affected by the global upwarping observed in post-seismic displacements, in agreement with previous observations on the flexural properties of the UC, with uplift velocities of 1.2 mm/yr in the hanging wall, close to the footwall. This tendency of a general upwarping can be appreciated also in Figure 8.7c, where relaxation involves only the mantle. For Figure 8.7c the same observations about Figure 8.6c can be made, except uplift of the footwall being enhanced with respect to the subsidence in the hanging wall.

These results suggest that detection of post-seismic deformation in the Umbria-Marche area is a challenging task, due to the smallness of the expected velocity signal, of the order of a few millimeters per year, for vertical and horizontal components. In order to reveal post-seismic effects in the area, a series of GPS campaigns has been undertaken in order to detect in particular the horizontal deformation components. Results from this surveying, along a baseline perpendicular to the major fault activated during the 1997 Umbria-Marche seismic sequence, are given in Figure 8.8, where isplacements are evaluated with respect to the town of Spello, indicated by A_0. These results represent an evidence for horizontal post-seismic deformation, triggered by shallow earthquakes, in the Mediterranean region, as shown in Aoudia *et al.* (2003). Figure 8.8, redrawn from Aoudia *et al.* (2003), compares GPS baseline variations with modeled ones, as a function of the baseline length in kilometers from the reference site of Spello (A_0).

All baseline variations are positive, except $A_0 - A_1$, which indicates a general trend of post-seismic extension across the fault. Estimates obtained from Bernese and GIPSY computer codes, for GPS data reduction, are shown with one sigma error bars, together with model results corresponding to a transition zone viscosity of 10^{18} Pa s and a lower crust viscosity of 10^{17} Pa s. Different fault models, indicated by FM1, FM2 and FM3, have been considered, as indicated in the caption. FM1 (dotted) is not consistent with observations at different sites and does not reproduce the general pattern of baseline variations from analyses. FM2 (dashed) reproduces only part of the observations and the general trend, while FM3 (solid) is the best fitting one, well reproducing almost all the observations and the general baseline variation pattern. These findings demonstrate the important role of GPS survey in seismogenic zones in revealing the physics of post-seismic deformation and providing a fundamental tool to unreavel plate deformation properties in active fault areas..

This section provides a first estimate of expected rates of vertical post-seismic deformation in a characteristic seismogenic zone of the central Mediterranean. Fine rheological layering in the crust plays a major role in the interplay between relaxation in the TZ and LC with respect to relaxation in the mantle, when shallow sources are considered in normal mode relaxation theory. Unlike large and generally deep earthquakes (Thatcher and Rundle, 1979; Melosh, 1983; Pollitz and Sacks, 1997; Suito and Hirahara, 1999), shallow and moderate normal-faulting earthquakes are unaffected by lower lithospheric viscous coupling, while the TZ and low-viscosity LC impose a pattern and a spatial scale on post-seismic deformation.

For normal faulting, these applications show the relevance of the depth of the source with respect to the thickness and neutral plane of the elastic layer in controlling post-seismic upwarping or downwarping, as emphasized by Thatcher and Rundle (1979) and Melosh (1983) for thrust faults. The elastic layer being here the upper crust rather than the whole lithosphere, this section shows that results from Thatcher and Rundle (1979) and Melosh (1983) are scale-invariant and applicable to dip-slip sources embedded in an elastic layer of any thickness overlying a viscoelastic one. Furthermore, the pattern of widening and narrowing of the deformation due to stress relaxation at different depths (Figures 8.6 and 8.7) agrees with King *et al.* (1988) findings from simpler models. This section also shows that an appropriate test area for dealing with viscoelastic normal mode theory at high harmonic degrees and for studying TZ and LC relaxation effects is central Mediterranean, in peninsular Italy in particular, where a clear cutoff in the depth distribution of earthquakes indicates that the seismogenic layer is limited to the first ten kilometers of the crust and a well-developed TZ lies below this depth, above a low-viscosity LC.

The Irpinia (1980) Earthquake

Comparison of measured vertical displacements from two leveling campaigns performed in 1981 and 1985 in the epicentral area of the 1980 Irpinia earthquake ($M_S = 6.9$) and predictions from viscoelastic earth models reveals the occurrence of post-seismic deformation due to stress relaxation in the ductile part of the crust, south of the area explored in the previous section 8.2.1. In proximity of the major fault, the leveling lines show a peculiar upwarping of the crust, accumulated during the time interval 1981 - 1985.

Here we test the mechanism of stress relaxation in the ductile parts of the crust after the occurrence of the shallow, normal-fault 1980 Irpinia earthquake, following the study by Dalla Via *et al.* (2003). The joint study of local and world-wide seismological data, static deformations and geological evidences provided a detailed picture of the complex mechanism of this event (Westaway and Jackson, 1984; De Natale *et al.*, 1988; Pantosti and Valensise, 1990; Pingue and De Natale, 1993). The main event consisted of three distinct subfaults at least, ruptured at intervals of about 20 s from each other. Surface faulting linked to this earthquake was well evident at several places and in particular on the main fault (first subevent), with dislocation up to 1.2 m. The total seismic moment inferred for this event is 3×10^{19} Nm.

The asymptotic expression of the fundamental solutions of the incompressible, self-gravitating, spherical, viscoelastic Earth for high harmonic degree described in Section 1.9 makes it possible, for this 1980 Irpinia application, to sum 40,000 spherical harmonic contributions in the co-seismic part and 6,000 ones in the post-seismic part, which guarantees the attainment of the highest resolution both in the co-seismic and post-seismic components. The earth model, described in Table 8.3, consists of 5 layers including a purely elastic upper crust (UC), a viscoelastic lower crust (LC), the mantle (M) and the core (IC).

Table 4. Viscoelastic model parameters for the 1980 Irpinia earthquake (UC=upper crust; LC=lower crust; LCB=lower crust bottom; M=mantle; IC=inviscid core)

Layer	Depth	ρ	ν	μ
	(km)	(kg/dm^3)	$(Pa\ s)$	(GPa)
UC	0-18.5	2.65	∞	32.5
LC	18.5-28.5	2.75	$1 \cdot 10^{18}$	33.7
			$0.75 \cdot 10^{19}$	
			$1 \cdot 10^{19}$	
			∞	
LCB	28.5-32.5	2.90	10^{18}	35.5
			∞	
M	32.5-2891	3.39	10^{21}	73.5
IC	2891-6372	10.93	—	—

The depths of the crustal layers and their elastic values have been taken from the average depths of the seismogenic crust and MOHO in the southern Apennine area (Mostardini and Merlini, 1986), while the deeper layers are based on standard global earth models. The upper limit of the lower crust, 18.5 km, has been chosen to match the maximum depths of earthquakes in this area on the basis of the assumption that the viscoelastic, ductile lower crust inhibits seismic fracture.

Viscosity in the lower crust has been varied from typical values of 10^{19} Pa s (Pollitz *et al.*, 1998). In order to mimic the reduction of the viscosity within the lower crust and the decoupling between the lower crust and the mantle, as expected on the basis of strength reduction with depth for the continental lithosphere under extension (Lynch and Morgan, 1987; Cosgrove, 1997), the bottom of the lower crust, LCB in Table 8.3, has been reduced by one order of magnitude with respect to the normal value of 10^{19} Pa s. Due to the simplified viscosity profile within the lower crust, only the effective viscosity resulting from the volumetric average within the two viscoelastic layers characterizing the lower crust can be compared with post-seismic results from other tectonic environments (Pollitz *et al.*, 1998; Pollitz *et al.*, 2000). A standard mantle (M) of 10^{21} Pa s below the lower crust does not portray any sizeable deformation over the time-scale of post-seismic deformation.

The assumed fault system consists of three normal subfaults as shown in Figure 8.9, including the three leveling lines considered, namely CZT2 (diamonds), CZT3 (crosses) and IGM81 lines (circles), measured immediately after the main shock and four years later; the thin curve in Fig 8.9, starting from Eboli and routing to Grottaminarda through Potenza and the IGM81 profile, represents the leveling line along which the co-seismic vertical displacement has been measured (Arca *et al.*, 1983; De Natale *et al.*, 1988). The surface projections of the three faults F1, F2 and F3 are shown by the light grey; the fault parameters of this model, shown in Table 8.4, are taken from Pingue and De Natale (1993). The total seismic moment has been inferred from seismological and geodetic data; for the main fault F1 the seismic moment M_0 is fixed at 24.4×10^{18} Nm, at 2.5×10^{18} Nm for F2 and at 3.2×10^{18} Nm for F3. The slip angle is fixed at $-90°$ for each fault. Slip on the three sub-faults has been considered homogeneous in the first tests. Subsequently, by maintaining the seismic moment constant to $M_0 = 3.0 \times 10^{19}$ Nm, the slip distribution on the main fault has been varied with depth in order to reduce the misfit between model predictions and observations.

Table 5. L1: fault length. L2: fault width along slip direction. Top: depth of fault top margin. Disl.: mean dislocation. Str.: strike.

Sub Event	L1 (km)	L2 (km)	Top (km)	Disl. (cm)	Str. (^{o})	Dip (^{o})
F1(0s)	25	20	1.0	150	317	60
F2(18s)	22	14	1.0	25	310	20
F3(40s)	13	10	1.3	75	120	85

Figure 8.10 shows the observed displacements resulting from the leveling lines in Figure 8.9, where the black vertical bars reproduce the observations and their average errors; the curves represent the modeled vertical displacements due to viscoelastic relaxation without the co-seismic component, computed for various combinations of slip distribution and viscosities. In order to carry out a comparison independent of the choice of the zero in the leveling, each model result has been uniformly shifted in such a way that the mean residual vanishes. The grey curves correspond to the reference model characterized by a uniform distribution of the seismic moment over the fault and by a lower crust viscosity, between 18.5 and 28.5 km depth, of 10^{19} Pa s (Table 8.3). Although the trends shown by this model agree with some basic features of the three lines, such as the subsidence along points 1 - 20 and the following uplift of IGM81, the general uplift along CZT2 and the uplift between points 1 - 50 of CZT3, the grey curves always underestimate the observations, especially the upward bulge of 20 - 30 mm of CZT3 and the drastic increase in the uplift at point 20 of IGM81. In order to increase the upwarping of the crust along CZT3, where this line crosses the major fault, it is necessary to distribute the seismic moment in such a way that the maximum slip occurs at depths of about 10 km, as shown in Table 8.5, resulting in the dotted curves in Figure 8.10.

Table 6. Seismic moment distribution along the fault width (L2); M=$M_0/5$ where M_0 is given in the text; the fault description is given in Table 8.4.

Fault fraction	F1	F2	F3
1/5	0.8 x M	2.0 x M	2.0 x M
2/5	1.2 x M	2.0 x M	2.0 x M
3/5	1.5 x M	1.0 x M	1.0 x M
4/5	1.0 x M	0.0 x M	0.0 x M
5/5	0.5 x M	0.0 x M	0.0 x M

This best-fit seismic moment distribution has been chosen by minimizing the L^1 norm of the residuals between observations and model predictions for all the three lines simultaneously. The long-wavelength uplift between points 1 - 35 of CZT3 increases with respect to the dotted curve to 12 mm; the same is true for CZT2 for points 1 - 25, where uplift reaches 10 mm. Particularly evident is the tendency to reproduce the change from subsidence into uplift along IGM81 from the benchmarks 15 to 35. The fit between the observations and model results can be substantially improved by reducing the viscosity in the lower crust from 10^{19} Pa s to 0.75×10^{19} Pa s, as shown in Table 8.3 for the LC layer in the case of non-uniform seismic moment distribution, as indicated by the black solid curves. The increase to 23 mm in the maximum along CZT3 is accompanied by an increase to 15 mm and 10 mm along CZT2 and IGM81, respectively. The general trend of observed displacements now appears well reproduced by the best-fit model, except for some high frequency signals at very localized zones and a rather systematic underestimation of displacements in the northernmost zone of the IGM81 line.

Figure 8.11 provides an areal view of the post-seismic vertical displacement accumulated in the period 1981-1985, with characteristic zones of uplift and subsidence;

this figure corresponds to the same viscosity model of Figure 8.10, characterized by lower crust relaxation (solid curves). Two major features are notable: the broadness of the area affected by subsidence and uplift and the amplitude of the displacement, of the order of 10 and 40 mm for uplift and subsidence. These effects are indicative of stress release at depth and of the ability of the viscoelastic part of the crust to channel the flow at large distances from the fault. The expected present-day peak-to-peak post-seismic vertical displacement, 22 years after the earthquake, is 120 cm, characterized by the same pattern as in Figure 8.11.

In the case of the 1980 Irpinia earthquake area, we thus have evidence that, once averaged over the thickness of the lower crust of 14 km, the effective average viscosity of 0.6×10^{19} Pa s agrees within the order of magnitude with the 'normal' lower crust viscosity of about 10^{19} Pa s obtained by Pollitz *et al.* (1998) from post-seismic relaxation following the 1989 Loma Prieta earthquake.

The characteristic wavelength of the post-seismic deformation in the Irpinia area is supporting evidence for viscoelastic relaxation in the lower crust rather than in the uppermost part of the mantle, unlike the 1992 Landers earthquake where there are indications that relaxation also involves a weak upper mantle. An additional result from post-seismic displacement modelling in this area consists of further details about the fault slip with respect to the previous results. It appears that higher slip concentration at around 10 km depth is able to give a best fit to the observed data.