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Rheological profiles in the Central-Eastern Mediterranean

Marcello Viti, Dario Albarello and Enzo Mantovani
Dipartimento di Scienze della Terra, Università di Siena, Italy

Abstract
Seismological investigations have provided an estimate of the gross structural features of the crust/upper mantle system in the Mediterranean area. However, this information is only representative of the short-term mechanical behaviour of rocks and cannot help us to understand slow deformations and related tectonic processes on the geological time scale. In this work strength envelopes for several major structural provinces of the Mediterranean area have been tentatively derived from seismological stratification and heat flow data, on the assumption of constant and uniform strain rate (10^{-16} s^{-1}), wet rocks and conductive geotherm. It is also shown how the uncertainties in the reconstruction of thermal profiles can influence the main rheological properties of the lithosphere, as thickness and total strength. The thickest (50-70 km) and strongest mechanical lithospheres correspond to the coldest zones (with heat flow lower than or equal to 50 mW m^{-2}), *i.e.*, the Ionian and Levantine mesozoic basins, the Adriatic and Eurasian foreland zones and NW Greece. Heat flows larger than 65 mW m^{-2}, generally observed in extensional zones (Tyrrhenian, Sicily Channel, Northern Aegean, Macedonia and Western Turkey), are mostly related to mechanical lithospheres thinner than 20 km. The characteristics of strength envelopes, and in particular the presence of soft layers in the crust, suggest a reasonable interpretation of some large-scale features which characterize the tectonic evolution of the Central-Eastern Mediterranean.

Key words *rheology – lithosphere – Mediterranean – geodynamics*

1. Introduction

The reconstruction of the recent geodynamic evolution of the Mediterranean area would greatly benefit from an estimate of the long-term rheological behaviour of the crust-upper mantle system in the major structural provinces. This information could give important insights into the interaction mechanisms of lithospheric blocks, both in the shallow and deep structures, which led to the present tectonic setting. In particular, the recognition of brittle and ductile layers inside the crust and upper mantle could help to evaluate the roles that some hypothesized tectonic mechanisms, such as slab pull, crustal delamination, etc. might have played in the Mediterranean evolution.

This kind of information can be obtained by experimental and theoretical studies. As widely recognized (see, *e.g.*, Ranalli, 1995), the mechanical behaviour of a given lithospheric body is mainly controlled by the nature of the stress regime (tensional or compressional), the strain rate, the mineralogical composition of the lithospheric rocks, the temperature profile, and the pore fluid pressure.

One-dimensional rheological profiles (strength envelopes), where the lesser of the brittle and ductile strengths is plotted, are usually taken as a first order representation of the

Mailing address: Dr. Marcello Viti, Dipartimento di Scienze della Terra, Settore di Geofisica, Università di Siena, Via Banchi di Sotto 55, 53100 Siena, Italy; e-mail: mantovani@unisi.it

mechanical behaviour of the lithosphere (see, *e.g.*, Lobkovsky and Kerchmann, 1991). In particular, this information can be used to constrain analogic and numerical modelling of finite-strength lithospheric plates (see, *e.g.*, Ratschbacher *et al.*, 1991; Sornette *et al.*, 1993; Spadini *et al.*, 1995; Bassi, 1995), to study lithospheric flexure both in continental and oceanic environments (Goetze and Evans, 1979; McNutt and Menard, 1982; Burov and Diament, 1995; Lowry and Smith, 1995; Mueller and Phillips, 1995; Cloething and Burov, 1996; Mueller *et al.*, 1996a) and to correlate seismicity with the local state of lithospheric stress and thermo-mechanical structure (see, *e.g.*, Meissner and Strelhau, 1982; Chen and Molnar, 1983; Smith and Bruhn, 1984; Govers *et al.*, 1992; Mueller *et al.*, 1996b).

In this work we tentatively computed the rheological profiles of some major structural provinces in the Central-Eastern Mediterranean region, by using suitable theoretical models and the available information on heat flow values and seismological layering of the crust-upper mantle system.

2. Methodological outline

Since the long-term mechanical behaviour of the lithosphere is here considered, attention has to be focused on anelastic deformations. Two major processes are considered: brittle failure and ductile flow. Both kinds of deformation are connected with complex microscopic mechanisms but, from the macroscopic point of view, some simplified descriptive approaches are currently adopted.

In a brittle layer, permanent deformations (related to sliding along pre-existing fracture zones) occur when the difference between maximum and minimum principal stresses overcomes a threshold value $\Delta\sigma_{br}$, which is mainly conditioned by the tectonic regime and orientation of faults, the lithostatic load (σ_v) and the pore fluid pressure (p) (Kohlstedt *et al.*, 1995). These two last parameters are related with the effective stress σ_e by:

$$\sigma_e = \sigma_v - p. \tag{2.1}$$

If fluid pressure is hydrostatic and the variations of gravity can be considered negligible in the range of depths under study, the effective stress at a given depth (h) can be expressed by:

$$\sigma_e\,(h) = g \int_0^h (\rho\,(z) - \rho_w)\,dz$$

where ρ and ρ_w are the densities of rocks and fluids respectively. The assumption of hydrostatic pore pressure can be considered a useful first order approximation despite the fact that deviations from hydrostatic equilibrium have been hypothesized for several tectonic environments (see, *e.g.*, Wang *et al.*, 1995) and that these deviations may play a major role in the determination of local lithospheric strength. The value of the threshold $\Delta\sigma_{br}$ can be considered, at least in a first approximation, as independent from temperature, rock composition and strain rate (Kohlsted *et al.*, 1995). It is expressed by the equations:

$$\begin{aligned} \Delta\sigma_{br} &= 3.68\ \sigma_e & (\sigma_e \le 113\ \text{MPa}) \\ \Delta\sigma_{br} &= 2.12\ \sigma_e + 176.6 & (\sigma_e > 113\ \text{MPa}) \end{aligned} \tag{2.2}$$

in the compressional regime and by:

$$\begin{aligned} \Delta\sigma_{br} &= 0.786\ \sigma_e & (\sigma_e \le 530\ \text{MPa}) \\ \Delta\sigma_{br} &= 0.679\ \sigma_e + 56.7 & (\sigma_e > 530\ \text{MPa}) \end{aligned} \tag{2.3}$$

in the tensional regime (Mueller and Phillips, 1995). These relations can be considered acceptable approximations of more realistic non linear relationships (Byerlee, 1978; Sibson, 1994; Lockner and Byerlee, 1995).

The threshold $\Delta\sigma_{br}$ can be considered a lower limit of the strength of a rock in the brittle regime, since it represents the maximum value of stress difference which can be actually sustained by lithospheric materials before frictional sliding.

In ductile layers, flow induces stress differences ($\Delta\sigma_{du}$) which depend on strain rate, temperature, and rock parameters. For crustal material in a steady-state flow, $\Delta\sigma_{du}$ can be obtained by the constitutive equation (Kirby, 1983):

$$\Delta\sigma_{du} = \left(\frac{\dot{\varepsilon}}{A}\right)^{\frac{1}{n}} e^{\frac{H}{nRT}} \tag{2.4}$$

where $\dot{\varepsilon}$ is the strain rate, R is the gas constant, T is the absolute temperature, A is an experimental parameter, H is the activation enthalpy and n depends on the experimental conditions and mineralogical composition of rocks (see, *e.g.*, Kirby and Kronenberg, 1987). When $\Delta\sigma_{du} > 200$ MPa and upper mantle rocks (assumed as mainly composed by olivine) are concerned, the constitutive eq. (2.4) becomes:

$$\Delta\sigma_{du} = \sigma_0 \left\{ 1 - \left[\left(\frac{RT}{H'}\right) \ln \left(\frac{\dot{\varepsilon}_0}{\dot{\varepsilon}}\right) \right]^{\frac{1}{2}} \right\} \tag{2.5}$$

where σ_0, and $\dot{\varepsilon}_0$ are experimental constants (8500 MPa and 5.7 10^{11} s^{-1} respectively) and H' is assumed to be equal to H in the eq. (2.4).

If rocks are assumed to yield either by frictional sliding or ductile flow, the maximum stress difference that rocks can sustain (strength) is given by:

$$\Delta\sigma_{max} = \operatorname{sign}[\dot{\varepsilon}] \min[\Delta\sigma_{br}, \Delta\sigma_{du}] \tag{2.6}$$

where $\Delta\sigma_{du}$ and $\Delta\sigma_{br}$ can be determined by eqs. (2.1)-(2.5) ($\dot{\varepsilon} > 0$ for the compressional regime). The depth distribution of this minimum is the strength envelope. It represents the state of stress of the lithosphere under large strain and a constant strain rate out of transient regimes.

Despite the fact that a number of assumptions and approximations are involved in the assessment of strain envelopes (see Ranalli, 1997, for a discussion), these curves can be considered a useful tool to clarify the dynamics of many long-term tectonic processes. From the strength envelope it is possible to deduce synthetic parameters such as the lithospheric thickness (L) and the total lithospheric strength (R), which characterize the long term mechanical behaviour of the lithosphere as a whole (Fadaie and Ranalli, 1990).

The lithospheric thickness can be defined in several ways. Fadaie and Ranalli (1990) proposed that the lower boundary of the lithosphere corresponds to the maximum depth where

$$|\Delta\sigma_{max}| > 1 \text{ MPa} \tag{2.7}$$

while McNutt (1984) proposed

$$|\Delta\sigma_{max}| > 100 \text{ MPa}. \tag{2.8}$$

The «total lithospheric strength» (dimensionally a force per unit length) is given by:

$$R = \int_0^L \Delta\sigma_{max}(z)\,dz \tag{2.9}$$

where L is the depth given by eqs. (2.7) or (2.8) (Fadaie and Ranalli, 1990).

Equations (2.4) and (2.5) can be interpreted as representative of a non-newtonian viscous flow (Ranalli, 1995). The effective viscosity η, representing the slope of the $\Delta\sigma_{du} = f(\dot{\varepsilon})$ curve, can be obtained by (2.4) and (2.5) respectively:

$$\eta(h) = \dot{\varepsilon}^{(1-n)/n} e^{\frac{H}{nRT(h)}} (2nA^{1/n})^{-1} \tag{2.10}$$

$$\eta(h) = \sigma_0 (RT(h)/H)^{1/2} \{4\dot{\varepsilon}[\ln(\dot{\varepsilon}_0/\dot{\varepsilon})]^{1/2}\}^{-1}. \tag{2.11}$$

Assuming that the strain rate $\dot{\varepsilon}$ does not vary over time and depth, a viscosity depth profile can be computed as a function of lithology and temperature only.

By using (2.10)-(2.11) and assuming a constant strain rate of 10^{-16} s^{-1}, the two threshold values given by (2.8) and (2.7) can be converted in the corresponding viscosity values 10^{23} and 10^{21} Pa s respectively (Ranalli, 1995). Anderson (1995) suggested that these values respectively mark the lower boundaries of the «mechanical» lithosphere (*i.e.*, the part of the lithosphere supporting most tectonic load) and of the «thermal» lithosphere (*i.e.*, the boundary between the conductive lithosphere and the convective asthenospheric mantle). These boundaries are generally identified by the terms «Mechanical Boundary Layer (MBL)» and «Thermal Boundary Layer (TBL)» especially in the petrological literature (see, *e.g.*, Anderson, 1995).

Table I. Parameters adopted to describe ductile flow in the lithospheric rocks (see equations (2.4) and (2.5)). H is expressed in kJ mol^{-1} and A is in MPa^{-n} s^{-1} (Fadaie and Ranalli, 1990; Bodri and Iizuka, 1993). The values of the activation enthalpy (H) have been equated to the corresponding values of activation energy, thus neglecting the pressure effect which seems of minor importance for the lithospheric rocks (Ranalli, 1995).

	Rock	n	H	A
Upper crust (sediments)	Quartzite	1.90	173	0.0320
Upper crust (basement)	Quartz-diorite	2.40	219	0.0013
Lower crust	Diabase	3.30	268	0.0032
Upper mantle	Dunite	4.00	471	2000.

3. Model parametrization

As discussed above, the mechanical behaviour of the lithosphere mostly depends on strain rate, fluid pore pressure, lithological composition, temperature and density profiles.

Average strain rates in the Mediterranean area, roughly estimated on the basis of the present-day kinematic models in the Mediterranean area (Argus *et al.*, 1989; Albarello *et al.*, 1995) are of the order of 10^{-16} s^{-1}. This value can be considered representative of stable areas lying far from active tectonic belts (see, *e.g.*, Bodri and Iizuka, 1993). In the first order approximation here considered, this strain rate is assumed uniform over the area despite the fact that, due to the tectonic complexity of the region, large deviations are expected which can strongly affect local lithospheric strength.

Following Meissner (1986) and Christensen and Mooney (1995) we assumed that the upper crust is constituted by two layers respectively identified by V_p velocities lower than or equal to 5.7 and 6.4 km s^{-1}, (respectively corresponding to sediments and basement) and the mantle is characterized by V_p velocities greater than 7.8 km s^{-1}. For each layer we assumed a constant lithological composition and suitable values of rheological parameters (A, H and n in eqs. (2.4) and (2.5)) corresponding to the rock type dominant in the respective layer (table I). These values correspond to deformation of wet lithospheric rocks. This assumption, corroborated by some experimental results (Kirby and Kronenberg, 1987 and references therein), significantly affects the expected strength of the crust and to a larger extent, of the lithospheric mantle. This choice is of course debatable and different hypotheses have been adopted by other authors (Cloetingh and Burov, 1996).

By assuming that heat transfer is conductive, the temperature profile can be computed from surface heat flow and suitable values of the heat production Q and heat conductivity k (table II) within each layer (see, *e.g.*, Chapman, 1986; Meissner, 1986).

For the upper crust, the lower crust and the upper mantle density values of 2.6, 2.9 and 3.3 10^3 kg m^{-3} have been assumed respectively (see, *e.g.*, Meissner, 1986).

In the framework of the simplified layering here adopted, the strength envelope for each zone and the associated integral parameters can be simply related with the lithological stratification of the lithosphere, determined by geophysical observations, and the surface heat flow.

Table II. Parameters adopted to characterize the thermal properties of the lithosphere. Thermal conductivity (k) is expressed in W m^{-1} $°C^{-1}$. Heat production (Q) is expressed in mW m^{-3} (Fadaie and Ranalli, 1990).

	k	Q
Upper crust	2.5	1.20
Lower crust	2.5	0.40
Upper mantle	3.0	0.02

4. Rheological characters in the Mediterranean zone

The complex tectonic evolution of the Mediterranean area, involving the formation of orogenic belts and relatively large basins (see Mantovani *et al.*, 1997 and references therein) led to a considerable lateral heterogeneity of structural features and thermal regimes (fig. 1). To explore such an heterogeneous system we computed strength profiles in 15 sites distributed in the main structural provinces (fig. 1). The location of the selected sites has also been conditioned by the availability of sufficient and reliable structural and heat flow data (table III).

The data used have been taken from Duschenes *et al.* (1986), Berti *et al.* (1988), Della Vedova *et al.* (1989) and Morelli (1994) for the Central Mediterranean and from Makris (1978), Dachev and Volvosky (1985) and De Voogd *et al.* (1992) for the Eastern Mediterranean. Only heat flow data corrected for the effects of recent sedimentation and the presence of water in submerged areas have been taken into account. Very large heat flow values (> 200 mW m^{-2}) have not been considered since they might be representative of volcanic areas or associated to convective heat transfer. Data and respective uncertainties have been taken from the review papers by Della Vedova and Pellis (1989), Čermák (1993) and Cataldi *et al.* (1995).

By using the data and the parameters reported in tables II and III, «average» temperature profiles have been computed for the sites shown in fig. 1. In order to make more explicit the effects of the large uncertainties involved in these estimates, two additional profiles have been computed for each site. These profiles, hereafter referred to as «hot» and «cold» geotherms, correspond to the extreme temperatures which are compatible with experimental data. In particular, the «hot» geotherm has been obtained by considering the maximum observed heat flow (table III), a heat production increased by 20% and a conductivity decreased by the same percentage. The «cold» geotherm corresponds to the lowest observed heat flow, a decreased heat production (20%) and an increased heat conductivity (20%). These intervals of variation have been suggested by the spreading of the experimental values (Chapman, 1986). Some examples of the geotherms so obtained are shown in fig. 2. It is possible to note that the uncertainty on temperature rapidly increases with depth and reaches values of the order of hundreds of degrees at few tens of kilometres of depth.

The strength profiles obtained for the 15 sites here considered (fig. 1) are illustrated in figs. 3, 4 and 5. For each site, six profiles are reported to show the effects of the cold, average and hot geotherms and of tensional and compressional regimes. In order to provide an indication on the lithosphere as a whole, we also computed, using eqs. (2.7)-(2.9), the synthetic parameters L (lithospheric thickness) and R (total lithospheric strength), both for the «mechanical» and «thermal» lithosphere (table IV).

The results obtained indicate that critical temperatures defining the mechanical thickness of the lithosphere show a bimodal frequency distribution (peaked around 600° and 350°C respectively) due to the fact that the transition may occur within the mantle or the crust. Different choices of the geotherm produce large variations in the computed lithospheric thicknesses both concerning mechanical (4-94 km) and thermal lithospheres (10-261 km), in spite of the relatively limited lateral extent of the region considered (table IV). The total strength of the lithosphere in compressional regimes is more than two times higher than in tensional regimes. Both in tensional and compressional regimes, the thickness of the mechanical lithosphere, in spite of being generally less than one half of the one of the thermal lithosphere, is responsible for about 80-90% of the total lithospheric strength.

In general, the total strength of the lithosphere seems to be much more influenced by the thermal regime than by crustal structure. In fact, zones characterized by thin crust and low heat flow (*e.g.*, the Ionian and Levantine mesozoic basins) present much thicker and «stronger» mechanical lithospheres with respect to foreland areas (*e.g.*, Moesia and Adriatic). The thinnest (less than 20 km) and softest

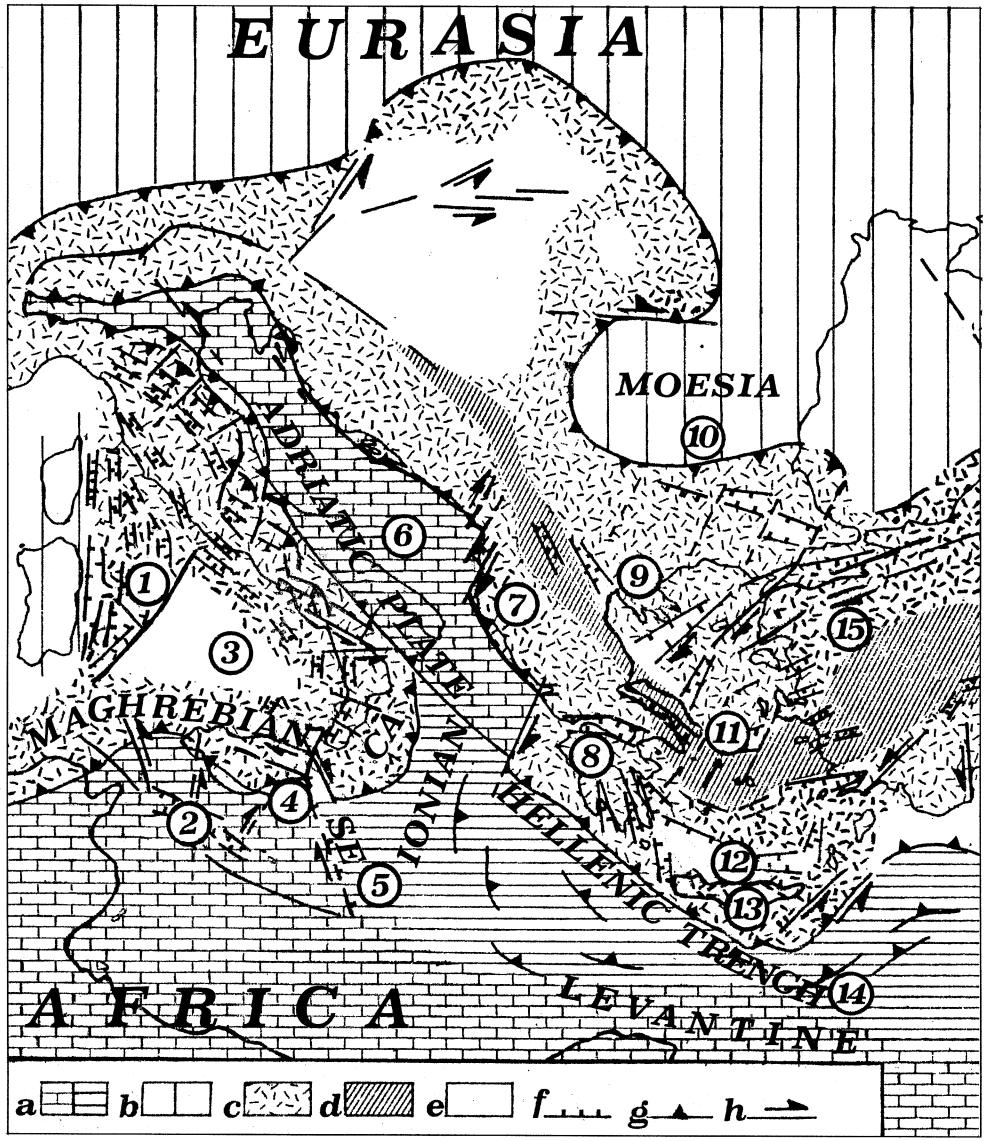

Fig. 1. Most important tectonic features of the Central-Eastern Mediterranean (a, b = African and Eurasiatic domains; c = orogenic belts; d = inner metamorphic massifs; e = oceanic or thinned crust; f, g, h = major tensional, compressional and transform features) and location of sites considered for the analysis of local rheological stratification of the lithosphere (1 = Northern Tyrrhenian; 2 = Sicily Channel; 3 = Southern Tyrrhenian; 4 = Iblean; 5 = Ionian basin; 6 = Adriatic; 7 = Northwestern Greece; 8 = Western Hellenic Arc; 9 = Macedonia; 10 = Moesia (Eurasia); 11 = Northern Aegean; 12 = Southern Aegean; 13 = Eastern Hellenic arc; 14 = Levantine basin; 15 = Western Turkey; SE = Syracuse escarpment).

Table III. Gross lithological stratification of the lithosphere for the sites shown in fig. 1, deduced from seismic and other geophysical data. For each site, average observed heat flow is reported along with the approximate range of values observed in the respective zone (see text for references).

No.	Site	Upper crust		Lower crust	Heat flux	
		Sediment thickness (km)	Basement thickness (km)	Thickness (km)	Average (mW/m^2)	Interval (min-max)
1	N. Tyrrhenian	1		8	100	75-150
2	Sicily channel	9	3	7	90	80-100
3	S. Tyrrhenian	1		6	150	100-200
4	Iblean	7	13	13	65	60-70
5	Ionian basin	8		6	40	35-50
6	Adriatic	12	8	15	55	50-60
7	NW. Greece	10	10	25	50	40-60
8	W. Hellenic arc	10	10	25	60	50-70
9	Macedonia		30	15	80	70-100
10	Moesia	7	8	17	50	40-60
11	N. Aegean	1	19	10	80	70-90
12	S. Aegean	2	10	8	70	60-80
13	E. Hellenic arc	5	19	6	60	50-70
14	Levantine basin	10		11	40	35-50
15	W. Turkey	4	20	20	80	70-90

lithospheres correspond to the zones affected by recent extensional regimes, such as the Sicily Channel, the Tyrrhenian, the Northern Aegean, the Macedonia region (Vardar trough) and Western Turkey (see, *e.g.*, Mantovani *et al.*, 1997 and references therein). A thin lithosphere (17 km) also results in a foreland zone, the Iblean plateau. This, however, is not surprising, since the Iblean area represents a fragment of the African foreland completely surrounded by active zones (the Syracuse escarpment, the Sicily Channel and the Maghrebian belt) and characterized by a higher than average heat flow (70 mW m^{-2}) probably connected with present and past magmatic activity (see, *e.g.*, Barberi *et al.*, 1974). The Southern Aegean basin presents a lithospheric thickness (28 km) which is somewhat greater than that observed in other extensional zones. This could be due to the fact that extensional activity in this zone ceased or slowed down some My ago (Mercier *et al.*, 1989; Meulenkamp *et al.*, 1994). At present, this zone is only characterized by minor deformations and seismic activity (Angelier, 1982; Jackson, 1993).

It is interesting to note that the eastern and western sectors of the Hellenic arc present significantly different lithospheric thicknesses (table IV), in spite of their lithological and genetic affinities. This difference may be explained by the fact that the western sector (Peloponnesus) is presently affected by a generalized extensional regime (Lyberis *et al.*, 1982), whereas the eastern sector (Crete-Rhodes) has been involved since the late Pliocene-early Pleistocene in a continental collision with the Lybian African margin (see *e.g.*, Armijo *et al.*, 1992; Chaumillon and Mascle, 1995).

From the tectonic point of view, a basic feature of strength profiles is the presence of in-

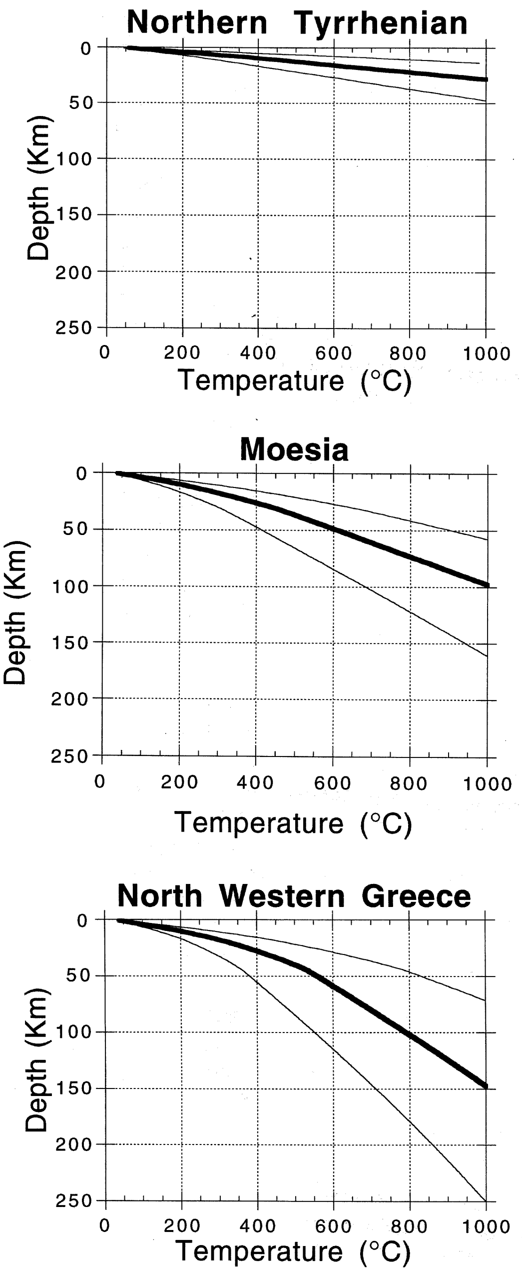

Fig. 2. Temperature profiles in three of the zones shown in fig. 1. For each zone, the thick line represents the «average» geotherm while the upper and lower thin lines respectively represent «hot» and «cold» geotherms (see text for details).

tra-lithospheric soft layers, which could behave as decoupling zones between the upper buoyant layer and the lower lithosphere, *i.e.*, the phenomenon which is often referred to as «delamination» or «ensialic subduction» (see, *e.g.*, Burchfiel, 1980; Boccaletti *et al.*, 1980; Bird and Baumgardner, 1981; Meissner, 1986; Van den Beukel, 1992; Mantovani *et al.*, 1997). The fact that the strength profiles in most Mediterranean zones here considered are characterized by pronounced soft intra-lithospheric layers (figs. 3 and 4), especially for average and hot geotherms, might thus imply that delamination processes have played an important role in the evolutionary pattern of this region. Of course, the reliability of this hypothesis and of all considerations on the possible influence of rheological profiles on evolutionary processes is strongly conditioned by the eventual changes that the rheological features of the various Mediterranean zones may have undergone in response to important changes in stress regimes. In any case, it seems interesting to note that the lack of soft intra-lithospheric layers in the Ionian and Levantine mesozoic basins could explain why in the trench zones where the Ionian/Levantine foreland is involved in subduction processes, as the Calabrian and Hellenic Arcs, the accretionary wedge is much more limited with respect to other orogenic belts, for example the Apennines, where the amount of imbricated crustal material is considerably larger.

In order to gain some insights into the role of the lithospheric mantle in Mediterranean tectonic processes, we determined the contribution of this layer to the total strength of the lithosphere in the sites considered (table V). These results suggest that in most zones, and in particular where a compressional environment is considered, the upper mantle makes a relatively minor contribution to the total lithospheric strength. Of course, these results largely depend on the assumptions underlying the present modelization, in particular as regards the wet mantle hypothesis. The assumption of dry mantle rocks leads to stronger lithospheric mantle which significantly contributes to the total strength (Cloetingh and Burov, 1996).

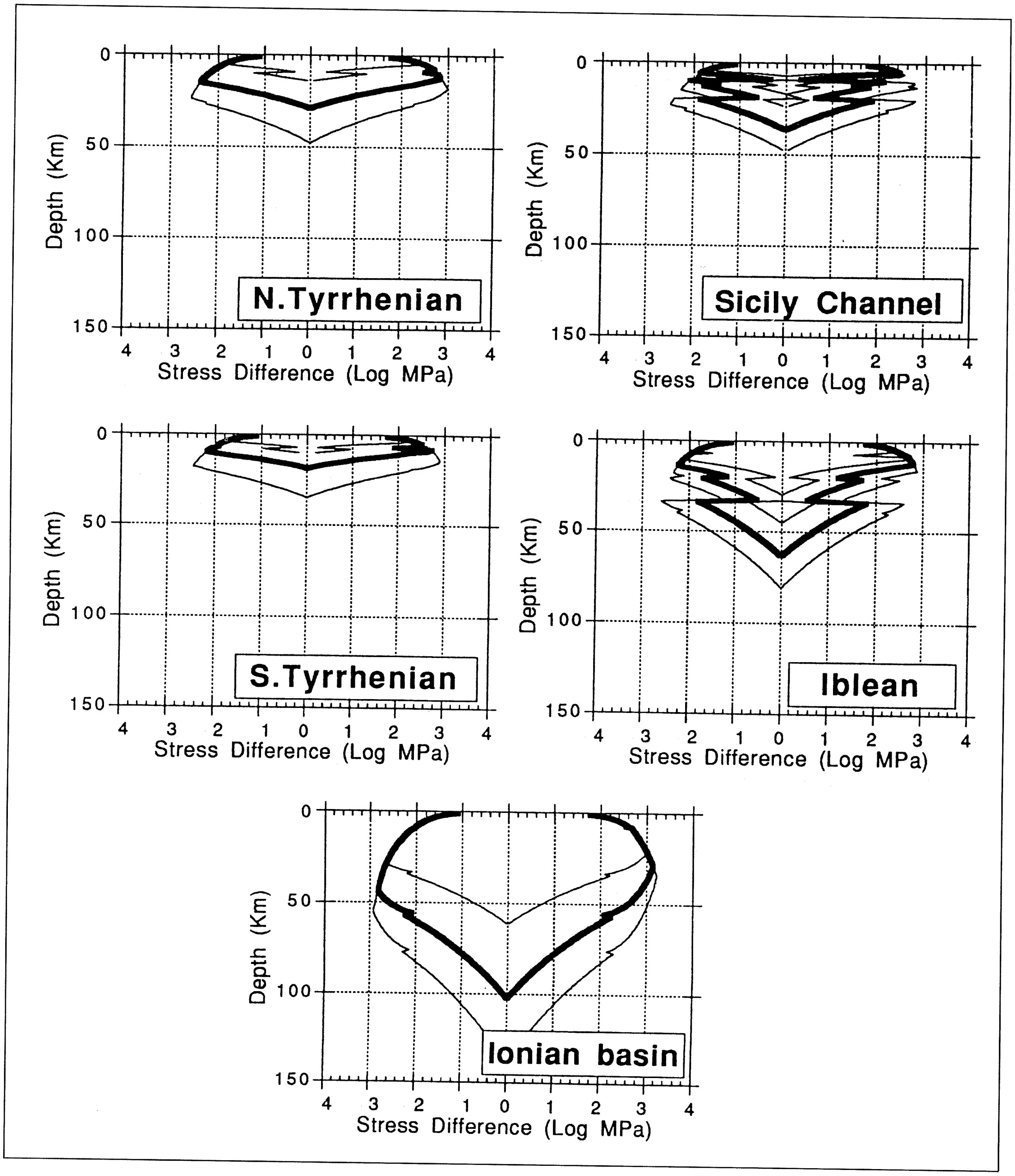

Fig. 3. Lithospheric strength envelopes at the sites from 1 to 5 in fig. 1 located in the Central Mediterranean. Decimal logarithms of stress difference values are plotted *vs.* depth. Left and right values in each plot respectively represent strength profiles in the tensional and compressional tectonic regimes. Thick lines are the strength envelopes computed using, for each site, the «average» geotherm. Inner and outer thin lines represent the strength envelopes computed using «hot» and «cold» geotherms respectively (see text for details).

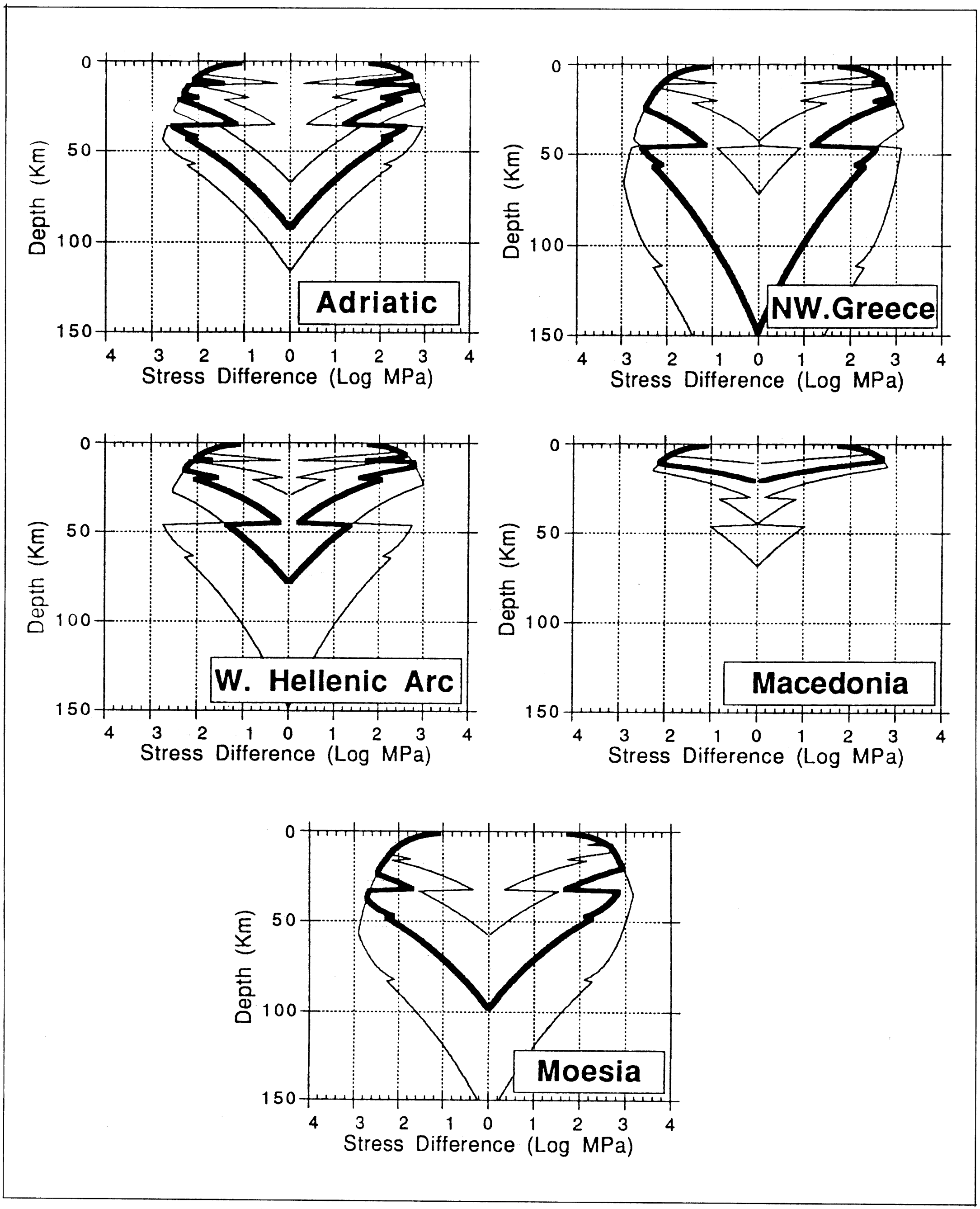

Fig. 4. Lithospheric strength envelopes at the sites from 6 to 10 in fig. 1 located in the Central Eastern Mediterranean (see caption of fig. 3).

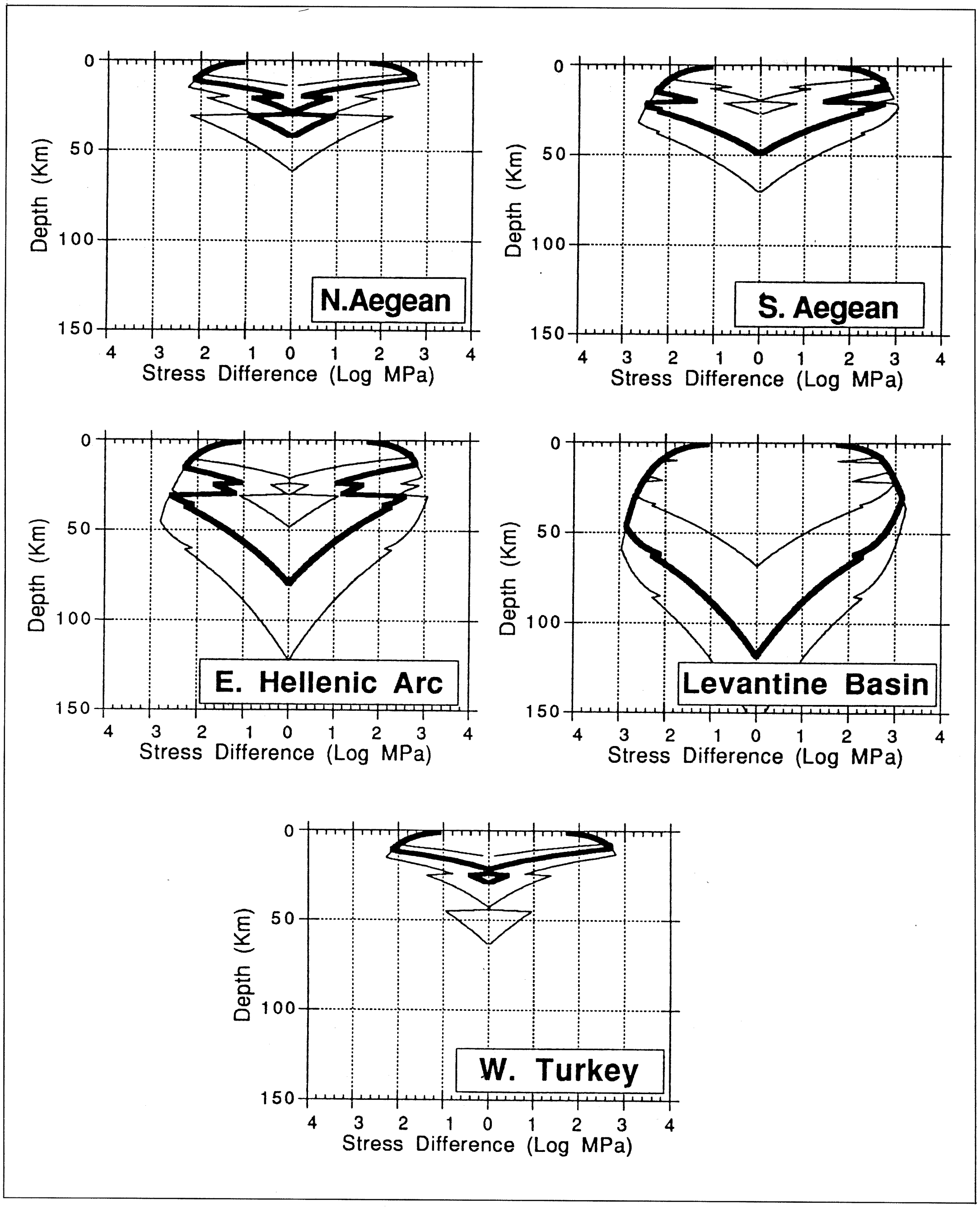

Fig. 5. Lithospheric strength envelopes at the sites from 11 to 15 in fig. 1 located in the Eastern Mediterranean (see caption of fig. 3).

Table IV. Synthetic parameters describing the most important rheological features of the lithosphere. The «thermal» and «mechanical» lithospheres are respectively defined by eqs. (2.7) and (2.8). L is the lithospheric thickness (km). T is the temperature (°C) at the lithosphere/asthenosphere boundary. Rc and Rt represent the total strength of the lithosphere (10^{12} N m^{-1}) in the compressional and tensional tectonic regimes respectively. These values are referred to the «average» geotherm. The figures in brackets in columns ΔL, ΔT, ΔRt and ΔRc are the values of L, T, Rt and Rc corresponding to «hot» and «cold» geotherms respectively.

No.	Site	L	ΔL	T	ΔT	Rt	ΔRt	Rc	ΔRc
					Mechanical Lithosphere				
1	N. Tyrrhenian	17	(5-29)	625	(389-635)	2.1	(0.2-5.6)	6.2	(0.8-14.8)
2	Sicily channel	11	(4-28)	387	(214-639)	0.5	(0.1-3.6)	1.1	(0.5-7.3)
3	S. Tyrrhenian	11	(4-21)	629	(416-627)	0.9	(0.1-3.1)	2.5	(0.5-9.2)
4	Iblean	17	(12-44)	392	(388-635)	1.6	(0.8-5.9)	5.2	(2.4-11.7)
5	Ionian basin	61	(36-84)	634	(632-639)	20.0	(7.5-34.8)	42.5	(18.7-67.6)
6	Adriatic	48	(15-64)	640	(389-640)	5.8	(0.7-16.2)	9.6	(1.7-28.8)
7	NW. Greece	67	(15-127)	639	(389-639)	9.6	(1.0-53.3)	17.2	(2.6-82.2)
8	W. Hellenic arc	22	(12-73)	432	(388-640)	2.1	(0.5-14.1)	5.4	(1.2-25.2)
9	Macedonia	12	(7-18)	369	(352-388)	0.9	(0.3-1.8)	3.3	(1.1-5.9)
10	Moesia	53	(17-91)	651	(427-655)	11.1	(1.5-35.3)	20.5	(4.3-66.1)
11	N. Aegean	12	(8-34)	369	(357-639)	0.9	(0.4-2.6)	3.3	(1.7-6.8)
12	S. Aegean	28	(8-41)	634	(357-638)	3.5	(0.4-9.3)	7.6	(1.7-21.6)
13	E. Hellenic arc	41	(12-68)	635	(388-641)	4.3	(0.8-20.0)	8.7	(3.0-35.2)
14	Levantine basin	69	(39-94)	639	(638-640)	23.2	(7.7-40.0)	47.3	(15.3-75.3)
15	W. Turkey	12	(8-18)	369	(357-388)	0.9	(0.4-1.8)	3.3	(1.7-5.9)
					Thermal Lithosphere				
1	N. Tyrrhenian	29	(14-48)	1007	(982-1137)	2.2	(0.3-5.8)	6.3	(0.9-15.1)
2	Sicily channel	37	(24-49)	1021	(1017-1024)	0.9	(0.3-3.9)	1.6	(0.5-7.6)
3	S. Tyrrhenian	19	(10-35)	1019	(955-1025)	1.0	(0.2-3.3)	2.6	(0.5-9.4)
4	Iblean	65	(47-83)	1022	(1020-1118)	2.2	(0.9-6.4)	5.9	(2.6-12.3)
5	Ionian basin	105	(63-142)	1024	(1023-1025)	21.0	(7.9-35.6)	43.1	(19.1-68.4)
6	Adriatic	96	(70-119)	1028	(1027-1030)	6.4	(1.2-17.0)	10.2	(2.2-29.5)
7	NW. Greece	154	(75-261)	1026	(1026-1027)	10.7	(1.3-55.1)	18.4	(2.9-84.0)
8	W. Hellenic arc	81	(46-154)	1021	(1020-1026)	2.6	(0.6-15.2)	6.0	(1.3-26.3)
9	Macedonia	46	(12-70)	1023	(568-1025)	1.0	(0.3-2.0)	3.4	(1.2-6.2)
10	Moesia	101	(59-166)	1024	(1017-1024)	11.8	(2.0-36.4)	21.1	(4.8-67.2)
11	N. Aegean	44	(21-63)	1022	(806-1023)	1.1	(0.5-3.0)	3.5	(1.8-7.1)
12	S. Aegean	50	(28-72)	1015	(1012-1021)	3.8	(0.6-9.7)	7.9	(1.9-22.0)
13	E. Hellenic arc	82	(50-126)	1020	(1024-1027)	4.9	(1.0-20.7)	9.2	(3.2-36.0)
14	Levantine basin	121	(70-163)	1023	(1022-1025)	23.9	(8.1-41.0)	48.0	(15.7-76.2)
15	W. Turkey	32	(16-66)	808	(648-1026)	1.0	(0.5-2.0)	4.4	(1.8-6.2)

It is interesting to note that the lithospheric thicknesses, deduced from surface wave analysis in the Mediterranean area (see *e.g.*, Panza and Suhadolc, 1990), are comparable with those obtained here for the thermal lithosphere. This correspondence, along with the fact that the temperatures at the base of the thermal lithosphere (see table IV) are slightly lower than the melting temperatures estimated for the wet peridotitic mantle (Kushiro *et al.*, 1968), could support the hypothesis that seismic low velocity zones are associated to partial melting in the upper mantle (Pollack and Chapman, 1977). An interesting discussion on the ambiguity which surrounds the concept of «lithosphere» in geophysical, geological and petrological literature is given by Anderson (1995).

It has been suggested that the frequency distribution of hypocentral depths of earthquakes is correlated with the strength profile of the lithosphere (*e.g.*, Meissner and Strelhau, 1982). However, this relationship seems to be rather vague from the experimental point of view (Deichmann and Rybach, 1989; Lamontagne and Ranalli, 1996; Okaya *et al.*, 1996) and unclear from the theoretical point of view (Scholz, 1990; Lamontagne and Ranalli, 1996). Due to these problems and to the fact that earthquake hypocentral depths routinely computed from seismological agencies and reported in the bulletins available to us (ISC, USGS, CSEM, etc.) cannot be considered fully reliable, this information has not been used to constrain rheological profiles in figs. 3 and 4.

Table V. Percentage of the total strength of the mechanical lithosphere supported by the lithospheric mantle for each site in fig. 1, computed for «cold», «average» and «hot» geotherms and for the tensional (t) and compressional (c) tectonic regimes on the assumption of wet rocks and uniform and constant strain rate. Empty cells correspond to the cases in which the boundary of the mechanical lithosphere is located within the crust and thus the contribution of the lithospheric upper mantle is null.

No.	Site	Cold		Average		Hot	
		t	c	t	c	t	c
1	N. Tyrrhenian	89	80	69	57	–	–
2	Sicily Channel	52	35	–	–	–	–
3	S. Tyrrhenian	87	79	57	47	–	–
4	Iblean	38	19	–	–	–	–
5	Ionian basin	96	91	93	86	82	70
6	Adriatic	62	41	41	25	–	–
7	NW. Greece	76	57	43	24	–	–
8	W. Hellenic arc	52	29	–	–	–	–
9	Macedonia	–	–	–	–	–	–
10	Moesia	81	63	58	35	–	–
11	N. Aegean	19	7	–	–	–	–
12	S. Aegean	71	55	48	27	–	–
13	E. Hellenic arc	72	57	49	24	–	–
14	Levantine basin	93	84	87	7	66	49
15	W. Turkey	–	–	–	–	–	–

5. Conclusions

Seismological stratifications of the crust and upper mantle and heat flow data in the Central-Eastern Mediterranean region have been used to compute first order rheological profiles in a number of sites representative of major structural provinces and stress regimes. Since strength envelopes are mainly controlled by the thermal regime, an attempt is made to estimate the influence of uncertainty in geotherms on the properties of the mechanical and thermal lithosphere.

Significant lateral heterogeneities in lithospheric thickness have been obtained for the area under study. As expected, the largest average mechanical lithospheric thicknesses (60-70 km) have been obtained in the «cold» zones, such as the Ionian and Levantine basins, a couple of Adriatic and Eurasia foreland sites and the overthickened collisional zone in North Western Greece, between the Adriatic block and the Balkan system. For heat flow values greater than 60 mW m^{-2} average mechanical lithospheric thicknesses are mostly thinner than 20 km. This occurs in recent basins (Tyrrhenian and Northern Aegean) and in extensional zones (Sicily Channel, Macedonia, Peloponnesus and Western Turkey). It is interesting to note that lithospheric thicknesses in extensional domains seem to depend on the time elapsed since the most recent extensional phase. This result can easily be interpreted in terms of transient cooling affecting young basins.

Within the limits of the simplified approach adopted here and on the assumption that rheological properties of lithosphere in the area under study have not significantly changed during the last My, strength envelopes can be used for tentative tectonic considerations on the Neogenic-Quaternary evolution. For this purpose, it seems interesting to note that most zones are characterized by strength envelopes with pronounced soft layers within the crust. This might imply that delamination processes have played a significant role in the recent Mediterranean evolution. Another important feature of rheological profiles is the fact that in many sites, and in particular for compressional regimes, the upper mantle seems to make a minor contribution to the total lithospheric strength.

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