Extensional regime in the Northern Apennines: indications from the lithospheric behaviour

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ABSTRACT The aim of this work is to assess the conditions leading to extension during the tectonic evolution of the Northern Apennine chain, starting from the reconstruction of the rheological behaviour of the lithosphere underneath the chain. The temperature field and the strength of the lithosphere during the evolution of the belt in different sectors of the Northern Apennines, are calculated. A model, which assumes that the heat is transported by conduction and advection, is used to mimic a thermal perturbation migrating from west to east. The thermal regime is influenced by simultaneous crustal thickening in the easternmost sector and by advection due to extensional processes in the western sector of the chain. Rheological profiles and the strength of the lithosphere for different sectors of the chain at different times are obtained and compared with the difference in potential energy connected to the stress induced by compensated topography. The results show that, for homogeneous crustal thickening, the advection term moving form west to east would increase the tensile deviatoric stress in the area where the transition between the advection and thickening processes is located: in this area the strength is reduced in such a way as to trigger extension.

1. Introduction

The Northern Apennines (NA) are commonly interpreted as the result of the convergence between the already formed Alpine orogen and the continental crust of the Adriatic promontory of the African plate [e.g. Doglioni *et al*. (1998) and references therein] (Fig. 1). The convergence began during the late Oligocene – early Miocene, along with the rotation of the Corsica-Sardinia microcontinent, continued throughout the Tertiary age and is possibly still going on along the Romagna-Adriatic front (Frepoli and Amato, 1997, 2000), even if the present-day activity of the easternmost compressive front is debateable (e.g. Di Bucci and Mazzoli, 2002; Lavecchia *et al*., 2003). In the NA chain, the eastward movement generated progressively younger, foreland basins [foredeep and piggy back basins, Ori *et al*. (1986)], that were successively incorporated, from late Oligocene to the present-time, in the collision zone (Merla, 1951): the Adriatic Sea represents the present-day, active foredeep. At the same time, extensional deformation, related to the opening of the Northern Tyrrhenian Sea, propagated towards the western side of the Apenninic belt, starting in the Middle Miocene in the Corsica basin (Bartole, 1995). Extension continued in the Tuscan mainland during the late Miocene-Middle Pliocene, affected the western Umbria in the late Pliocene and is presently going on in the Umbria-Marche Apennine ridge, producing progressively younger, continental and shallow marine "hinterland" basins (Bartole, 1995; Jolivet

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et al., 1998; Pascucci *et al*., 1999). The coexistence and continuous eastward migration of both extensional and compressional structures has dominated the NA evolution since the early Miocene and is one of the distinctive characters of the NA orogen, as early recognized by Elter *et al*. (1975) (Fig. 1c).

As a consequence of this tectonic regime, an overall thinning of the lithosphere characterized the westernmost sector of the chain [with the lithosphere-asthenosphere boundary localized at about 30 km, Calcagnile and Panza (1981); Suhadolc and Panza (1989); Amato and Selvaggi (1991), Piana Agostinetti *et al*. (2002)] and it is reflected in the high surface heat flow [>150 mW/m^2, Mongelli and Zito (1991)] providing also an explanation for the positive Bouguer anomalies found in this domain (Marson *et al*., 1998) (Fig. 1a). The easternmost domain is characterized instead by a lithospheric thickness of about 70-90 km and the Moho is located at a depth of about 35-40 km (Ponziani *et al*., 1995; De Franco *et al*., 1998). The Bouguer anomalies are negative and there are relatively low values for the heat flow (between 70 and 40 mW/m^2) (Fig. 1a).

The transition between the two domains is located west of the Val Tiberina basin, where the zero of gravity anomaly occurs. In this region, geophysical soundings have shown a clear image of a regional east-dipping, low-angle, normal fault (Alto Tiberina fault) that crops out west of the Tiber Valley and dips towards NE, beneath the Apennine Chain, to a depth of about 12 km. This fault is presently active and represents the yougest and eastern expression of crustal extension in the NA (Barchi *et al*., 1999; Boncio *et al*., 2000; Pauselli *et al*., 2002), (Fig. 1b).

The recent revision of the available data collected during the last forty years underlines the role of the Neogene extensional tectonics on the recent evolution of the NA. The extensional deformation through the NA is realized by a system of normal faults dipping toward the east and the west-dipping normal faults are interpreted as their antithetic (Fig. 1). But, even if the entire orogen is dislocated by this east-dipping normal fault system, the geological and geophysical setting where the Tuscan faults are located is different from where the AltoTiberina fault rests. In fact, the extensional faults of the Tuscan region are located in a thinned crust, in a regime of anomalously high heat flow (>150 mW/m^2) and they are detached at the top of the lower crust, at ~12 km depth, while the underlying lower crust shows an anomalously reduced thickness (about 7 km), possibly due to a ductile flow (Fig. 1b).

The Alto Tiberina fault rests on a thickened crust, in a low heat flow setting with no evidence of post-kinematic granitoids. Many authors (Barchi *et al*., 1998a; Doglioni *et al*., 1998; Decandia *et al*., 1998; Jolivet *et al*., 1998) agree that the presence of the extensional east-dipping master faults along the entire Apennines has a common geodynamic origin. It is probable that the present difference in the heat flow conditions for the Tuscan faults and the Alto Tiberina fault reflects different periods of the activity: perhaps the eastward migration of the heat flow occurred after the migration of the tensional front responsible for the crustal thinning. In this scenario, the sector where the Alto Tiberina rests, represents a transitional zone between the extended western sector and the compressional eastern sector. Here, the presence of the Alto Tiberina fault indicates that extensional tectonics is now active, but the process of crustal thinning has not yet been completed.

The aim of this work is to investigate if stress induced by lithospheric thickening due to the eastward migration of the compressional front across the Apennines could induce a tensile stress regime and could trigger extension during the tectonic evolution of the NA.

For that aim the temperature field and the strength of the lithosphere during the evolution of the belt will be reconstructed in different sectors of the NA.

The strength of the lithosphere obtained will then be compared with the difference in potential energy stress connected to the stress induced by compensated topography (Mareschal, 1994), in order to assess the conditions leading to extension during the tectonic evolution of the NA.

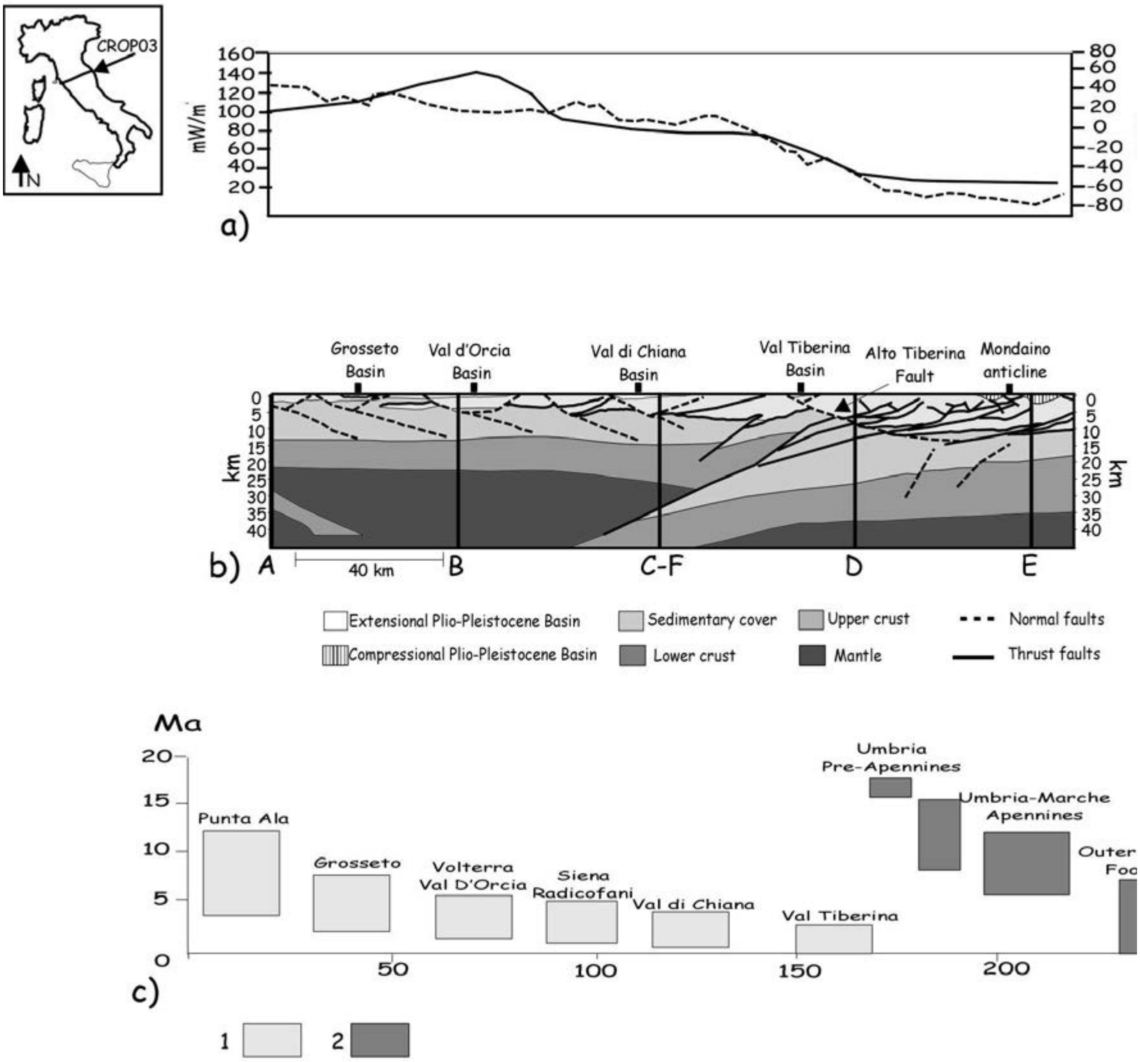

Fig. 1 - Location map of the study area. a) Surface heat flow (broken line) (Mongelli and Zito, 1991) and Bouguer anomaly (continuous line) along the CROP03 line (Marson *et al*. 1998). b) Geological interpretation of the seismic reflection line CROP03 [modified, after Barchi *et al*. (1998a)]. c) Migration of extension and compression along the CROP03 profile, as calibrated by the age of syntectonic 1) hinterland and 2) foreland basins [data from seismic reflection profiles Barchi *et al.* (1998b); Pascucci *et al*. (1999)]. A-B-C-D-E-F Reference points of rheological simulations.

2. Stresses induced by lithospheric thickening

Deformation induced by lithospheric thickening is testified by several pieces of evidence in present collision belts and has been largely investigated [for example: Love (1911); Bird and Piper (1980); England and McKenzie (1982); Buck (1991); Zhou and Sandiford (1992)]. Thin sheet models with vertical averaged stress and rheology have been studied and several papers have suggested that the topography of orogenic belts induces a tensile deviatoric stress that could trigger extension. The mode of extension was further investigated considering its dependence on the crustal thickness and thermal regime in order to compare the difference in potential energy induced by crustal and lithospheric thickening with the total strength of the lithosphere [see Mareschal (1994) for an exhaustive bibliography].

If the wavelength of the compensated topography is large as compared with lithospheric thickness, the stress induced in the lithosphere by the surface loading and density heterogeneities can be calculated. The deviatoric stress at depth is given by the difference between the vertical and horizontal stresses; the integration of the stresses over the entire lithospheric thickness yields the force per unit length, whose physical dimensions are equal to that of the energy density. This quantity is equivalent to the difference in potential energy per unit surface between the "thickened" and the reference lithosphere (Ramberg, 1981). In our modelling of the NA, "thickened" lithosphere refers to a lithosphere in which the thickening processes involve only the crust. The reference lithosfere (Table 1) is a initial lithosphere, in which the depth of the lower crust (d_{LCI}), the depth of the Moho (d_M), and the depth of the lithosphere (d_L) are considered constant along the profile (for the adopted values see Table 1). If the density distribution in the reference state is known, with the thickening involving only the crust, the difference in potential energy at the bottom of a constant thickeness lithosphere, is given by Mareschal (1994) in the following form:

$$\Delta E = \left(\gamma^2 - 1\right)\left[\int_0^{d_M} P(z)dz - \frac{P_M^2}{2g\rho_a}\right] + \left(\gamma - 1\right)P_M\left[(d_L - d_M) - \frac{P_L - P_M}{g\rho_a}\right] \qquad (1)$$

where γ is the crustal thickening factor (ratio of thickened crust to not a thickened crust), $P(z)$ is the lithostatic pressure at depth z in the non-thickened lithosphere, d_M and d_L denote initial (i.e., not thickned) depth of Moho (M) and initial (i.e., not thickned) base of the lithosphere (L), (for the adopted values see Table 1). P_M and P_L are the lithostatic pressure at Moho and at the base of the not thickened lithosphere, g is the acceleration of gravity, and ρ_a =3125 kg/m^3 is the assumed asthenosphere density.

Table 1 - Parameters for thermal simulation for MOD1 and MOD2. γ: crustal thickening factor; d_{LCI} = initial, not thickened depth of the lower crust; d_M = initial, not thickened depth of Moho (M); d_L = initial, not thickened base of lithosphere (L); v = pure shear constant strain-rate velocity; W = constant stretching zone.

Reference lithosphere	d_{LCI} (km)	d_M (km)	d_L (km)	v (mm/a)	γ	W (km)
MOD1	10	25	100	4	2	40
MOD2	10	25	100	1	2	40

In order to obtain the strength of the lithosphere, the temperature of the lithosphere during the NA evolution is needed.

3. Thermal evolution and mechanical behaviour of the lithosphere across the Northern Apennines

The strength of the lihosphere is the total force per unit width necessary to deform a lithospheric section at a given strain rate and it is a function of the composition, crustal thickness and geotherm. The concept of rheological profile, or strength envelope, was first developed by Goetze and Evans (1979) and on the basis of laboratory experiments, it can be assumed that the deformation regime for any given rock can be subdivided into two domains: brittle (B) and ductile (D). For a given strain rate, the B/D transition is defined by the equality of frictional strength of favorably-oriented pre-existing faults and the strength of the rock to steady-state ductile flow.

The brittle behaviour of rocks, for practically all rocks, could be described in terms of stress difference, overburden pressure, and pore fluid pressure, by the following expression:

$$(\sigma_1 - \sigma_3) = \beta g \rho z (1 - \lambda) \tag{2}$$

where σ_1 and σ_3 and are maximum and minimum compressive stresses, ρ is the density of the material, g the acceleration of gravity, z the depth, λ the pore fluid factor (ratio of pore fluid pressure to lithostatic pressure) and β is a numerical factor depending on friction coefficient and fault type [3, 1.2 and 0.75 for thrust, strike-slip, and normal faulting, respectively, if the friction coefficent $\mu = 0.75$; Ranalli and Murphy (1987)].

In practice, for all crustal and upper mantle rocks the creep regime is empirically described by a ductile flow law called power-law dislocation creep (Kirby and Kronenberg, 1987). This law relates the strain-rate to a n-power of the critical principal stress difference ($\sigma_1 - \sigma_3$), necessary to maintain a given steady-state strain-rate [see for example Ranalli (1995)]:

$$(\sigma_1 - \sigma_3) = \left(\frac{\dot{\varepsilon}}{A} \right)^{\frac{1}{n}} \exp\left(\frac{Q}{nRT} \right) \tag{3}$$

where $\dot{\varepsilon}$ is the strain rate, T is the temperature in degrees Kelvin, A and n are material creep parameters, Q is the activation entalpy for creep, R is the gas constant (8.314 J/mol K).

Laboratory experiments show that the activation entalpy for creep, Q, depends on several factors such as lithology, hydration degree and temperature, determining an increase of Q with depth (Kirby and Kronenberg, 1987). The material constants A and n also depend on the type of deformation and, slightly, on the temperature and suffer uncertainties due to their experimental determination (i.e., strain rate). In this work, we adopt values from Pasquale *et al*. (1993) for n, A and Q (Table 2). The strain rate has been chosen following Dragoni *et al*. (1996) and its value is 10^{-15} s^{-1} for sites between Tuscany and Val Tiberina and 10^{-16} s^{-1} for the Adriatic foreland, with a pore fluid factor, λ, changing with the depth and depending on the density of each layer (for the adopted values see Table 2). The strength of the lithosphere at any depth is thus the minimum between the stress for brittle deformation given by Eq. (2) and the stress required to maintain a given strain rate. Because the stress-strain relationship depends strongly on temperature, the stress required to trigger extension depends on the assumed temperature profile.

To obtain the temperature distributions across the NA, the general equation of heat diffusion where the advective transfer term is also present has been numerically solved over a 2D grid. Physical parameters, density, specific heat, thermal conductivity and volumetric radioactive heat production rate are assumed variable in space and their values are reported in Table 2a. It is to be noted, that the geological section, coming out from the interpretation of the Crop03 profile [Fig. 1b Barchi *et al*. (1998a)] has been simplified in only three different layers: the upper crust, the lower crust and the upper mantle (Table 2).

The boundary conditions are: the free Earth surface, at $z = 0$, is mainted at a constant temperature $T = 273$ K; the lower surface, at $z = D = 100$ km (base of the model) is set at a lateral varying value of the heat flow density (HFD) $Q_D(x)$ in order to simulate the thermal influence of the underlying mantle and of the anomalous thermal perturbation in the western side of the region. The assumed depth has been chosen in order to identify the 1600 K isotherm which thermally defines the lithosphere-asthenosphere boundary along the profile and during the time evolution. The lateral dimension of the 2D model is set equal 250 km; at the right-hand and left-hand sides of the model, the boundary condition is that the horizontal heat flow is zero. It is to be noted that the horizontal dimension of models is greater of the lenght of the seismic profile Crop03 to avoid the influence of the boundary conditions.

An initial temperature distribution is nedeed in order to start the computations and it has been obtained using results of our previous simulations (Federico and Pauselli, 1998) at the time $t = -12$ Ma (Punta Ala basin age, Fig. 1).

Our attempt is now to introduce the variety of dynamic processes characterizing the geodynamical scenario in the modelling: these different processes are reflected in the highly variable surface heat flux and, consequently, in the deep thermal regime. In particular, the peculiar characteristic of the NA, i.e. the coexistence and continuous eastward migration of both extensional and compressional structures since early Miocene, has been introduced considering that:

- the compressional front induces a crustal thickening, eastward migrating;
- the extensional front produces a thinning of the lithosphere through the effect of an advective term, eastward migrating.

For the sake of simplicity the velocity of these two migrations has been assumed equal and the needed value has been estimated interpreting the time-space evolution of the syntectonic basins from the data of Barchi *et al*. (1998b) and Pascucci *et al*. (1999).

In our modelling, both the thermal regime and the “mechanical” stratification of models are influenced by simultaneous crustal thickening in the easternmost sector and by advection due to extensional processes in the western sector of the chain, both eastward migrating. As regards the thermal regime, an advective term has been introduced. The advective velocity field has been simulated following the approach of Jarvis and Peltier (1980) and Jarvis (1983) for finite stretching rates, within a zone with a lateral extension of 2W, where a pure shear constant strain-rate velocity field is assumed.

Table 2 - a) Physical parameters for thermal simulation (Pasquale, 1987; Cermak and Bodri, 1995); b) Creep parameters after Pasquale *et al*., 1993). z = depth in km; d_{LC} = depth of the lower crust; d_{Moho} = depth of the Moho. Note that during the simulation the value of d_{LC} and d_{Moho} change following the thickening of the lithosphere; D = base of the model; ρ= density; c_p = specific heat; K = thermal conductivity; H = volumetric radioactive heat production rate; A and n are material creep parameters, Q is the activation entalpy for creep.

a)

MOD1-MOD2	ρ (kg/m^3)	c_p (J/kg K)	K(W/mK)	H(μW/m^3)	λ
Upper crust ($z<d_{LC}$)	2500	1000	2.5	$0.04qs\,e^{\left(-\frac{z}{10}\right)}$	0.40
Lower crust ($d_{LC}<z<d_{\mathrm{Moho}}$)	2700	1100	2.5	$2.5\,e^{\left(-\frac{z}{10}\right)}$	0.37
Mantle ($d_{\mathrm{Moho}}<z<$D)	3200	2900	3.5	0.01	0.31

b)

MOD1-MOD2	A(MPa^{-n} s^{-1})	n	Q(kJ/mol)
Upper crust ($z<d_{LC}$)	1.3×10^{-3}	2.4	219
Lower crust ($d_{LC}<z<d_{\mathrm{Moho}}$)	3.2×10^{-3}	3.3	268
Mantle ($d_{\mathrm{Moho}}<z<$D)	2.0×10^{3}	4.0	471

As regards the crustal thickening, it is to be notes, that the depth of the lower crust (d_{LC}) and the depth of the Moho (d_{Moho}), in Table 2, change during the simulationin time and in space along the geological section, starting from the Reference Model (Table 1) up to the actual lithosphere stratification coming out from the interpretation of the Crop03 profile (Barchi *et al*., 1998a).

We analysed two different models in which the maximum velocity is put equal to 4 mm/a (Mod1) and equal to 1 mm/a (Mod2) on the two external sides of the stretching zone whose initial half-length W is equal to 40 km [see Federico and Pauselli (1998) for details].

4. Results and interpretations

4.1. Thermal models

In Fig. 2b to 2d the temperature distributions for Mod1 at different time steps are reported. At present the temperature field (Fig. 2d) provides a calculated surface HFD in agreement with the measured one (Fig. 2a) (Mongelli and Zito, 1991) and the maximum surface HFD along the Crop03 seismic profile is located near the Val D'Orcia basin. However, it is to be noted that the positive anomal with pronounced horizontal gradient is not reproduced unless an unrealistic horizontal variation of heat flow coming from below is assumed. For example, the highest value of surface heat flow along the seismic profile east of the Val D'Orcia basin (Fig. 1) is located in correspondence with the projection on the seismic profile of the Mt. Amiata geothermal field, and this anomaly is probably due to the superposition of a local anomaly, connected with the presence of a magmatic body (Mongelli and Zito, 1991), and a more intense regional temperature field due to the astenospheric rise beneath Tuscany. Thus, the regional crustal model would have to be refined, taking into account the distribution of zones of hydrological disturbance or considering possible conductivity contrast. Such additional modelling is beyond the scope of the present work. During the temporal evolution, the thermal anomaly laterally extends and increases its temperature (Fig. 2b to 2d). At present (Fig. 2d) the thermal anomaly is located at about 130 km from the beginning of the profile. In the easternmost side of the profile, it is possible to note the changes in the temperature field induced by the bottom boundary condition (HFD variations) and by the thickening of the crustal layer at the expense of the mantle one.

In Fig. 3a the temperature defining the lithosphere-asthenosphere boundary (1600 K) is also reported. The intersection of the different geotherms with the 1600 K isotherm identifies the values of the thickness of the "thermal" lithosphere (Fig. 3b): going from west to east the thickness of the thermal lithosphere increases from about 30 km to 90 km in Mod1. For comparison, see also Fig. 3b where the thickness of the "seismic" lithosphere, obtained by Panza *et al*. (2003) using the data of the surface wave velocity tomography and applying a non-linear inversion method, is reported. The thermal lithosphere calculated here is thinner than the seismic one, even if the general trend is similar. This means that the obtained temperature field is higher than the one indirectly constrained by the results of Panza *et al*. (2003).

Simulations performed using different values for the thickness of the layers, semilength *W* and rate of deformation do not give results differing more than 10% compared to Mod1. A decrease in the velocity (1 mm/a, Mod2) determines a general decrease of temperature field as shown in Fig. 4a, where the present day geotherm in the same Reference Points of Fig. 2 (Mod2) are reported. Inspection of Fig. 4b shows that the thickness of the Mod2 thermal lithosphere in the western sector is in better agreement with the seismic lithosphere, but the surface HFD values largely differ (more than 50%) from the measured ones (Fig. 4a).

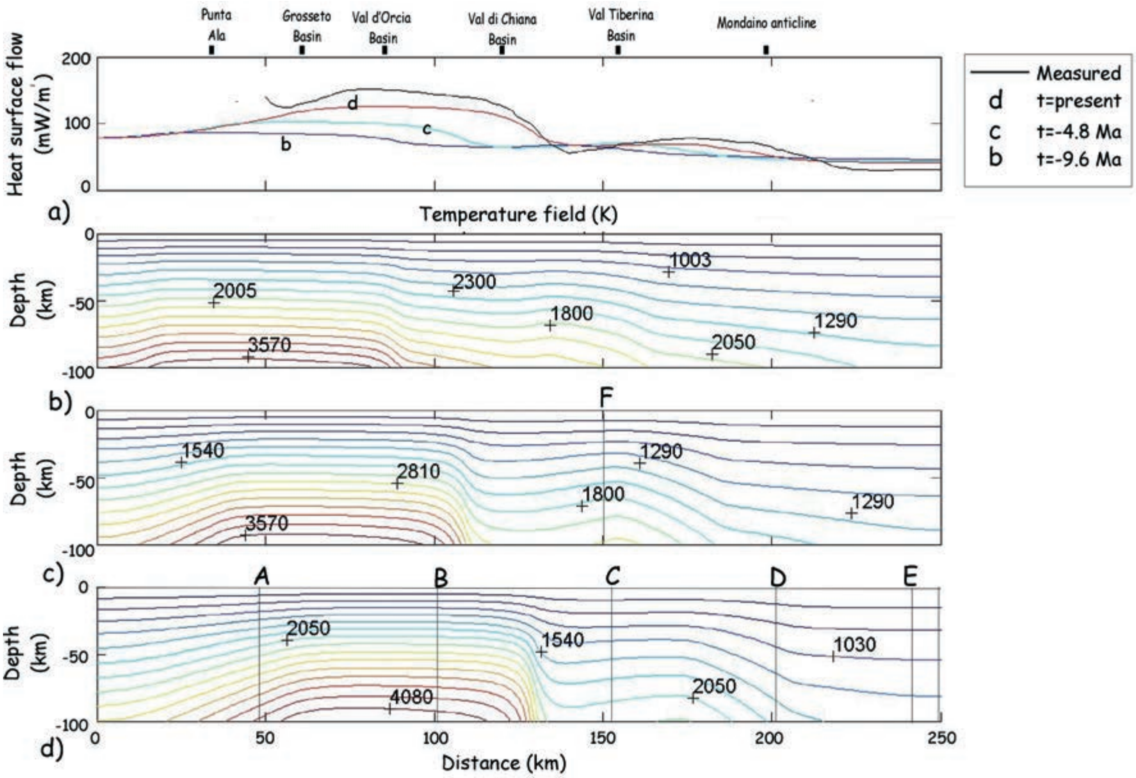

Fig. 2 - a) Comparison between measured and calculated surface heat flow for different time steps (Mod1) b): temperature field (Kelvin) for t = -9.6 Ma. c) temperature field (Kelvin) for t = -4.8 Ma d) temperature field (Kelvin) for t = present. A-B-C-D-E-F Reference Points. Crosses and number in figure are the values of temperature in those points.

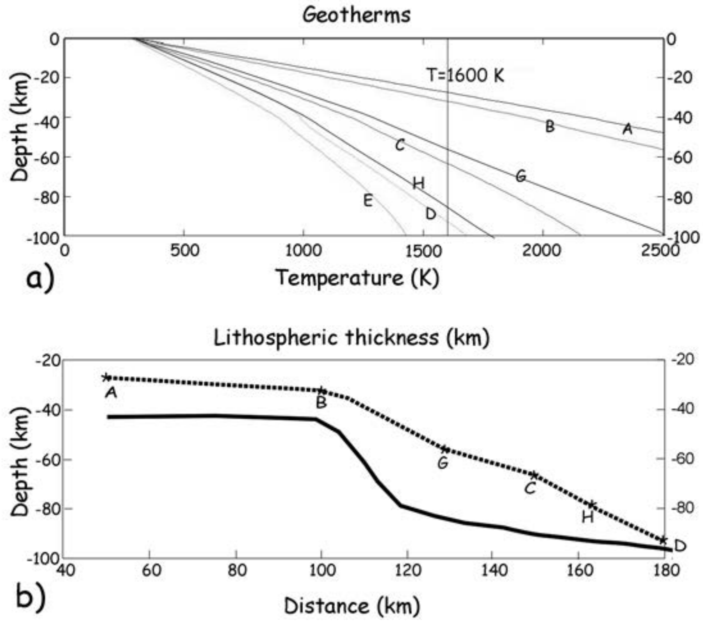

Fig. 3 - a) Geotherms for A-B-C-D-E reference points, and G and H additional points used to validate the model (see text for explanation), for Mod1. The 1600 K isotherm is also reported. b) calculated thermal lithospheric thickness (broken line) and seismic lithospheric thickness (continuous line) after Panza *et al.* (2003).

In the intermediate region (from Reference Point B to D) where the transition between compressional and extensional front is present, the two thermal models do not substantially differ and as far as the aim of this paper consists in verifying the influence of the thermal regime on the tectonic, we prefer Mod1 that better agrees with the surface HFD. It is to be noted that the discrepancies between depth of the lithosphere calculated here and the depth of the lithosphere obtained by Panza *et al*. (2003) arises from the different meaning of lithosphere: the lithosphere indentified by our modelling refers to a "thermal lithosphere" defined by the isotherm 1600 K, whilst the seismic one is set by the seismic velocity of the Low Velocity Zone discontinuity. For that reason, the discrepancies are tolerable, and they are also due to uncertainties of the thermal parameter values and surface heat flow mesaurements. The intersection of other geotherms with the 1600 K isotherm (Point G and H in Fig. 3), in addition to the Reference Points, confirms the similarity of the general trend between the "thermal" lithosphere and the "seismic" one, validating Mod1.

4.2. Lithospheric strength (S) vs. potential energy difference (DE)

Using the Mod1 thermal field, stress envelopes for the different geotherms are calculated and reported in Fig. 5, assuming strain rates of $10^{-15}s^{-1}$ for the Reference Points B,C and D (sites between Tuscany and Val Tiberina) and $10^{-16}s^{-1}$ for Refence Point E (Adriatic Foreland).

The B/D obtained transition is in good agreement with the depth distribution of the earthquake ipocentres recorded and deepens eastwards. A satisfactory agreement is found with the frequence

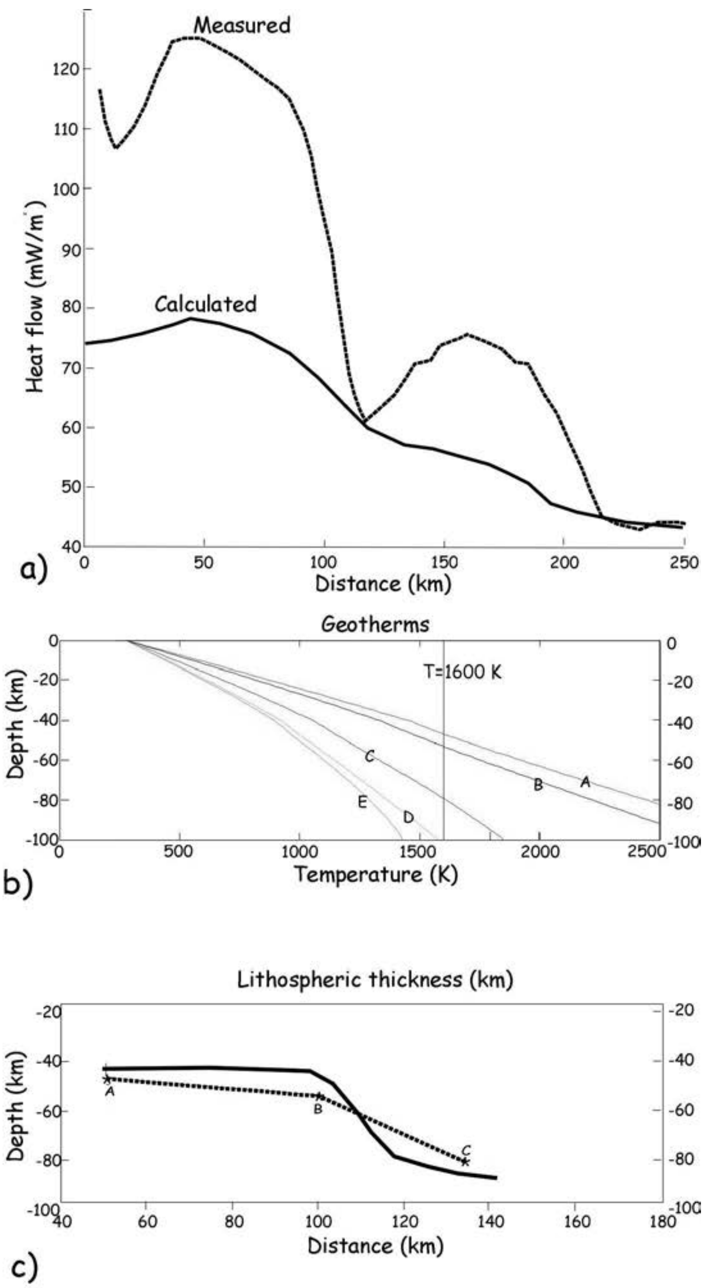

Fig. 4 - a) Calculated and measured surface heat flow. b) Geotherms for A-B-C-D-E Reference Points for Mod2. The 1600 K isotherm is also reported. c) calculated thermal lithospheric thickness (broken line) and seismic lithospheric thickness (continuous line) after Panza *et al*. (2003).

of the occurence of events with magnitude 3 in the NA: in an area with surface heat flow of 80 mW/m^2 (Reference Point B), about 90% of the ipocentres of earthquakes are localized within a depth of 10 km; in the area with a surface heat flow of between 60 and 80 mW/m^2 (Reference Point C), the greater concentration of the ipocentres of earthquakes is within a depth of 15 km; in the area with a surface heat flow of between 40 and 60 mW/m^2 (Reference Point D), about the 90% of the ipocentres of earthquakes are localized within a depth of 10 km with 4% of events localized between depth of 10 km and 30 km; in the area with a surface heat flow of 40 mW/m^2

(Reference Point E), the greater concentration of the hypocentres of earthquakes is within depth of 25 km and few events in a depth greater than 30 km, with a gap of seismicity between a 25 and 30 km depth. A "sandwich stratification" of crustal layers is present in correspondence with the Alto Tiberina fault.

The integrated strength of the lithosphere, *S*, has been calculated for each stress envelope. The potential energy difference induced by lithospheric thickening, ΔE, has been calculated by using Eq. (1). Following Mareschal (1994), when *S* is less than ΔE, the conditions are such as to trigger an extension.

For the B, C and D rheological profiles, the strength of the crust is less than ΔE ($S<\Delta E$). Going towards east (profile E), the strength of the lithosphere is higher than the potential energy difference ($S>\Delta E$): in this condition the extension would never follow a crustal thickening. The temperature distribution is critical in determining the strength of the lithosphere, as is well explained by Mareschal (1994). In the case of the NA, the potential energy difference in the area of the Alto Tiberina fault (profile D in Fig. 5) is greater than the strength in most of the crust and

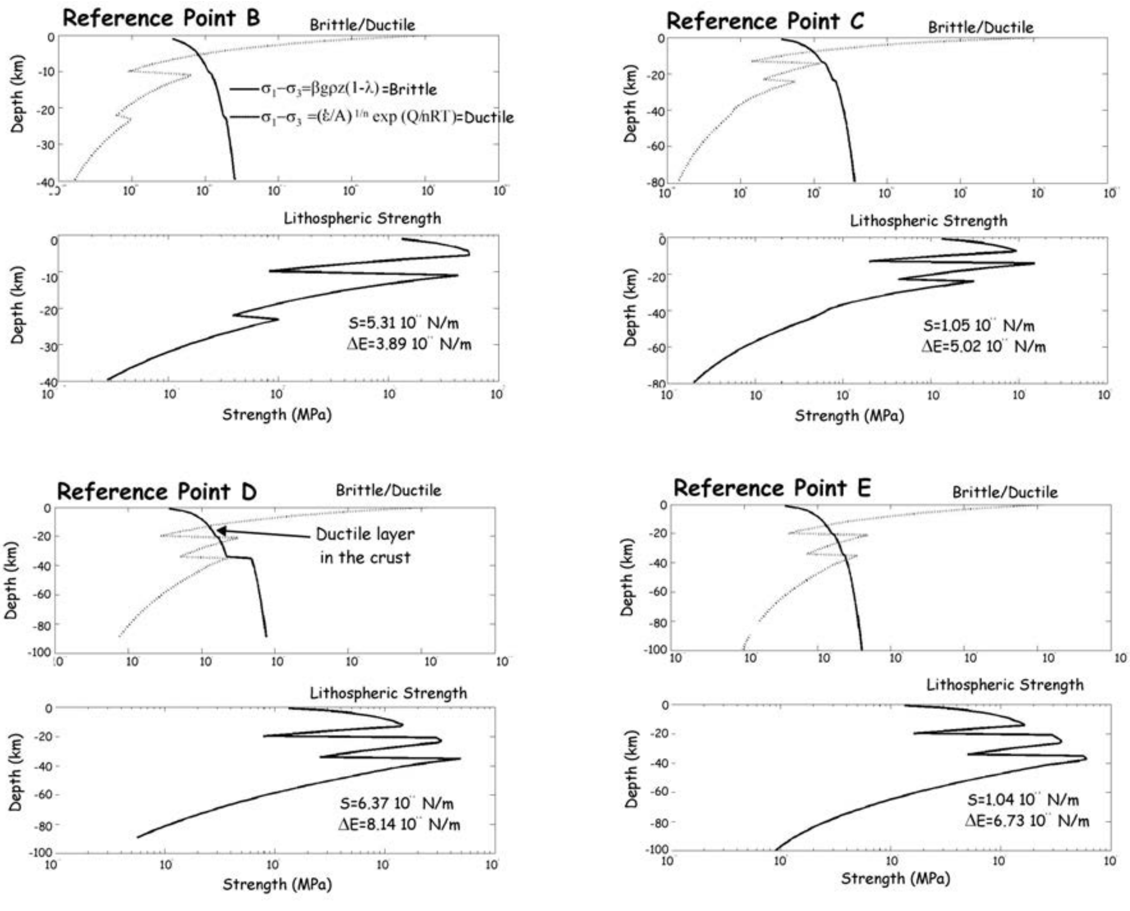

Fig. 5 - Stress envelopes and integrated lithospheric strength at present time for Mod1 for reference points B-C-D (area of Alto Tiberina fault)-E.

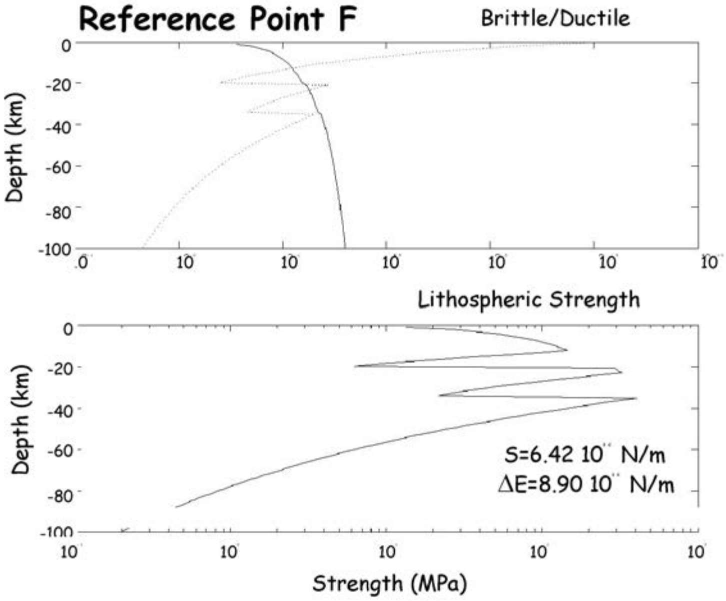

Fig. 6 - Calculated stress envelopes and integrated strength at time = -4.8 Ma in the area of Val Di Chiana basin for Mod1 (Reference Point F, see location, surface heat flow and temperature field in Fig. 2).

it could produce extension ($S<\Delta E$). This result is also valid for Mod2. In fact, as we have already noted, the geotherms C, D and E are similar for both the models.

In Fig. 6 the temperature field, the rheological profiles and lithospheric strength at time -4.8 Ma, when the extensional tectonics was localized in the Val di Chiana region (Fig. 1), are reported. The values of the strength are higher than the present day ones (profile F in Fig. 6) determining a rheological profile characterized by a sandwich structure within the crust. An inspection of the rheological profiles and of the values of the strength shows that at the present time a similar structure characterizes the crust beneath the Val Tiberina (profile D in Fig. 5), suggesting that the thermal evolution of the NA progressively determines a weakening of the lithosphere towards east, favouring the eastward migration of the extensional process.

5. Conclusions

The specific knowledge of the thermal structure and thermo-mechanical history is important for the understanding of the dynamics of geological systems. The performed procedure here allows us to assess the conditions leading to extension during the tectonic evolution of the NA. It is interesting to note that in our model the formation mechanisms of the belt have not been taken into account, because only the known geophysical characteristics of the lithosphere have been used. In this way, it has been possible to quantitatively determine the rheological conditions and where and when the extension could follow a compressional event, even in a belt characterized by a complex geodynamical history like the Apennines.

In the modelling, the peculiar characteristic of the NA, i.e. the coexistence and continuous eastward migration of both extensional and compressional structures since early Miocene, has been introduced considering that the compressional front induces a crustal thickening, eastward migrating, whilst the extensional front produces a thinning of the lithosphere through the effect of an advective term, eastward migrating.

The calculations show that the thermal effects of the continuous eastward migration of the crustal thickening term and of the advective term during the evolution of the NA, determines a strength reduction in the area where the transition of the two terms occurs. At the present time in the NA, the area of Alto Tiberina fault represents the last zone where the potential energy difference, being greater than the strength in most of the crust, is able to produce extension. An inspection of the rheological profile and of the value of the strength shows at time -4.8 Ma, when the extensional tectonics were localized in the Val di Chiana region, a similar structure of the present crust beneath the Val Tiberina suggesting that the thermal evolution of the NA progressively determines a weakening of the lithosphere towards east, favouring the eastward-migrating extensional process. Similar results that look only at the dynamics of the process, but use a more general physical formulation have been recently obtained by Ricard and Husson (2005).

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