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Ductile crustal flow in Europe's lithosphere

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ABSTRACT

Potential gravity theory (PGT) predicts the presence of significant gravity-induced horizontal stresses in the lithosphere associated with lateral variations in plate thickness and composition. New high resolution crustal thickness and density data provided by the EuCRUST-07 model are used to compute the associated lateral pressure gradients (LPG), which can drive horizontal ductile flow in the crust. Incorporation of these data in channel flow models allows us to use potential gravity theory to assess horizontal mass transfer and stress transmission within the European crust. We explore implications of the channel flow concept for a possible range of crustal strength, using end-member 'hard' and 'soft' crustal rheologies to estimate strain rates at the bottom of the ductile crustal layers. The models show that the effects of channel flow superimposed on the direct effects of plate tectonic forces might result in additional significant horizontal and vertical movements associated with zones of compression or extension. To investigate relationships between crustal and mantle lithospheric movements, we compare these results with the observed directions of mantle lithospheric anisotropy and GPS velocity vectors. We identify areas whose evolution could have been significantly affected by gravity-driven ductile crustal flow. Large values of the LPG are predicted perpendicular to the axes of European mountain belts, such as the Alps, Pyrenees–Cantabrian Mountains, Dinarides–Hellenic arc and Carpathians. In general, the crustal flow is directed away from orogens towards adjacent weaker areas. Gravitational forces directed from areas of high gravitational potential energy to subsiding basin areas can strongly reduce lithospheric extension in the latter, leading to a gradual late stage inversion of the entire system. Predicted pressure and strain rate gradients suggest that gravity driven flow may play an essential role in European intraplate tectonics. In particular, in a number of regions the predicted strain rates are comparable to tectonically induced strain rates. These results are also important for quantifying the thickness of the low viscosity zones in the lowermost part of the crustal layers.

1. Introduction

Previous Potential Gravity Theory (PGT) studies show that surface topography and lateral variations in crustal thickness and composition may lead to significant gravity-induced horizontal stresses (e.g., England and Molnar, 2005; Flesch et al., 2001). However, PGT does not treat differential movements that can be produced within the lithosphere at depths where the rheological strength is small enough to result in ductile flow under gravitationally induced stresses. Although rheology has uncertainties, all conventionally assumed parameter estimates of crustal flow should take place in plates having thermotectonic ages younger than 300 Ma or surface heat flow about and higher than 60 mW/m^2 (e.g., Burov and Diament, 1995; Cloetingh and Burov, 1996). In older lithosphere the possibility of crustal flow depends on crustal composition and fluid content. For example, crustal flow is likely to happen for any wet rheology. Ductile flow in weak parts of the lithosphere has been proposed to play an important role in tectonic deformations of the continents and might drastically reduce the flexural rigidity of the lithosphere (Avouac and Burov, 1996; Bird, 1991; Burov and Cloetingh, 1997; Cloetingh et al., 2005), Fig. 1. The effectiveness of this process depends to a large extent on the rheological stratification of the lithosphere and its density structure. Both are far from uniform in the continental lithosphere, depending, for example, on the age of tectonic structures and the thermal state of the lithosphere. These features can lead to strong spatial variations in the degree of mechanical decoupling between crust and mantle, as shown by reduction of the observed equivalent elastic thickness of the lithosphere compared to predictions based on assumption of strong non-stratified rheology (e.g., Burov, 2011; Watts and Burov, 2003). Multiple studies of the equivalent elastic thickness, T_e, in Europe (e.g., Pérez-Gussinyé and Watts, 2005) reveal strong spatial variations of T_e, ranging from higher values corresponding to rheologically coupled lithosphere with strong crust (T_e>60 km), specifically in the old blocks to the

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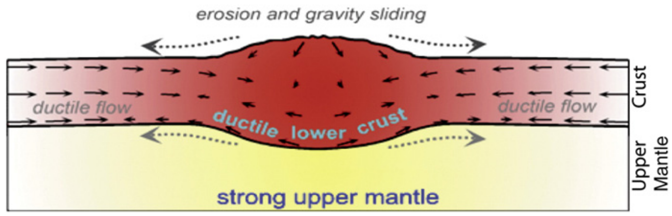

Fig. 1. Concept of gravitational crustal flow (after Avouac and Burov, 1996): topographic and crustal thickness variations result in pressure differences that may drive lateral crustal flow.

north/north-west of the Trans European Suture Zone, to considerably smaller values south of this zone (T_e<40–50 km), indicating the possibility of crust–mantle decoupling (Burov and Diament, 1995). Indications for ductile flow come also from observed deformations, such as those inferred from seismic anisotropy (e.g. Meissner et al., 2002). The latter is generally aligned parallel to the structural axes of the mountain belts (e.g. Hearn, 1999). This alignment has been interpreted as indicating deformation in the uppermost mantle, caused by mountain-perpendicular compressive stresses (Meissner et al., 2002). This compression might contribute to crustal thickening and, under appropriate conditions, to lateral escape of middle and lower crustal material (and presumably weak mantle material) in the direction of the weakest part of the belt. Young mountain belts have thick crust, higher than average heat flow, and higher upper mantle temperatures (Artemieva and Mooney, 2001), which might be a consequence of additional heat from friction or melt associated with deformation (e.g. Chapman, 1986). Recent models of orogenic belts (Whittington et al., 2009) show that the decrease of thermal diffusivity with depth leads to retention of heat generated by strain and radiogenic heating or other sources, and favors melting and flow channel formation in the lower crust, even without unusually high radiogenic heat production at mid-crustal levels. Lower crustal flow has been proposed to explain crustal thickening and surface uplift in the absence of upper crustal shortening in a number of orogens, notably beneath the eastern Tibetan plateau and its margins (Clark and Royden, 2000). Ouimet and Cook (2010) and Yang et al. (2003) have proposed that in the central Andes, ductile flow within the lower crust redistributes crustal material along strike, translating material from areas of greater crustal shortening toward areas with less shortening. Therefore, lower crustal flow has a potential role in controlling the morphology and development of high topography. The latter process can have a direct influence on plate motions, because gravitational spreading from high topography can contribute to the global plate tectonic force balance (Avouac and Burov, 1996; Iaffaldano and Bunge, 2009; Pascal and Cloetingh, 2009).

In previous studies (Kaban et al., 2010; Koulakov et al., 2009; Tesauro et al., 2008, 2009a,b) we developed a model of the European crust and upper mantle (Fig. 2), by assembling most of the existing geophysical data and estimates of physical (density, temperature, velocity) and rheological parameters (Fig. 3a–k). These models have been used to estimate the integrated strength and the effective elastic thickness of the European lithosphere (Tesauro et al., 2009a,b). Here we use these models to compute lateral pressure gradients (LPG) resulting from variations in crustal thickness and density (Fig. 3k). The LPG distribution is then employed to predict ductile flow strain rates, adopting end-member rheology models. Comparison of these estimates with the mantle anisotropy and GPS velocity vectors allows to examine the fine structure of deformation within the lithosphere. The results are particularly important for quantifying the thickness of low viscosity zones in the lowermost part of the crustal layers.

2. Lateral pressure gradient and crustal flow: basic considerations

Pressure variations associated with topography and Moho undulations may reach 10–75 MPa (e.g. 1000 m of local topography or 5000 m of Moho depression produce 20–30 MPa pressure difference). Ductile flow is produced in areas where tectonic or gravitationally induced stresses approach the ductile strength of the rock. Rheological studies show that for quartz-rich crust, such flow occurs at depths corresponding to 250–350 °C (15–20 km) and requires stresses as small as 10 MPa (Ranalli, 1995). The crustal composition is vertically stratified, with rocks becoming more basic with depth. With decrease of quartz content, ductile creep activation occurs at higher temperatures for compositions such as quartz-diorite or diabase. Consequently, ductile flow may occur at different sub-parallel levels separated by zones of stronger material. An analytical solution for the evolution of the topography due to flow in the crustal channel for the case of local isostatic equilibrium of mantle lithosphere is provided by previous studies, including Bird (1991), Kaufman and Royden (1994) and Lobkovsky and Kerchman (1992). For more complex rheologies

Fig. 2. European topography (www.arcgis.com) with labels of main features.

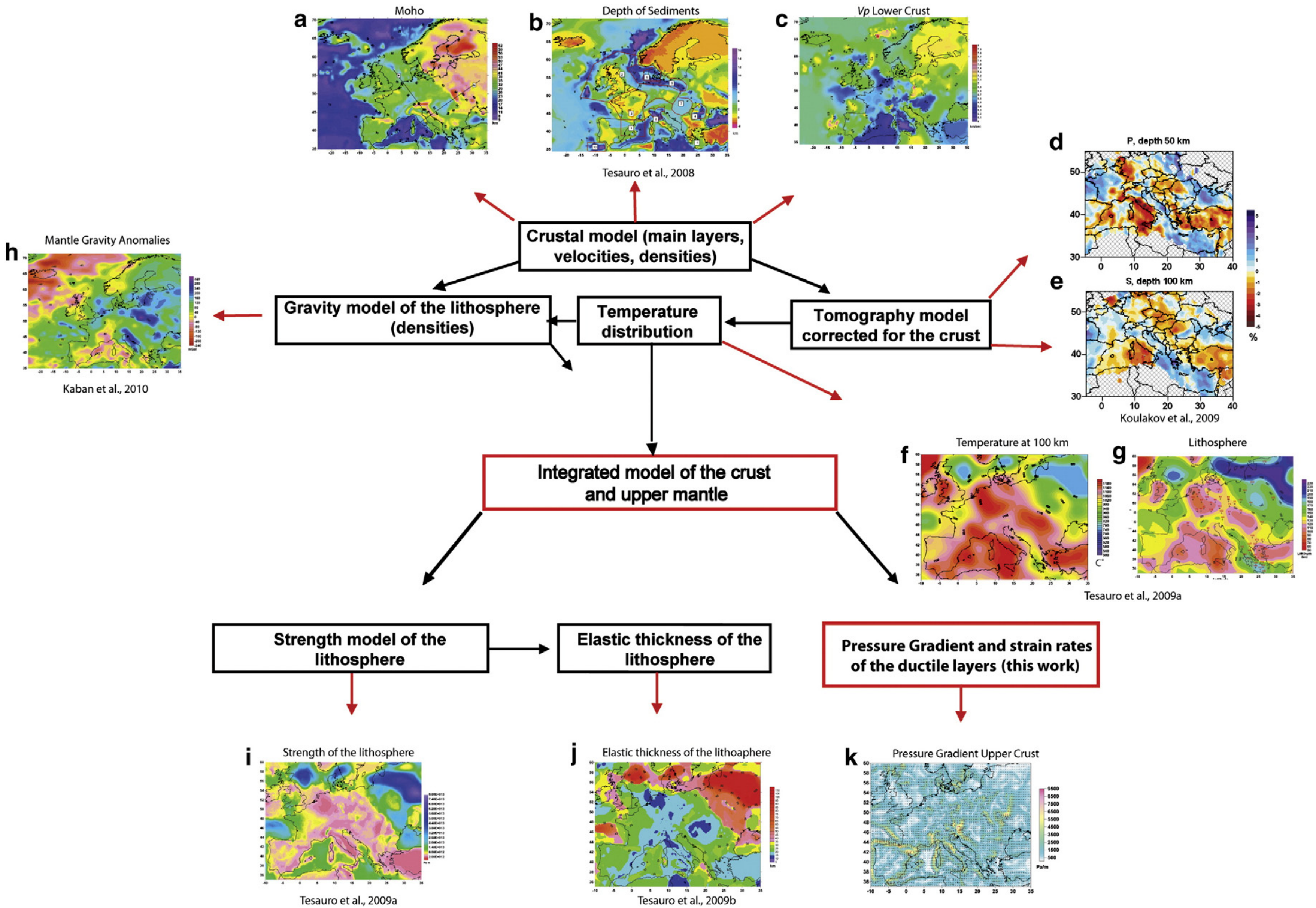

Fig. 3. (a–k). Workflow for modeling crust and upper mantle, involving determination of: (a) Moho depth (km); (b) Thickness of sediments (km); (c) Lower crustal velocity (km/s); (d–e) *P*- and *S-wave* mantle velocity anomalies (%); (f) Temperature at a depth of 100 km (°C); (g) Lithospheric thickness (km); (h) Mantle gravity anomalies (mGal); (i) Lithospheric strength (Pa m); (j) Effective elastic thickness of the lithosphere (EET)(km); (k) Lateral pressure gradient (LPG) (Pa/m).

involving inelastic bending of the lithosphere, surface processes and non-linear ductile flow, Avouac and Burov (1996) and Burov and Cloetingh (1997) have suggested a semi-analytical approach. In most of these studies, lubrication theory and/or close thin layer/sheet approximations are used to simplify the calculations (Bird, 1991; Medvedev and Podladchikov, 1999), and thickness of the thin ductile channel varies with time due to the evolving crustal flow and isostatic response of the lithosphere (see Supplementary material A1 for a full explanation of the method).

The goal of this study is not precise modeling of the crustal flow, but rather mapping of potentially affected zones. Hence we implement a widely used approximate approach that allows us to evaluate instantaneous flow patterns. This corresponds well to the goals of this study, which is based on present-day "snapshot" data. The upper surface topography and crustal structure are known a priori from independent data (Kaban et al., 2010 and Tesauro et al., 2008). Using these data we first estimate the LPG in orthogonal horizontal directions, x and z, at depths (y-axis is downward vertical) corresponding to the bottom of the ductile upper and lower crusts. In these areas the effective viscosity is expected to be minimal, resulting in maximal possible flow rates. The rheologies adopted for the strength models and LPG estimates are used to calculate gravitationally induced stresses and strain rates at the bottom of the ductile upper and lower crusts. At these depths we estimate the average shear stress caused by lateral pressure gradients. This stress is estimated under the assumption of incompressible continuous flow in a thin layer, by integrating LPG over the ductile thickness Δh_m of the given crustal layer (Avouac and Burov, 1996; Burov and Cloetingh, 1997; Lobkovsky and Kerchman, 1992), as explained in the supplementary material A1.

Our model shows ductile channels in the upper and lower crust with very sharp jumps in the rheology. These jumps mainly reflect small-scale artifacts of the parameterization rather than real features of the structure. As a result, the initially estimated LPG and ductile flow patterns are affected by strong noise, which masks the potential tectonic signal. To eliminate these artifacts, the horizontal variations of the ductile channel boundaries are smoothed by a low-pass filter leaving the wavelengths greater than 111 km (maximum resolution 0.5°). These results were used to estimate the LPG and ductile flow distribution.

3. Rheological and density structure of Europe

In this section the crustal density and thickness variations, which are the main parameters needed to estimate the LPG, are shown along specific cross-sections. In addition, strength envelopes for typical structures, estimated using the rheological and thermal model of Tesauro et al. (2009a) are displayed. The cross-section in Fig. 4a

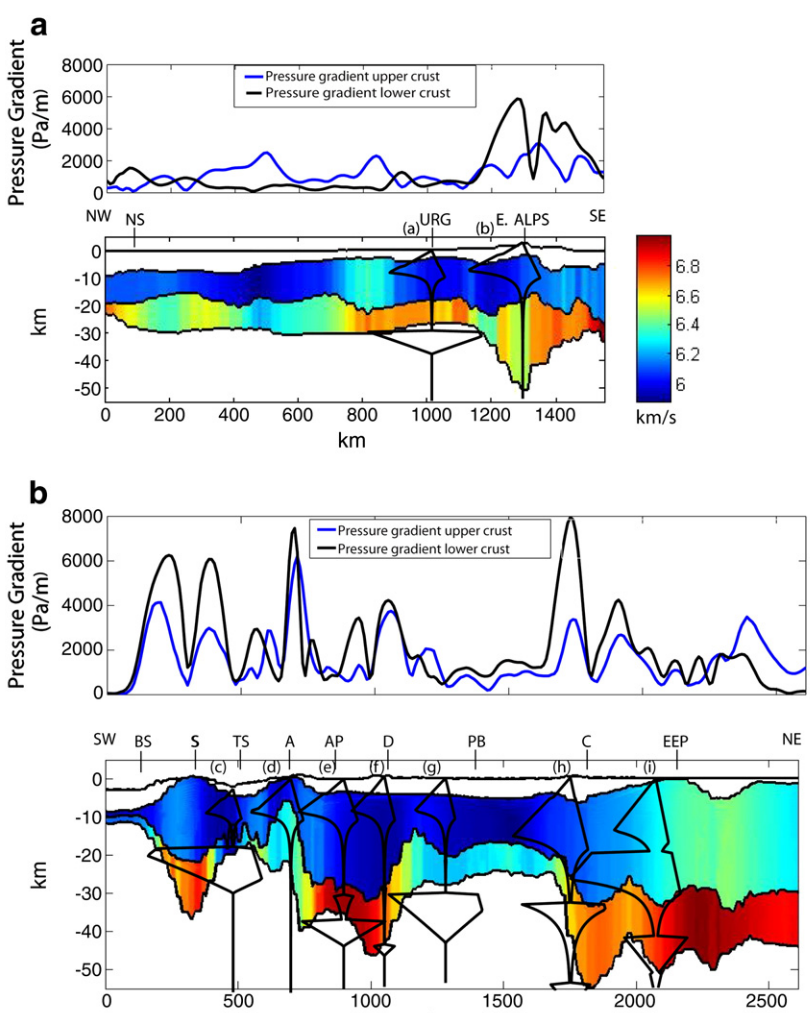

Fig. 4. (a–b). Amplitude of the absolute pressure gradient (top panels) at the bottom of the upper and lower crust (Pa/m) and density (g/cm^3) and thickness (km) distribution (bottom panel) along two cross-sections. Strength profiles are displayed for selected points. Cross-sections are depicted in Fig. 7a. Abbreviations are as follows: A, Apennines; AP, Adriatic Promontory; BS, Balearic Sea, C, Carpathians; D, Dinarides; E. Alps, Eastern Alps; EEP, East European Platform; NS, North Sea; PB, Pannonian Basin; S, Sardinia; TS, Tyrrhenian Sea; ECRIS, European Cenozoic Rift System. Strength profiles are displayed in selected points (see text for explanations).

illustrates general features of the crustal structure in Central Europe. Crustal thickness increases by about 10 km from the North Sea to the North German Plain. To the southeast the crust thickens up to ~50 km under the eastern Alps and thins again to 25–20 km under the northern Adriatic coast, while the lower crust density increases up to 2.90 g/cm^3. The cross-section in Fig. 4b shows the difference in crustal structure between the Eastern European platform (EEP) with high density and crustal thickness and the heterogeneous areas west to the Trans European Suture Zone (TESZ). The Tyrrhenian Sea, Apennines, Eastern Alps, Pannonian basin and European Cenozoic Rift System (ECRIS) are characterized by different crustal thickness (Fig. 4a and b) and thermal conditions, but by the same rheology (Tesauro et al., 2009a). The associated strength profiles (points a, b, c, d, g) strongly differ only below the Moho. The European Cenozoic Rift System (ECRIS), Pannonian and Tyrrhenian basins (points a, c and g), have relatively thin crust and thus a high fraction of the strength in their upper mantle. In contrast, the lithospheric mantle is ductile at the bottom of the thick crust of the Eastern Alps and Apennines (points b and d), as shown by the absence of a strong mantle zone. In all these areas the ductile lower crust causes crust–mantle decoupling. The Dinarides and the Adriatic plate have a crustal thickness and thermal regime similar to the Eastern Alps and the Apennines, but show brittle conditions in their lower crust (points e and f), on account of the stronger rheology used (Tesauro et al., 2009a). In particular, the brittle behavior of the lower crust of the Adriatic plate (point e) controls its coupling with the mantle lithosphere. In the Carpathians and the EEP, characterized by deep Moho and different thermal conditions, both the upper and lower crust are mostly in a brittle regime (points h and i), which results in a full coupling of the lithospheric layers.

4. Uncertainties in the model and their potential impact

As shown in the previous section, the crustal thickness and the thermal regime appear to influence mostly the strength partitioning between the crust and mantle, but not the strength distribution within the crust, nor the coupling/decoupling conditions, which in turn mainly depend on rheology. Therefore, we discuss below a range of integrated crustal strengths for alternative models with 'soft' and 'hard' rheology, in order to construct a set of possible end-member strength models.

4.1. End-member rheological models of the crust

The first crustal strength model ('soft rheology', Fig. 5a) implies 'dry quarzite' and 'wet diorite' rheology for upper and lower crust (Carter and Tsenn, 1987). The second one ('hard rheology', Fig. 5b) implies 'dry granite' and 'mafic granulite' rheology (Carter and Tsenn, 1987; Wilks, and Carter, 1990) for the same layers. The other assumptions and parameters (Table 1) are those used in Tesauro et al. (2009a). Even though there is a growing amount of more recent rheology data, we decided here to stick to the commonly referenced sources because it is the extrapolation of rock mechanics data to geological spatial and time scales that brings major uncertainties to rheological models of the lithosphere. The two end-member models actually show quite different strength distributions. In the first, crustal strength varies between mostly weak Western and Central Europe ($<1\times10^{13}$ Pa m) and strong areas to the north-east of the TESZ. In the second model, the strength shows similar minima and maxima, but varies over a larger range. The crust is weak in the Tyrrhenian and Balearic Sea, but strengths sharply increase along the continental

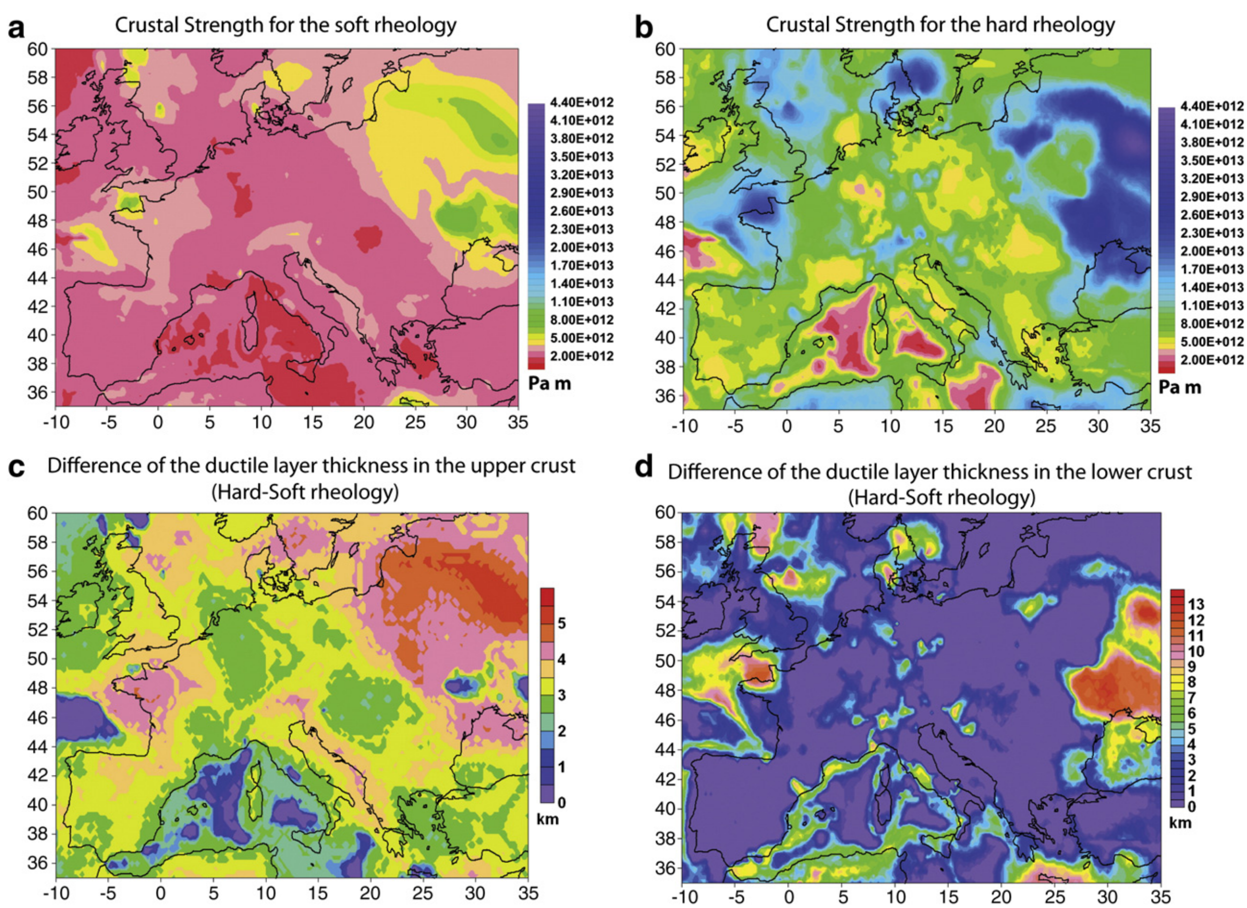

Fig. 5. (a–d). Crustal strength and ductile thickness variations for two rheologies. (a) Crustal strength for a soft rheology; (b) Crustal strength for a hard rheology; (c) Difference between the upper crustal ductile thickness estimated for the two rheologies; (d) Difference between the lower crustal ductile thickness estimated for the two rheologies.

Table 1
Rheological model parameters. Numbers in square brackets indicate source: [1] Carter and Tsenn (1987); [2] Wilks and Carter (1990); [3] Goetze and Evans (1979).

Parameter	Symbol	Units	Sediments	Upper crust	Lower-crust	Upper mantle
Composition	–	–	–	Quarzite (dry) [1]/ Granite (dry) [1]	Mafic ganulite[2]/ Diorite (wet) [1]/	Olivine (dry)[3]
Friction coefficient ext/com	f	–	0.75/3	0.75/3	0.75/3	0.75/3
Pore fluid factor	λ	–	0.36	0.36	0.36	0.36
Power law exponent	n	–	–	2.72/3.3	4.2/2.4	3
Power law activation energy	E_P	KJ mol^{-1}	–	134/186	445/212	510
Power law material constant	A_P	$Pa^{-n}s^{-1}$	–	$6.03x10^{-24}/3.1x10^{-36}$	$8.83x10^{-22}/1.26x10^{-16}$	$7.00x10^{-14}$
Dorn law activation energy	E_D	KJ mol^{-1}	–	–	–	535
Dorn law strain rate constant	A_D	s^{-1}	–	–	–	$5.70x10^{11}$
Dorn law stress	s_D	Pa	–	–	–	$8.50x10^{9}$
Strain rate	$\dot{\varepsilon}$	s^{-1}	–	10^{-15}	10^{-15}	10^{-15}
Brittle Strength	$\sigma = f\rho gz(1-\lambda)$					
Power law creep	$\sigma = \left[\frac{\dot{\varepsilon}}{A_P}\right]^{\frac{1}{n}} \cdot \exp\left[\frac{E_P}{nRT}\right]$					
Dorn law creep	$\sigma = \sigma_D\left(1-\left[-\frac{RT}{E_D}\cdot \ln\left(\frac{\dot{\varepsilon}}{A_D}\right)\right]^{1/2}\right)$					

margins. Pronounced strength variations occur at boundaries of the main geological features of western and central Europe (e.g. Pannonian Basin, Bohemian Massif, Massif Central). In contrast, the EEP's strength in the second model is four times higher. Thus rheological changes have more influence in areas with a large crustal thickness and a low thermal regime. Consequently, estimates of strength in these areas have largest uncertainties (Tesauro et al., 2010).

As in previous studies (e.g., Avouac and Burov, 1996; Burov and Diament, 1995), the thickness of the ductile crustal channel(s) (TDC) in the two models is computed as the depth interval between the BDT (brittle–ductile transition depth) in the corresponding lithologic layer and the bottom of the crustal layer (Fig. 5c and d). The variations of the TDC in the upper crustal layer are relatively modest (up to ~5 km). They tend to increase from the oceanic domains to the continental margins. In the continental areas, the smallest variations are predicted in areas characterized by a high thermal regime (Pannonian basin, ECRIS and Massif Central). In contrast, the largest variations are predicted in the easternmost EEP, where upper crustal thickness is large (Tesauro et al., 2008). For the lower crust, more than half of the study area shows no difference in the TDC estimated

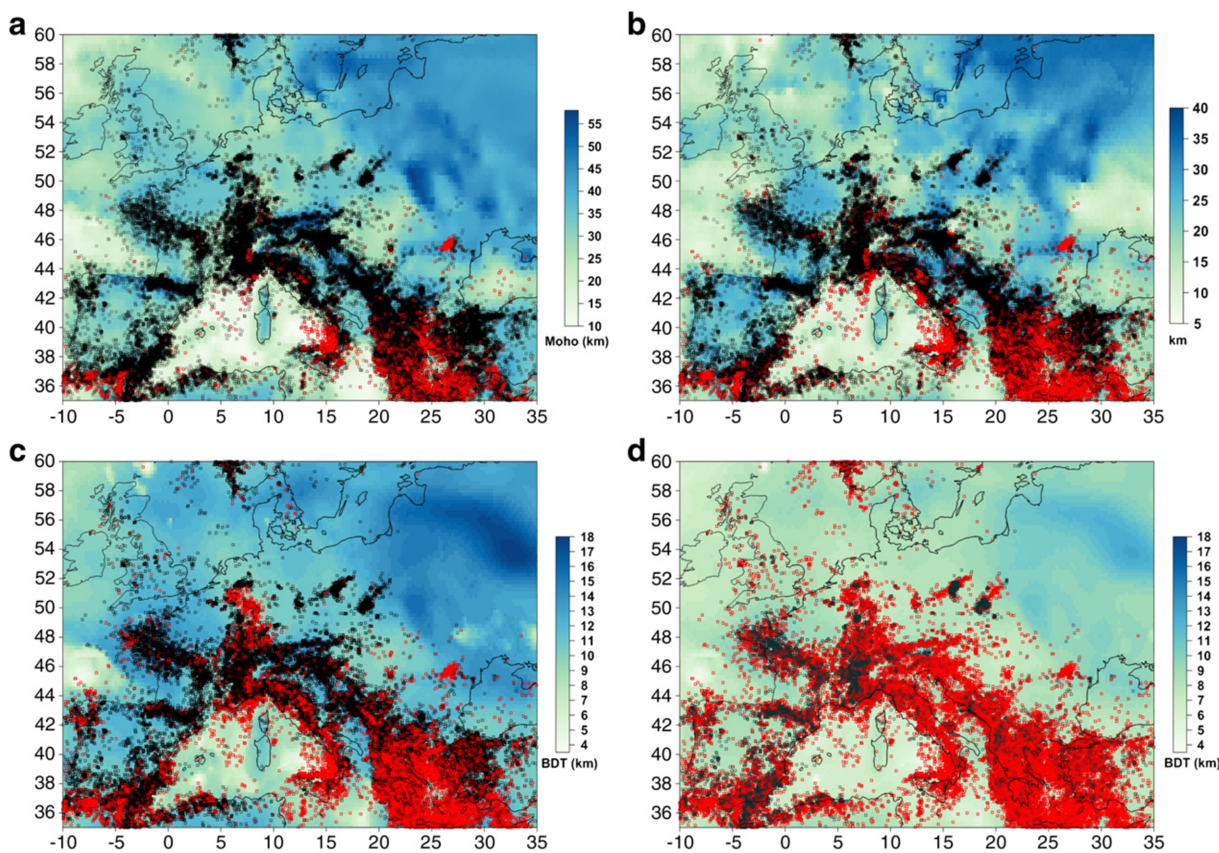

Fig. 6. (a–d). Seismicity distribution. The database contains more than 100,000 seismic events with magnitudes between 1 and 9 for the period 1973–2010. (a) Black and red circles correspond to earthquakes in the crust and mantle lithosphere, respectively. Moho depth is diplayed in the background; (b) Black and red circles correspond to earthquakes in the upper crust and below this layer, respectively. Moho and boundary between upper and lower crust are from EuCRUST-07 (Tesauro et al., 2008); (c) Black and red circles correspond to earthquakes in the brittle upper crust and below this layer, respectively. The brittle–ductile transition, displayed in the background is from the strength model of Fig. 5b; (d) Black and red circles correspond to earthquakes in the brittle upper crust and below this layer, respectively. The brittle–ductile transition, displayed in the background is from the strength model of Fig. 5a.

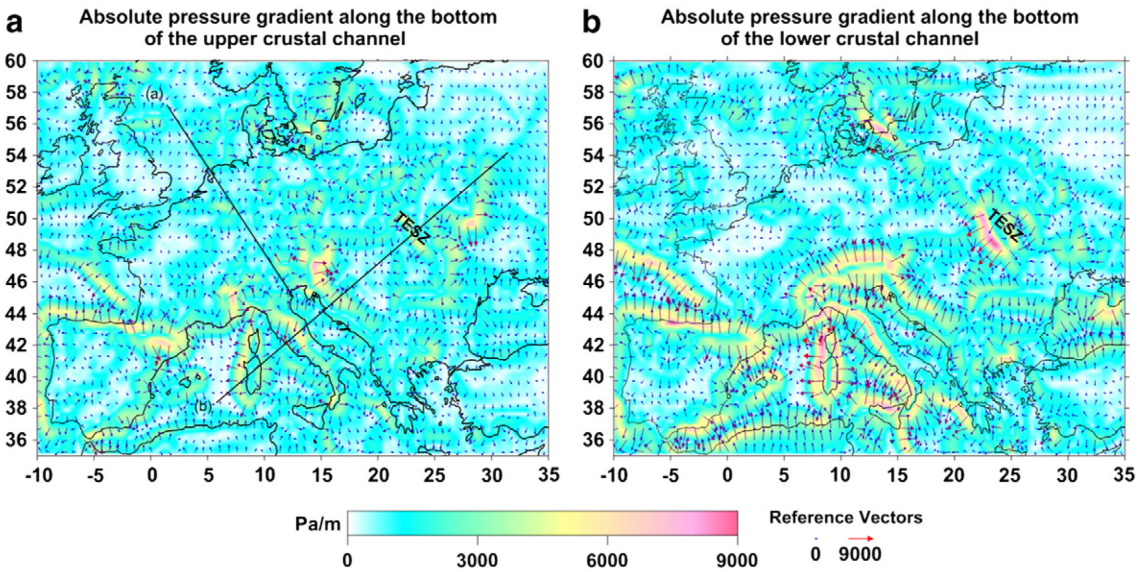

Fig. 7. (a–b). Absolute, or maximal lateral pressure gradient at the bottom of the crustal channels. The absolute lateral pressure gradient is computed along the bottom of the channel and is naturally higher than non-lithostatic pressure gradient. Yet, the hydrostatic pressure in the channel compensates pressure differences due variations in its depth, if density of the channel is the same as that of the embeddings. This assumption is not valid in case of high fluid content or temperature differences. The absolute pressure gradient provides upper bounds on pressure differences in the crustal channel. (a) Upper crust; (b) Lower crust. Black lines in Fig. 7a show the location of the cross-sections in Fig. 4(a–b). The amplitude of the absolute pressure gradient is displayed in the background.

from the two rheological models. Thus in the lower crust, other parameters (e.g. temperature) might have a stronger effect on the distribution of ductile strain with depth. Large differences between the TDC predicted by two models (up to 10 km) are mostly observed along the continental margins, where the TDC changes sharply. The largest variations (>10 km) are found in the southeastern part of the EEP, corresponding to the largest thickness of the lower crust (Tesauro et al., 2008).

4.2. Comparison of the end-member models and seismicity

It is generally accepted that frictional earthquakes generally occur within crustal and mantle layers that remain brittle on long time scale (e.g., Burov, 2011; Watts and Burov, 2003). The highest earthquake concentrations (Fig. 6a–d) occur in areas with low crustal strength, confirming the hypothesis (Burov and Watts, 2006; Watts and Burov, 2003) that concentrations of crustal seismicity reflect stress levels approaching brittle strength in areas of long-term rheological weakness. The Armorican Massif is the only exception from this correlation, because it is characterized both by relatively high seismicity and significant crustal strength. More than 90% of the earthquakes are located within the crust (Fig. 6a), while those below the Moho are mostly concentrated in subduction–collision zones. However, this subcrustal seismicity is probably related to non-frictional failure, e.g. metastable mechanisms that are not directly related to rock strength (e.g. Kirby et al., 1991). Hence, this seismicity is not discussed in the present study. Furthermore, most of the earthquakes (~85%) occur in the upper crust, while those in the lower crust mostly occur in subduction–collision areas, similar to the subcrustal earthquakes (Fig. 6b). Therefore, along plate boundary zones the stress induced by the subducting plate may generate faulting in the deeper part of the upper plate's crust. Fig. 6c and d shows earthquakes located within the brittle part of the upper crust, as defined by the strength envelopes for the two end-member rheologies. Although, as discussed above, the two rheologies employed in strength calculations produce similar estimates for the depth of the brittle–ductile transition of the upper crust (on average ± 2.5 km), the percentage of the earthquakes occurring in the brittle part of the upper crust significantly drops (from 72% to 32%), when a 'soft rheology' ('dry quarzite') instead of a 'hard rheology' ('dry granite') is used. The maximum depth of seismicity should generally not exceed the BDT depth, at which aseismic ductile creep processes start to dominate the deformation (e.g. Meissner and Mooney, 1998; Meissner et al., 2002). Therefore, the BDT depths in the two end-member rheological models might define the upper and lower bound of their possible range. On the other hand, most of the intraplate seismicity in the ECRIS and the Molasse basin occurs below the BDT defined by the two models (Fig. 6c). Deep crustal earthquakes in the Molasse Basin have been interpreted as indicating high fluid pressure (Deichmann, 1992). Deep seismicity in the ECRIS is probably related to its extensional tectonic regime (Cloetingh et al., 2005). Earthquake focal mechanisms in this region indicate that deformation of the upper crust is controlled by a strike slip to compressional stress regime while the lower crust is subjected to extension (Deichmann et al., 2000; Plenefisch and Bonjer, 1997). Extension of stabilized continental crustal segments precludes ductile flow of the lower crust and faults will propagate toward the basin center (Bertotti et al., 2000). Therefore, the lower crust will deform by distributing ductile shear in the BDT domain.

5. Lateral pressure gradients and strain rate distribution

The absolute LPG estimated at the bottom of the upper and lower crusts (Fig. 7a–b) show similar patterns but different magnitudes. These results suggest that the direction of the horizontal stress affecting both crustal layers remains constant with depth. Higher variations in thickness and density (Fig. 4(a–b)) occur at the bottom of the lower crust than at the bottom of the upper crust (Kaban et al., 2010; Tesauro et al., 2008). This increase promotes formation of ductile flow channels in the lowermost crustal layer. Tectonic processes can give rise to significant changes in the distribution of crustal material. The gravitational forces associated with the high topography of young orogens might also drive lateral crustal flow. To estimate pressure gradients potentially driving crustal flow, we computed absolute pressure gradient along the bottom of the crustal channels (Fig. 7a–b) and lower-bound non-lithostatic horizontal pressure gradient (Fig. 8a–b). Our models predict large values of the absolute and non-lithostatic LPG perpendicular to the axes of mountain belts, such as the Alps, Apennines, Pyrenees–Cantabrian Mountains, Dinarides–Hellenic arc and Carpathians (Figs. 7a–b and 8a–b). In general, crustal flow is directed away from orogens towards adjacent

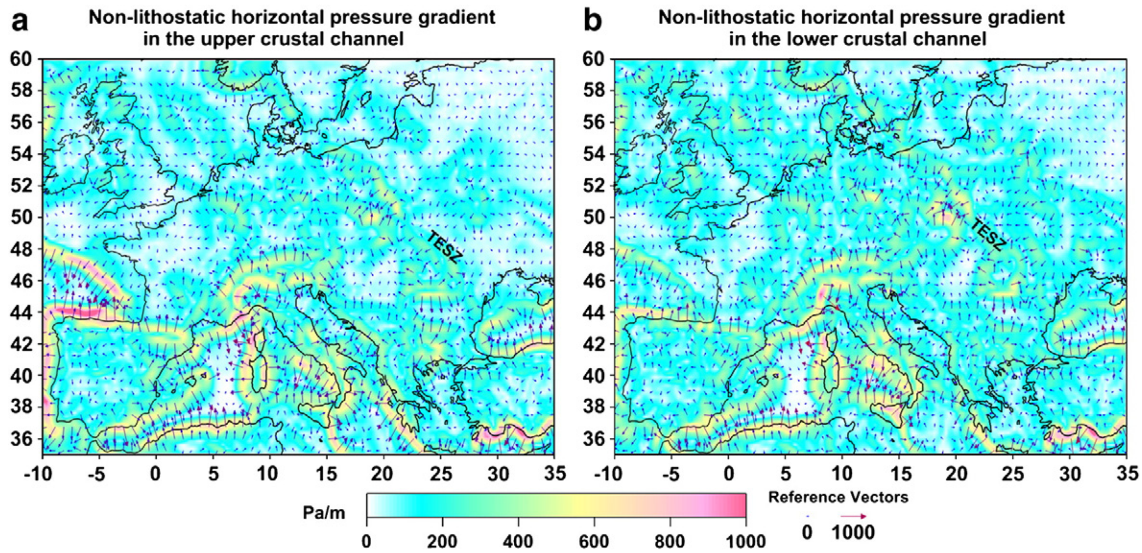

Fig. 8. (a–b). Non-lithostatic horizontal pressure gradient in the crustal channels. Pressure difference between two neighboring grid points is computed at same depth in global Cartesian coordinate framework assuming no over/under pressure and constant channel density within the interval between these points. The result yields lower bounds on horizontal pressure gradient. (a) Upper crust; (b) Lower crust. The amplitude of the non-lithostatic horizontal pressure gradient is displayed in the background.

areas (Figs. 7a–b and 8a–b). For many of these areas extension is confirmed by seismic and geodetic observations (e.g., Kastrup et al., 2004). The pronounced change in the LPG occurring between the orogens and the basins is also visible in the cross-sections of Fig. 4(a–b). Gravitational forces directed from areas of high gravitational potential energy to subsiding basin areas can strongly reduce extension in the basins, leading to a gradual late stage inversion of the entire system. This process is observed, for example, in the Pannonian Basin, which originated in an extensional regime, but is currently under compression (Bada et al., 2001).

Strain rate distributions estimated using Eq. (5) at the bottom of the upper and lower crust for the soft and hard rheologies are shown in Fig. 9(a–d). For a soft rheology, strain rates at the bottom of the upper and lower crusts largely vary and are higher than 10^{-13} s^{-1} (Fig. 9a and c) in some parts of Western Europe, which implies a high velocity flow in the ductile crust. Strain rates obtained assuming the strong rheology, show a different distribution (Fig. 9b and d). At the bottom of the upper crust high velocity flow (for strain rates $>10^{-13}$ s^{-1}) might be limited to some parts of the orogens, like the Pyrenees, Eastern Alps and Southern Carpathians. At the bottom of the lower crust the flow extends under entire orogens. In these areas orogenic stresses perpendicular to the axes of mountain ranges and lateral pressure gradients due to thick crust are sufficient to cause mountain-parallel flow.

The maximum predicted strain rates in most cases are high compared to background strain rates. This implies that the stability of the associated structures is of dynamic nature. As expected, mountain belts drive important gravity flow and their stability is conditioned by the presence of tectonic forces that partly or largely compensate gravity driven processes. Soft lower crust appears to promote intensive but relatively homogeneous flow under topographic highs, and strain rate contrasts on the order of 6–7 orders of magnitude. Surprisingly, the assumption of a stronger lower crust leads to a higher scatter and sharper gradients in the flow rates, even though the average flow rates are slower than for a soft upper crust. This difference can be explained by more non-linear behavior of the strong ductile rheology (higher "n" factor, Table 1). As mentioned above, the absolute values of maximum predicted strain rates are several orders of magnitude higher than average tectonic strain rates. Strain rates slower than 10^{-17} s^{-1} are negligible because they signify practical absence of gravity-driven components in the deformation. Even though those high values might be an artifact of, for example, overestimated temperatures at depth or inapplicability of dislocation creep laws in certain stress–strain rate intervals, it should be also noted that simple deformation models based on surface GPS data also often yield vertical strain rate values that are several orders of magnitude higher than typical tectonic strain rates, thus implying very low effective viscosities ($<10^{15}$ Pa s) at GPS time scales (e.g., Caporali et al., 2009 for Northern Anatolia). However, stability of geological structures requires mean lithosphere viscosities $>10^{22}$ Pa s (e.g. Gratton, 1989; Perazzo and Gratton, 2010). This might partly imply that surface deformation in areas of steep topographic gradients is largely controlled by weak ductile crustal deformation, and unrelated to the slower deformation in the mantle lithosphere. This finding is relevant in the context of the ongoing discussion on the possible partial decoupling of surface and mantle lithosphere deformation. Thus models based on vertical and sometimes horizontal GPS measurements may provide strain rates estimates different from geological rates in case of weak ductile crust (e.g. Liu et al., 2000). On the other hand, horizontal GPS rates are expected to be more compatible with geological strain rates because they reflect deformation of the strong brittle layers, at rates imposed by far-field tectonics. These findings evidently are pertinent only to continental lithospheres with weak intermediate or lower crustal layers. Hence, they do not affect classical plate tectonic models based on paleomagnetic reconstructions, which assume rigid plates (e.g., NUVEL — 1, DeMets et al., 1994).

Below, we compare these results to observations in a number of key areas in Europe. To this aim we examine in particular relationships between the LPG, the orientation of the GPS velocity vectors with respect to stable Eurasia, and mantle lithospheric anisotropy.

6. Zones potentially influenced by gravitational flow

6.1. The Alpine chain

The Alps began to develop in the Mesozoic, but their modern morphology is produced by the rapid Neogene counterclockwise rotation and N-NE directed indentation of the Adriatic microplate, which appears to be the dominant source of compression ("Adria-push"), e.g. Stampfli et al. (1998). This caused some movement of rigid and ductile material along the indentor, probably with a significant dip perpendicular component. GPS observations reveal that the Alpine-North Pannonian unit moves eastward at 2 mm/yr on average along two main faults as a consequence of the compression caused by the

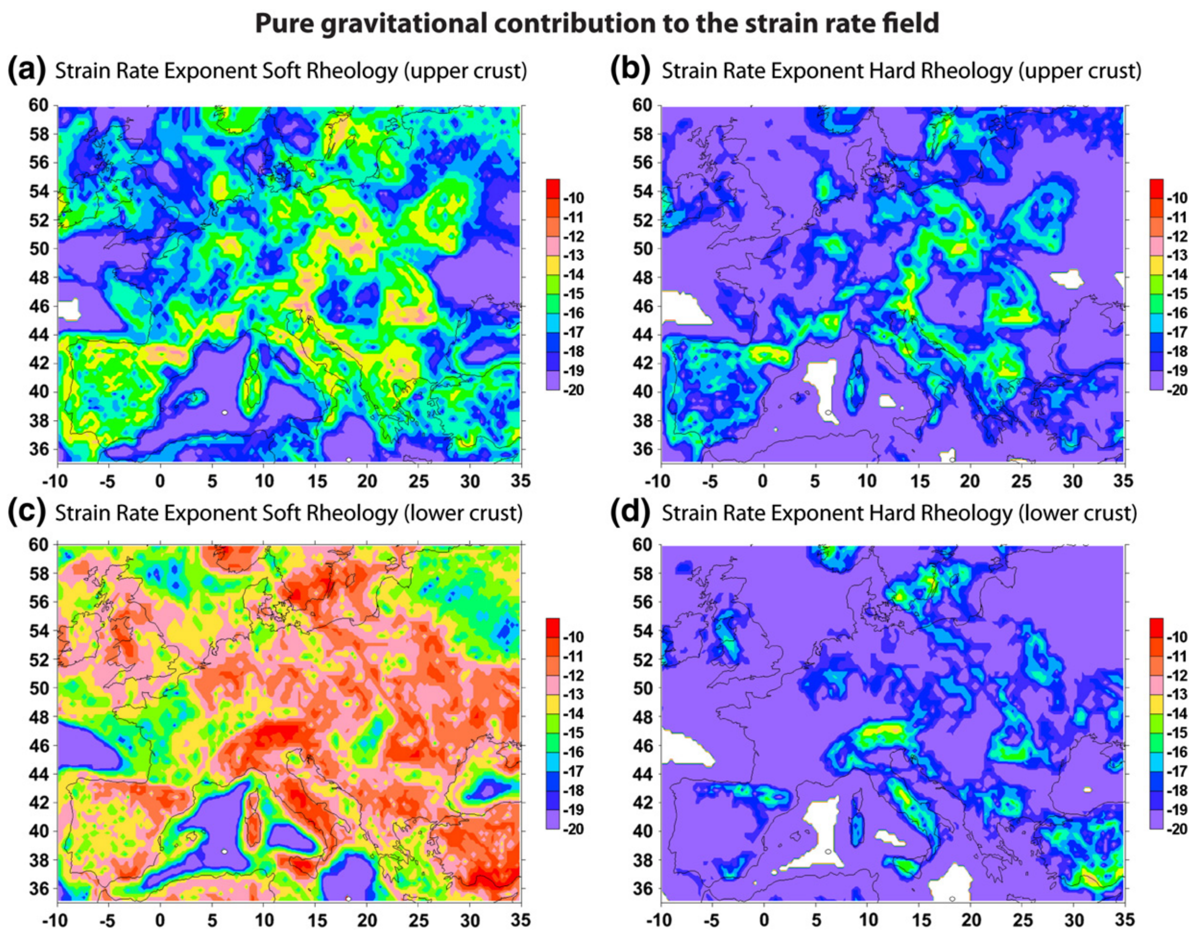

Fig. 9. (a–d). Strain rate exponent of the ductile crustal layers. (a) Strain rate exponent of the ductile upper crust predicted for a 'soft' rheology; (b) Strain rate exponent of the ductile upper crust predicted for a 'hard' rheology; (c) Strain rate exponent of the ductile lower crust predicted for a 'soft' rheology; (d) Strain rate exponent of the ductile lower crust predicted for a 'hard' rheology. White zones correspond to areas in which deformation in the crustal layers is restricted to the brittle regime.

Adria-Alpine collision (Grenerczy et al., 2000). A rotation of the compressional axes of the geodetic strain rates from NW-SE in the western part to NNE-SSW in the eastern part of the Alpine chain is observed (e.g. Caporali et al., 2009; Grenerczy et al., 2000; Tesauro et al., 2005, 2006). This rotation can be explained largely as a function of compressive forces acting on plate boundaries and of the geometry of the latter (Gruenthal and Stromeyer, 1992; Müller et al., 1992). The LPG at the bottom of ductile lower crust of the Alpine chain shows a similar rotation (Figs. 7b and 8b) and is compatible with previous local models (e.g., Jiménez-Munt et al., 2005). In addition, in both rheological models (Fig. 5a and b), the strain rates at the bottom of the lower crust are very high in the Eastern Alps ($>10^{-14}\,s^{-1}$), promoting crustal flow towards the Pannonian basin (Fig. 9(c–d)). Stresses are transferred far from Adria into the Pannonian basin, and the dominant style of deformation gradually changes from pure contraction through transpression to strike–slip faulting (Bada et al., 2007). Therefore, the Central and Eastern Alps are laterally escaping eastward toward the thin and weak lithosphere of the Pannonian basin, as observed by GPS data (e.g. Caporali et al., 2009; Grenerczy et al., 2000). Other evidence for tectonic escape, involving both the crust and the lithospheric mantle, is provided by the fast axes of olivine in the mantle lithosphere directed W–E and by faulting in the Eastern Alps (e.g. Ratschbacher et al., 1989).

6.2. The western Mediterranean and Pannonian back-arc basins

The back-arc extension of the western Mediterranean and Pannonian back-arc basins was initiated in Neogene times. Current deformation in the Pannonian basin is driven by the counterclockwise rotation of the Adriatic plate (e.g. Grenerczy et al., 2005). GPS data reveal 1.5 ± 0.5 mm/yr shortening across this basin, reflecting present-day inversion of extensional structures inherited from its Neogene formation (Grenerczy et al., 2005). The LPG in the Tyrrhenian and Balearic seas and Pannonian Basin, tends to decrease from the margins to the centres, predicting possible crustal flow in the same direction (Fig. 7a–b). However, only for the soft rheology model, are strain rates at the bottom of the lower crust (Fig. 9c) high enough ($>10^{-13}\,s^{-1}$) to support the crustal flow hypothesis. Other indications for possible crustal flow in these areas are given by the absence of a high-velocity (>6.8 km/s) lower crustal layer (Tesauro et al., 2008). This absence may be explained by delamination wherein decoupling occurs at the top of the lower crust. Non-uniform crustal extension over a large region gives rise to Moho undulations, which in turn might drive lateral lower crustal flow. The strong E–W anisotropy in the Tyrrhenian Sea (Mele, 1998) also suggests possible flow of the ductile lithospheric mantle.

6.3. The Apennines and the Dinarides

The Italian Apennines are of Tertiary age, and formed in response to the westward subduction of lithosphere beneath the Corsica–Sardinia continental margin (e.g. Doglioni, 1991). The GPS velocities and the observed earthquake slip vectors (e.g. D'Agostino et al., 2008) are NE directed, as a result of the Tyrrhenian extension and the convergence of the African plate. These data, together with roughly NW-SE trending normal faulting (e.g. Ascione et al., 2007), are consistent with tectonic escape of the ductile part parallel to the strike of the belt. At the same time, the strong uplift recognized in the Northern Apennines (e.g., Bartolini, 2003) might be interpreted as the effect of shortening and thickening of the deep ductile crustal layers induced by the belt-parallel compression (e.g. Mantovani et al., 2009). In particular, the combination of brittle extrusion in the upper crust and ductile thickening of the lower crust might explain why the inner Apennine belt has contemporaneously undergone horizontal extension and uplift (Mantovani et al., 2009). At the northeastern boundary of the Adria plate, the Dinarides respond to the Adriatic identation with a horizontal motion of 3–4 mm/yr in NE direction (Grenerczy et al., 2000). The increase of the LPG toward the margins of the Adriatic plate (Fig. 7a and b) supports the hypothesis that the convergence of the African and Adriatic plate involves detachment and transfer of crustal material between the latter and the two fold belts (e.g., Bennett et al., 2008). The strain rates results show that both the upper and lower crust of these orogens might flow when a soft rheology model is used (Fig. 9a and c). On the other hand, for the hard rheology, the flow might be limited to the lowermost part of the Apennine core (Fig. 9d). Evidence for possible flow of the ductile upper mantle, as a consequence of the same convergence process, is given by the fast axes of lithospheric mantle anisotropy, orientated parallel to the strike of the Apennines and the Dinarides (Mele, 1998; Smith and Ekström, 1999).

6.4. The Pyrenees–Cantabrian mountains

The formation of the Pyrenees–Cantabrian Mountains is connected to the Cretaceous counterclockwise rotation of Iberia, accompanied by opening of the Bay of Biscay, and with the final docking of Iberia to Europe. This N–S convergent movement led to the partial closure of the Bay of Biscay at the end of the Oligocene and the formation of the Pyrenees and the Cantabrian Mountains. The LPG in the Bay of Biscay appears to coincide with the transition of oceanic to continental lithosphere, inducing some additional stretching of this boundary. The N–S collision between the European and Iberian plates possibly results in the flow of the ductile mantle in E–W direction, as evidenced by the anisotropy orientation observed beneath the belt (Barruol and Souriau, 1995; Barruol et al., 1998). The inferred presence of high velocities at mid-crustal depths of the Cantabrian Mountains can be associated with portions of a lower crustal wedge from the northern (European) domain indenting the southern (Iberian) crust during the Alpine stage of compression (Pedreira et al., 2003). This indentation produces delamination of the Iberian crust, with northward underthrusting of its lower half and consequent crustal thickening. The latter process is also reflected by the northward orientation of the LPG (Figs. 7a–b and 8a–b). Both rheological models support the occurrence of upper and lower crustal flow beneath the Pyrenees. (Fig. 9a–d). In contrast, in the Cantabrian Mountains crustal flow might affect the crust, when it is characterized by a soft rheology (Fig. 9a–c).

6.5. The Carpathians

The collision between Europe and Moesia, the two major units composing the Carpathians, with a distinct mechanical contrast (i.e. strong vs. weak in terms of lithospheric scale rheology), resulted in deforming plate boundaries by gradual accretion of material from the lower (Moesian) plate (Matenco et al., 2010). The deformation thickens the lower plate along reverse faults inclined at a higher angle than the sole thrust. Therefore, the lower plate is not always a conveyer belt that transfers material to the upper plate during subduction and /or collision, such as in the case of the orogens with high convergence rates (e.g. the Alps, Schmid et al., 1996). Orogens such as the Carpathians produced by a 'foreland-coupling' collision have a characteristic first-order feature: crustal and lithospheric roots are not located beneath the core of the orogen, but shifted towards the foreland (Matenco et al., 2010). The NE orientation of the LPG in the outer Carpathians (Fig. 7b and 8b) supports this hypothesis, reflecting a possible indentation of the Carpathians crust towards the East European Platform (EEP). By contrast, the SW orientation of the LPG in the inner part of the Carpathians shows the tendency of the lowermost crustal layers to flow toward the Pannonian basin (Fig. 9a–d). If the crust is represented by a soft rheology, the high strain rates (>10^$^{-13}$ s^{-1}) at the bottom of the upper and lower crust predict a crustal flow of the entire Carpathian belt (Fig. 9a and c). In contrast, if the rheology of the crust is hard, the crustal flow might be limited to the inner part of the Eastern and Southern Carpathians (Fig. 9b and d). On the other hand, the crustal stress pattern in Romania is heterogeneous, prevalently controlled by local stress sources (Müller et al., 2010). Mantle anisotropy parallel to the strike of the orogen (Ivan et al., 2008; Russo and Mocanu, 2009), inferred from tomography studies (e.g. Koulakov et al., 2010) and numerical models (Ismail-Zadeh et al., 2005) provides evidence for the involvement of the entire lithosphere in the convergence process. Furthermore, a deeper fabric with E–W trend points to asthenospheric flow to the east, guided by the eastward thickening of the EEP cratonic lithosphere (Russo and Mocanu, 2009). The LPG southeast of the Carpathians is SSE oriented in agreement with GPS velocity vectors, which predict a horizontal movement in the same direction of ~2.5 mm/yr (Van der Hoeven et al., 2005).

6.6. The Anatolian Plateau and adjacent areas

The collision between Arabia and Asia started about 10 Ma ago (e.g. Burke and Sengör, 1986). The rapid convergence has produced crustal thickening and an elevation of 3 km in the Zagros and Caucasus Mountains. Consequently, the portion of Turkey between the northern and southern Anatolian faults is escaping westward, as documented by fault geometry, vertical cracks in the upper crust orientated in an approximate E–W direction (Mindevalli and Mitchell, 1989), focal mechanism solutions (Sengör et al., 1985) and GPS velocity vectors (Hollenstein et al., 2008; McClusky et al., 2000). We observe that the LPG is perpendicular to the northern Anatolian fault (NAF) (Figs. 7a–b and 8a–b), which started a dextral strike–slip movement since the end of the Miocene with a total displacement of ~50 km (Yilmaz et al., 1997). However, the chance of a crustal flow channel formation becomes high only at the bottom of the lower crust in the case of a soft rheogical model (Fig. 9c). Furthermore, anisotropy orientation along the NAF varies from NE-SW in the east to E–W and N–S in the central parts and to NW-SE in the west close to the Marmara Sea (Al-Lazki et al., 2004). The sudden change of the lithospheric mantle anisotropy direction contrasts with the relatively uniform GPS observed westward motion of the Anatolian plate (e.g. Hollenstein et al., 2008). This discrepancy possibly suggests that GPS data reflect present-day motions of Anatolia and that present-day surface deformation in this region may be relatively independent of deep crustal and mantle deformation (Brun and Sokoutis, 2010).

In the Crimean peninsula the LPG N–S oriented is consistent with stress indicator data and with the regional N–S Plio-Quaternary compression, which followed the NW-SE compression (Saintot et al., 2006). This process took place since the post-Middle Eocene inversion of the Black Sea, as a consequence of the closure of the Tethys Ocean to the south (Finetti et al., 1988; Nikishin et al., 2003; Robinson, 1997). A flow of the lowermost crust toward the Black Sea might be possible if the rheology is soft (Fig. 9c). The fast direction of anisotropy in the mantle of southern Crimea (Dricker et al., 1999) almost perpendicular to the LPG, supports a lithospheric ductile mantle flow in NW-SE direction.

7. Discussion

Spatial variations in crustal thickness and density in Europe's lithosphere induce patterns of lateral pressure gradients consistent with independent observations of surface motions and intraplate seismicity. Predicted non-lithostatic pressure and strain rate gradients suggest that gravity driven flow may play an important role in European intraplate tectonics (Figs. 8a–b and 9a–d). In particular, in a number of regions the predicted strain rates compete with tectonically induced strain rates. The predicted maxima of gravitational pressure gradients generally correlate well with intraplate seismicity patterns (Fig. 6a–d). The computed non-lithostatic lateral pressure gradients also predict strain rates at the bottom of crustal layers, while the absolute pressure gradients also allow to estimate potential directions of fluid migration (Figs. 7a–b and 8a–b). These results are also important to quantify the thickness of the low viscosity zones present in the lowermost part of crustal layers. Apart from being zones of potential crust-mantle decoupling resulting in differential deformation and drastic reduction of flexural resistance of the lithosphere, crustal low-viscosity zones are potential channels for lateral crustal flow, a process often referred to as crustal escape. As expected, ductile flow appears to be more vigorous in the lower crust than in the upper crust thus hinting that GPS data only partly reflect lithosphere scale dynamics. Comparison of these estimates with the orientation of the mantle lithospheric anisotropy and GPS velocity vectors allows us to detect and analyze relationships between crustal and mantle movements within the lithosphere. We expect that the predicted crustal flow correlates with the surface GPS data, while for lower level deformation it might be not the case. Our findings are consistent with the first order patterns for horizontal motions for the Alpine–Pannonian basin region where tectonic escape has been documented. This agreement points to a primarily gravitational nature of the eastward crustal extrusion from the Alps towards the Carpathian–Pannonian system, thus mitigating the potential role of the "Adria indenter" (Caporali et al., 2009; Horváth and Cloetingh, 1996). Furthermore, we estimated the possible range of integrated crustal strength values and strain rates at the bottom of the ductile upper and lower crust, using different sets of parameters for crustal rheology. Our estimates confirm the notion of gravitational collapse of the present-day Alps, Apennines and Pyrenees. The TESZ is also characterized by major gradients in crustal flow, suggesting that thrusting along this area is also partly of gravitational origin. Of particular interest is the potential gravitational nature of the Carpathian push to the southwest and the 3D character of compressional gravity forces acting on the Tyrrhenian and Balearic sea. In Iberia, intraplate compression and folding appear to be counteracted by lower crustal flow of gravitational nature (Fernández-Lozano et al., 2010).

8. Conclusions

The results of this study demonstrate the importance of gravity-driven lower crustal flow in the rheologically stratified lithosphere of Europe. The gravity-driven contribution to the ductile flow in the lithosphere may provide a strong background signal, superimposed on the regional flow patterns induced by plate tectonic forces and mantle flow. We have estimated possible contrasts in gravity-driven flow rates for the end member models characterized by 'hard' and 'soft' rheology. Such contrasts may be several orders of magnitude, leading to a strong dependence of the flow rate on surface and Moho topography. These findings suggest that the mantle lithosphere should be strong enough to preserve the observed Moho gradients over geological time scales.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at doi:10.1016/j.epsl.2011.09.055.

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