# Geothermics of the Apenninic subduction

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**Abstract**
The subduction of the Adriatic microplate is analysed from a geothermal point of view. In particular four main geodynamic units are distinguished: foreland, foredeep and slab, accretionary prism, and back-arc basin. Each of them is examined from a geothermal point of view and the related open question are discussed. The most relevant results are the determination of the undisturbed geothermal gradient in the aquifer of the foreland; the discovery of a «hot» accretionary prism; and a new model of instantaneous extension of the back-arc basins. The main conclusion is that geothermal data are consistent with a westward dipping subduction that migrated eastward producing a sequence of several episodes at the surface.

**Key words** *Apennine chain – geothermics – subduction*

## 1. Introduction

The heat flow map of Italy and surrounding seas (fig. 1) is characterised by wide variations of heat flow over a limited surface. Some portions are relative to either areas that have not been involved in recent geodynamic phenomena, or to areas that are not particularly active at present that are:

– The Alpine region with a heat flow of about 60 $mWm^{-2}$ that results from a continent-continent collision and radiogenic heat produced within a doubled crust.

– Eastern Sardinia, with a heat flow of about 40 $mWm^{-2}$ (Mongelli *et al.*, 1992), is a microcontinent of ercinic age.

– The Ionian Sea, with a heat flow of about 40 $mWm^{-2}$, is a typical ocean of about 200 Ma (Della Vedova and Pellis, 1986).

The remaining portion of the map covers Central-Southern Italy, where the Adriatic microplate subducts westward beneath the European plate. It is characterized from east to west by a deep trench (mostly filled by sediments), an accretionary prism with average low elevation, a steep subduction angle and a wide back-arc basin. The heat flow is strictly related to the main geodynamic units that are:

– The foreland outcropping in the Apulian and Iblean platform with a heat flow of about 40 $mWm^{-2}$.

– The foredeep, represented by the Adriatic and Bradano Troughs, characterised by low values (less than 40 $mWm^{-2}$).

– The accretionary prism, represented by the Apennines, with low to high values (40 to 80 $mWm^{-2}$).

– The back-arc basin, represented by the Ligurian Sea, the Tuscan-Latial-Campanian pre-Apenninic belt, and the Thyrrenian Sea, with very high values (up to 200 $mWm^{-2}$).

Each of them presents problems of determination and interpretation of the heat flow.

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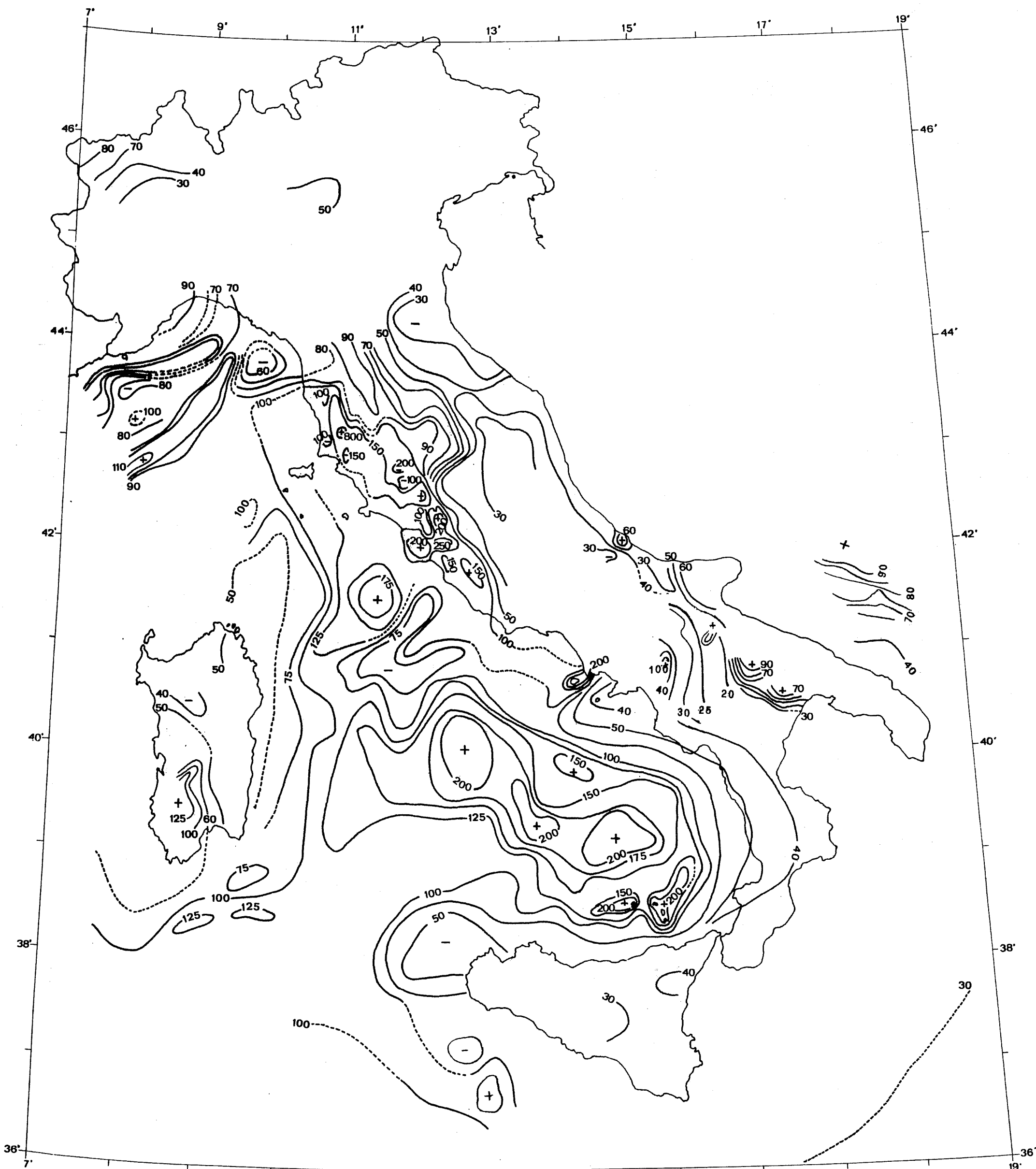


**Fig. 1.** Heat flow map of Italy and surrounding seas (after Mongelli *et al.*, 1991, revised).

## 2. Foreland

The foreland is generally assumed to be in a steady state from a geothermal point of view. The heat flow in this geodynamic unit is taken as reference value respect to which variations in the other units are studied. Consequently its accurate determination is of great importance.

The Apulian and Iblean platforms are constituted by limestones; there the heat flow is perturbed by ground-water flow. The determination of the undisturbed value can be performed according to a model proposed by Mongelli and Pagliarulo (1997). It is relative to a semi-infinite aquifer, with groundwater flowing parallel to the terrestrial heat flow in the recharge zone and perpendicular to heat flow in the remaining portion of the aquifer (fig. 2).

In the recharge zone the energy equation governing heat transfer at steady state is

$$\lambda \frac{\partial^2 T}{\partial z^2} = \rho c_p v_v \theta \frac{\partial T}{\partial z} \tag{2.1}$$

where $T$ is the temperature; $\lambda$ is the matrix thermal conductivity; $\rho$ is the water density; $c_p$ is the water specific heat capacity at constant pressure; $v_v$ is the water pore velocity, directed vertically downward; $\theta$ is the porosity.

In the horizontal portion of the aquifer, where the water moves with horizontal pore velocity $v_o$ and with the same values of $\theta$, $\lambda$, and $c_p$, the energy equation is

$$\lambda \left( \frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} \right) = \rho c_p v_o \theta \frac{\partial T}{\partial x}. \tag{2.2}$$

The temperature gradient at the surface is

$$\left( \frac{\partial T}{\partial z} \right)_{z=d} = G \left\{ -1 + \sum_{n=1}^{\infty} (-1)^n n\pi A_n \exp\left( \alpha_n \frac{x}{d} \right) \right\} \tag{2.3}$$

where

$d$ is the aquifer thickness;

$$A_n = \frac{2}{n\pi} \left\{ \frac{n^2\pi^2 [\exp(Pe_v) - (-1)^n]}{(n^2\pi^2 + Pe_v^2)\exp(Pe_v) - 1} + \right.$$

$$\left. + \frac{(-1)^n - 1}{\exp(Pe_v) - 1} - 1 \right\};$$

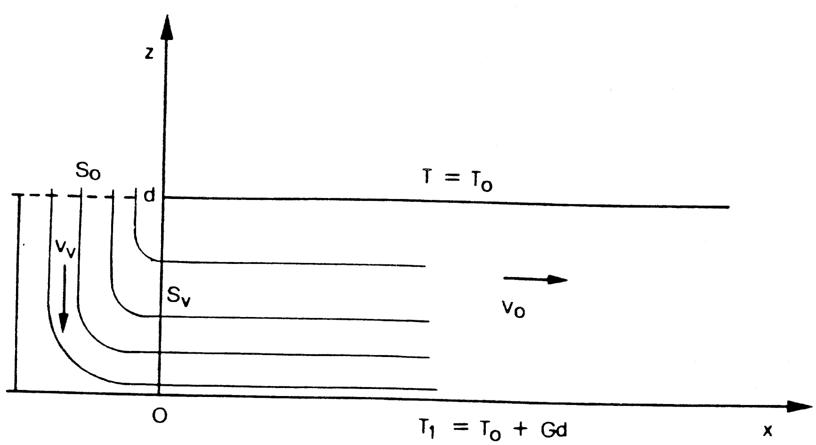


**Fig. 2.** Reference system of an unconfined semi-infinite aquifer; $v_v$ is the vertical velocity, of the groundwater; $T_0$ is the temperature of the water table at $d$ (aquifer thickness); $S_0$ is the surface of the recharge area; $T_0 + Gd$ is the temperature of the impermeable base. Streamlines are also drawn (after Mongelli and Pagliarulo, 1997).

$$\alpha_n = \frac{Pe_o}{2} - \sqrt{\frac{Pe_o^2}{4} + n^2\pi^2};$$

$Pe_v = \dfrac{\rho c_p v_v \theta d}{\lambda}$ is the Peclet number for the water vertical flow;

$Pe_o = \dfrac{\rho c_p v_o \theta d}{\lambda}$ is the Peclet number for the water horizontal flow.

Figure 3 shows the variation of the surface gradient with distance from the recharge zone for different values of $Pe_v$.

This model is applied to the Murgian plateau where the geothermal gradient has been measured in boreholes sited at different distances from the recharge zone. A fit of the observed gradient with different sets of curves where $Pe_o$ were fixed and $G$ let to vary, has yielded $G = 23$ °Ckm$^{-1}$ and a heat flow of 53 mWm$^{-2}$.

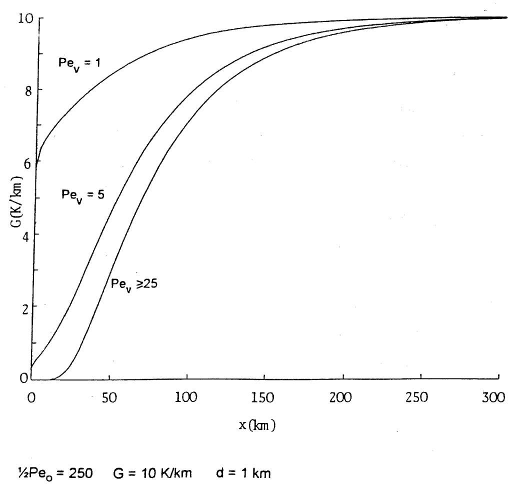


**Fig. 3.** Surface temperature gradient for different values of *Pe* numbers and aquifer thickness of 1 km. $Pe_v$ is the Peclet number for the water vertical flow in the recharge area (after Mongelli and Pagliarulo, 1997).

## 3. Foredeep and slab

This is the site of convergence of the plates and of the beginning of the subduction. The heat flow in this area is influenced by the combined effects of the subduction and the sedimentation process in the foredeep.

The thermal field within the subducted slab is governed by the energy equation

$$\lambda\left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2}\right) = \rho c_p v_x \frac{\partial T}{\partial x}. \tag{3.1}$$

With appropriate boundary conditions shown in fig. 4, the solution is (McKenzie, 1969)

$$T = T_1 \left\{1 + 2\sum_{n=1}^{\infty} \frac{(1)^n}{n\pi} \exp\left(\alpha_n \frac{x}{l}\right) \sin\left(n\pi \frac{z}{l}\right)\right\} \tag{3.2}$$

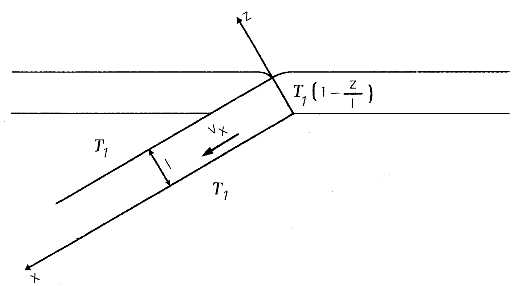


**Fig. 4.** Reference system of a subducted slab.

where

$T_1$ is the temperature of the isothermal mantle;

$l$ is the width of the slab;

$\alpha_n = R - \sqrt{R^2 + n^2\pi^2}$ where in turn $R = \dfrac{\rho c_p v_x l}{2\lambda}$

and $\rho$ is the density; $c_p$ is the specific heat capacity at constant pressure; $v_x$ is the subduction rate.

The temperature within the slab depends strongly on the subduction velocity through $\alpha$; a high subduction velocity corresponds to low temperatures. Figure 5 shows the isotherms within the slab calculated by eq. (3.2).

Many authors have proposed improvements to this simple but fundamental model. One of the most interesting is that proposed by Hsui and Tang (1988) that considers an upper mantle with a temperature gradient; in this case the angle of subduction becomes very important.

The thermal state is crucial in determining the relative slab weight according to

$$\rho_T = \rho_{T_1}(1 - \alpha_v \Delta T) \qquad (3.3)$$

where $\alpha_v$ is the coefficient of thermal expansion and $\Delta T = T - T_1$.

The weight of the slab is in fact considered by many people as the «engine» of plate tectonics. To properly estimate the slab weight it is necessary to model the geological structure of the subducting slab; then eq. (3.3) must be applied according to the various petrological associations within the slab. Theoretical and experimental works have been carried out by petrologists, yielding density distribution models like the one proposed by Irifune and Ringwood (1987) (fig. 6). These models also account for phase changes such as the olivine-spinel transition that occurs at about 400 km within the upper mantle and at shallower depth within the slab.

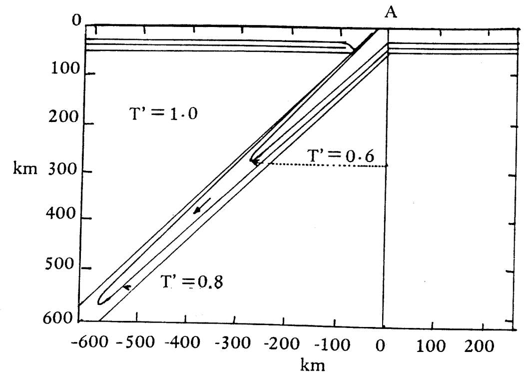


**Fig. 5.** Temperature within a subducted slab. $T' = T/T_1$ (after McKenzie, 1969).

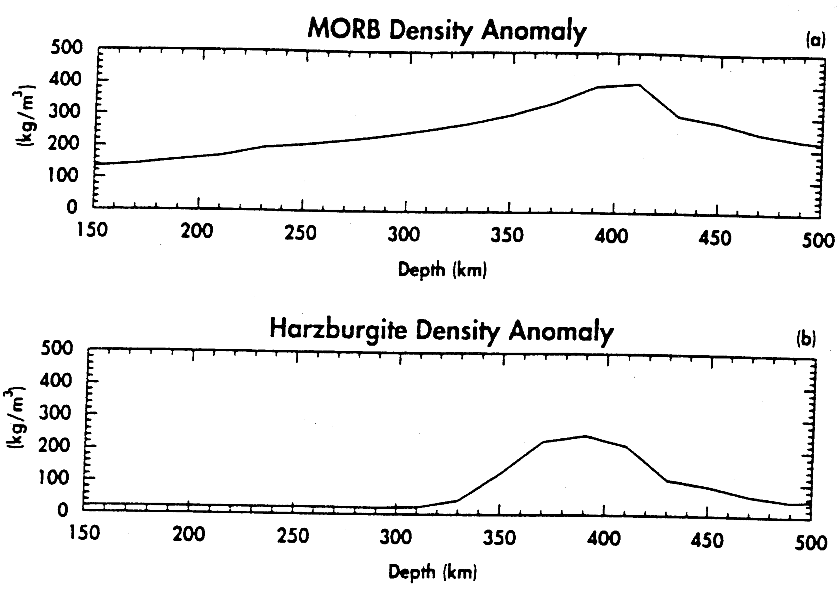


**Fig. 6.** Density difference between rocks of the subducting slab and surrounding mantle and its variation with depth (after Irifune and Ringwood, 1987).

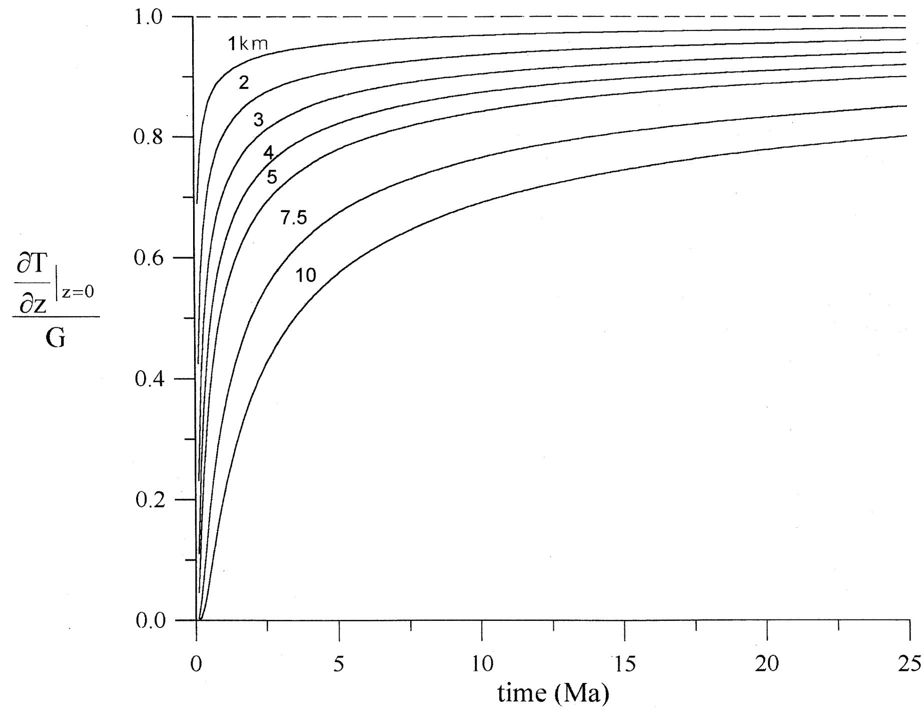


**Fig. 7.** Ratio between observed and undisturbed geothermal gradient for different thicknesses of sedimentary layers, and its variation with time (after Mongelli *et al.*, 1984).

A method of verifying the validity of curves of this type is that of calculating the gravity field given by a slab assuming a density distribution in accord with these models and to compare it with the observed gravity field at the surface. Harabaglia and Doglioni (1997) have investigated all the major west-dipping subductions worldwide. They stacked a series of satellite gravimetric profiles across these subductions, after appropriate multiple cross-correlation to filter out local effects; this stacking results in a background value of about 10 mgal. This phenomenon is still unexplained. Mongelli *et al.* (1997a) also verified that the density distribution of fig. 6 does not yield a good fit with the observed gravity anomaly in the Thyrrenian Sea.

In any case, the surface heat flow anomaly due to the deepening of the isotherms into the slab is masked by the effect of sedimentation in the foredeep. Sedimentation occurs slowly at sea bottom at variable sedimentation rates. However, if a considerable amount of time has elapsed since the sediments emersion, it is possible to consider the sedimentation as a sudden deposition (Birch *et al.*, 1968). In this case the surface temperature gradient is

$$\left(\frac{\partial T}{\partial z}\right)_{z=0} = G \cdot \operatorname{erfc}\frac{H}{2\sqrt{kt}} \tag{3.4}$$

where $G$ is the undisturbed gradient; $H$ is the thickness of deposited sediments; $t$ is the time elapsed since deposition, $k$ the thermal diffusivity.

Figure 7 shows how fast or slow, according to the thickness $H$, the restoration of the thermal equilibrium is, and gives the possibility, once $(\partial T/\partial z)_{z=0}$ is measured, to obtain $G$ (Mongelli *et al.*, 1984). Observed geothermal gradients of about 20 °Ckm$^{-1}$ on the eastern side of the Apennine chain have been corrected by this method and a value of about 35 °Ckm$^{-1}$ has been obtained (Mongelli *et al.*, 1989a). After the construction of the sedimentation curve, and the verification of the sequence of high and low sedimentation rates in different periods, it is sometimes possible to consider the whole pile of sediments as due to the stacking of single events (Mongelli *et al.*, 1997b).

It is noteworthy that the study of thermal effects of sedimentation have a very useful application in understanding oil maturation processes.

## 4. Accretionary prism

In the subduction process, the lithosphere is skinned and an accretionary prism is formed by vertical impilation of slices of limited thickness (Oxburg and Turcotte, 1974). The heat sources in this process are heating from the asthenosphere, radiogenic heating, and the frictional heating produced by the sliding of one slice over the other. It is here recalled that the surface erosion is another phenomenon that influences the thermal field and a further correction might be applied.

Many authors have studied the case of a sudden thrusting with different boundary conditions (*e.g.*, Molnar *et al.*, 1983 and references therein). More recently, Royden (1993) studied the case of a continuous stationary accretion. However, in some cases two features have been evidenced in recent years:

a) The thrusting is quite often episodic.

b) The slices are thin and mostly constituted by upper crust and sedimentary rocks; in some cases sedimentary rocks represent the bulk of the prism.

Under this hypothesis, the heat produced by the rock friction decays very rapidly (Brewer, 1981; Mongelli *et al.*, 1989b).

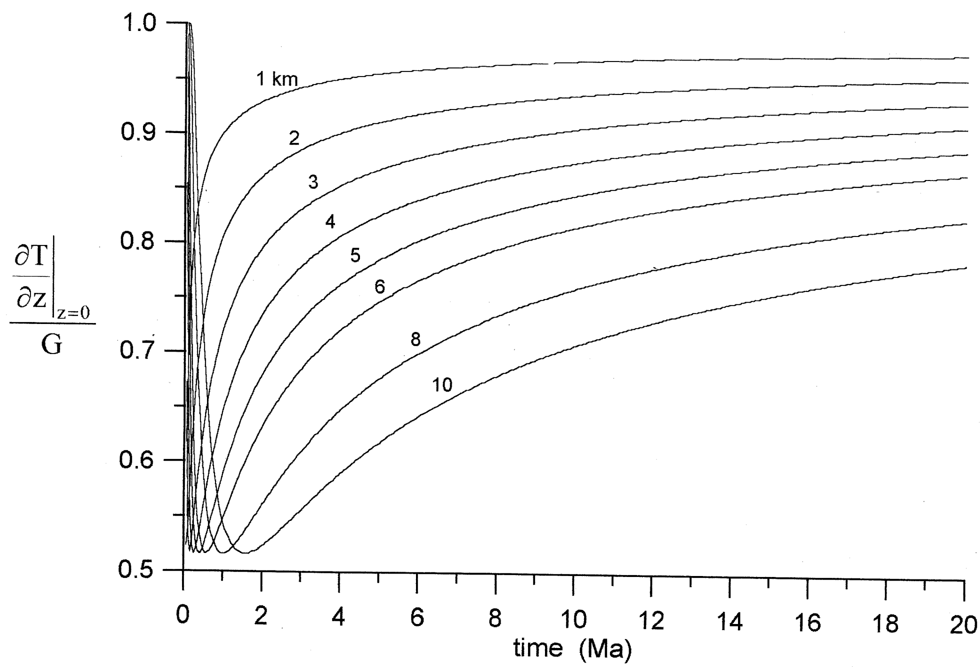


**Fig. 8.** Example of restoration of the geothermal gradient after emplacement of thrusts of different thicknesses.

The surface geothermal gradient in the case of a single thrust is

$$\left(\frac{\partial T}{\partial z}\right)_{z=0} = -G\left\{\frac{H}{\sqrt{\pi k t}}\exp\left(-\frac{H^2}{4kt}-1\right)\right\} \quad (4.1)$$

where $H$ is the thickness of the overthrusted slice. Figure 8 shows the restoration of surface gradient for particular values of the parameters involved. It is possible to observe that during the restoration, the gradient is less than the reference value.

Mongelli *et al.* (1989b) explained the low heat flow values observed on the eastern side of the Apennine chain by considering the stacking effect of two sudden episodes of thrusting. The second episodes further lower the value of gradient.

More recently, new heat flow measurements over a strip crossing the Southern Apennines evidenced high values on the western side of the chain (Mongelli *et al.*, 1996). This is consistent with a model (Doglioni *et al.*, 1996) of an accretionary prism mostly warmed laterally from the hot back-arc basin (fig. 9). The authors proposed a thermal model of a semi-infinite plate

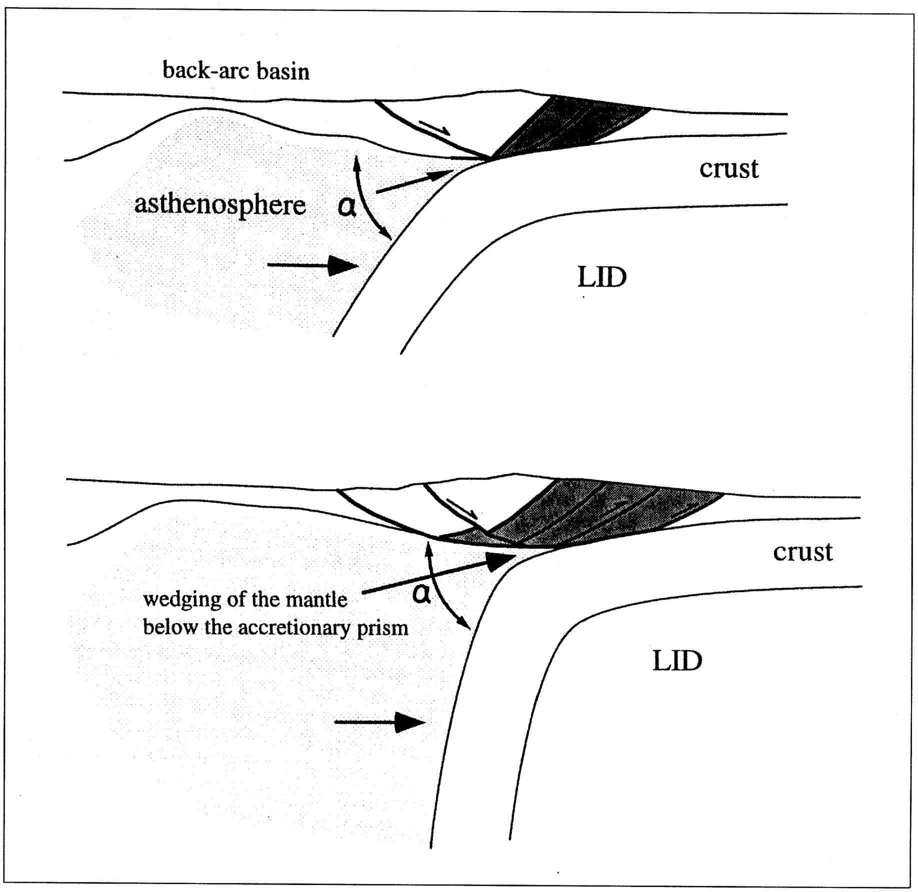


**Fig. 9.** The accretionary prism floored in its western side by the hot mantle and in contact with the warm thinned lithosphere to the west (after Doglioni *et al.*, 1996).

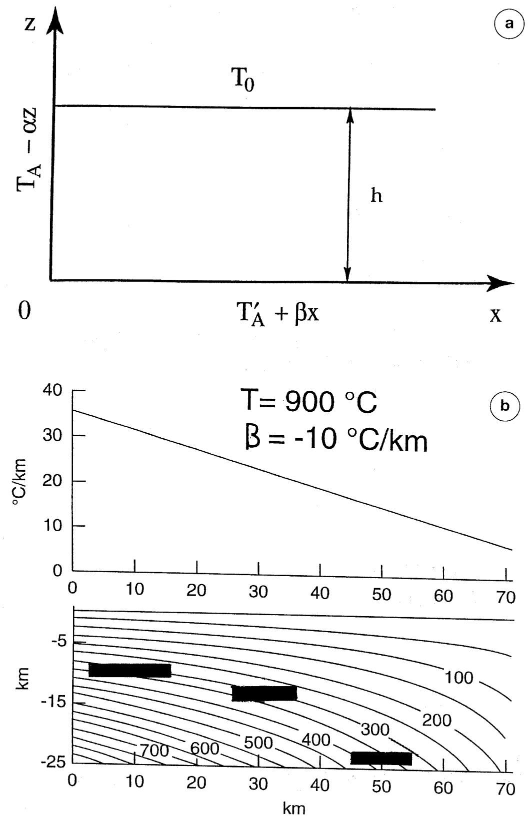


**Fig. 10a,b.** a) Idealised thermal model of the accretionary prism. $T_0$ is the surface temperature; $T_A - \alpha z$ is the decreasing temperature at the boundary of the overriding plate; $T_A' + \beta x$ is the increasing temperature at the base of the prism; $h$ is the thickness of the prism. b) Isotherms and surface gradient of an accretionary prism for $T_0 = 10$ °C, $T_A = 900$ °C, and $\beta = -10$ °Ckm$^{-1}$. Solid lines represents deep aftershocks of the northern and southern portion of the 1980, Irpinia, and 1990, Potenza, earthquakes and are supposed to mark the brittle-semibrittle transition (after Doglioni *et al.*, 1996).

in steady state with boundary conditions of fig. 10a; the surface temperature gradient is

$$\left(\frac{\partial T}{\partial z}\right)_{z=h} = \frac{T_0 - T_A' - \beta x}{h} +$$

$$+2\,(T_A' - T_A) \sum_{n=1}^{\infty} (-1)^{2n+1} \exp\left(-n\pi \frac{x}{h}\right). \quad (4.2)$$

The isotherms and surface gradients in the area (fig. 10b) were obtained with the seismological constraints given by the deeper aftershocks of the 1980 Irpinia and 1990 Potenza earthquakes that were assumed to mark the lower limit of the brittle portion of the crust in the area. This model is consistent with the low velocity anomaly in the crust extending underneath the whole Apennine chain (Chiarabba and Amato, 1996).

Also in this case the geothermal study is important to understand the oil maturation process.

## 5. Back-arc basin

The high heat flow in the back-arc basin is due to the thinning of the lithosphere. The most intriguing problem is how and why the lithosphere thins. It is generally assumed that there is no active asthenospheric heating of the lithosphere; on the contrary, thinning is considered of mechanical origin with consequent «passive» rising of the asthenosphere.

McKenzie (1978) modelled the thermal effects of uniform *pure shear extension* of the entire lithosphere. In this classic one-dimensional model (fig. 11) a lithosphere with vertical thickness $a$ and temperature gradient $G$, is uniformly stretched by a factor $\beta$ and correspondingly thinned to a final thickness $a/\beta$. In this model the temperature of the lithosphere increases by advection during the extension to assume a new gradient $G \cdot \beta$.

The temperature during cooling is

$$T = T_1 \left\{ 1 - \frac{z}{a} + \frac{2}{\pi} \sum_{n=1}^{\infty} \frac{(-1)^{n+1}}{n} \left[ \frac{\beta}{n\pi} \sin \frac{n\pi}{\beta} \right] \cdot \right.$$

$$\left. \cdot \exp\left(-\frac{n^2 t}{\tau}\right) \sin\left(\frac{n\pi z}{a}\right) \right\} \quad (5.1)$$

where $\tau = \dfrac{a^2}{\pi^2 k}$.

The model allows for the calculation of the initial tectonic subsidence, the thermal subsidence, and the decrease with time of the man-

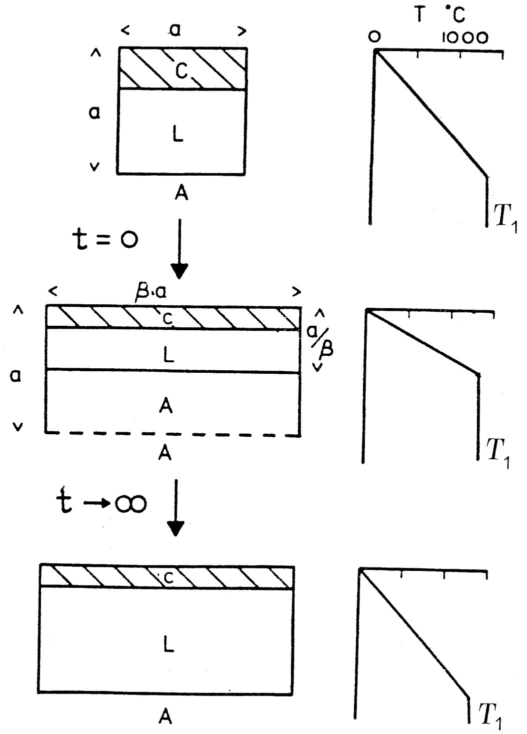


**Fig. 11.** Stretching of the lithosphere. $T_1$ is the temperature of the asthenosphere (after McKenzie, 1978).

tle heat flow at the surface. The latter (fig. 12) is given by

$$q(t) = \frac{\lambda T_1}{a}\left\{1 + 2\sum_{n=1}^{\infty}\left[\frac{\beta}{n\pi}\sin\frac{n\pi}{\beta}\right]\exp\left(-\frac{n^2 t}{\tau}\right)\right\}. \tag{5.2}$$

This model has been applied to the Tyrrhenian Sea (Hutchison *et al.*, 1985), Tuscan-Latial pre-Apennine belt (Mongelli *et al.*, 1989b) and the Northwestern Mediterranean (Pasquale *et al.*, 1995). Some authors (*e.g.*, Cochran, 1983) have extended this method to the 2D case.

This model however leaves unresolved the important question of how the thinning advances within a finite time interval, that is, if it does it vertically or horizontally. Jarvis and McKenzie (1980) studied the effects of vertical heat loss during a finite duration extension event by moving the base of the lithosphere upward. Cochran (1983) combined the effects of lateral heat conduction and finite duration of extension by considering the lithosphere divided into a series of narrow blocks each of which stretched by different amounts in subsequent times; in this way the author simulated the extension as a series of discrete short rifting events.

However, recent geological and geophysical observations have emphasised the importance of *simple shear* extension in the crust which gives rise to asymmetric structures. Moreover, a back-arc basin like the Northwestern Mediterranean (Doglioni *et al.*, 1997) seems formed by a series of sub-basins. In correspondence of them the lithosphere thins so that its base is marked by a series of undulations, as if this propagates eastward in subsequent episodes. This may indicate that the extension of the lithosphere and the upwelling of the asthenosphere is a pulsating phenomenon which occurs following the roll-back of the subducting slab. It is possible that each sub-basin may extend either by pure shear or by simple shear. Mongelli and Zito (1997) have proposed a model of instantaneous stretching according to

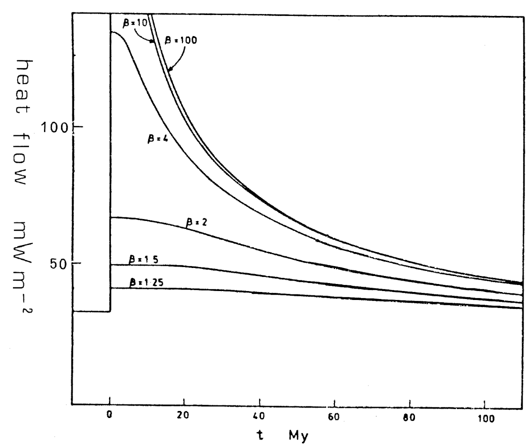


**Fig. 12.** Surface heat flux as a function of time of a stretched lithosphere for various values of $\beta$ (after McKenzie, 1978).

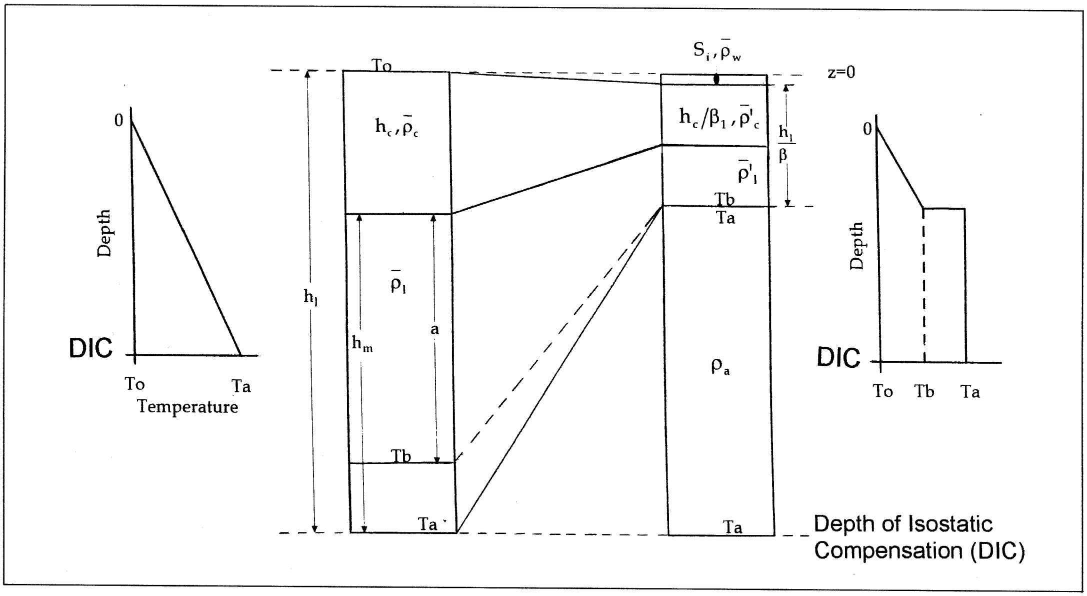


**Fig. 13.** Instantaneous stretching of the lithosphere. $h_c$ is the thickness of the crust; $h_l$ is the thickness of the lithosphere; $a$ is the portion of the LID which is stretched; $T_b$ is the temperature at its base (after Mongelli and Zito, 1997).

which the lithosphere thins partly by advective distension and partly by sub-lithospheric erosion (fig. 13).

By application of the isostatic principle it is possible to calculate both the tectonic subsidence

$$S_i = \frac{h_c\left(\frac{\bar{\rho}'_c}{\beta_1} - \frac{\bar{\rho}'_l}{\beta_1} + \bar{\rho}_l - \bar{\rho}_c\right) + h_l\left(\frac{\bar{\rho}'_l}{\beta} - \frac{\rho_a}{\beta} + \rho_a - \bar{\rho}_l\right)}{\rho_a - \rho_w} \tag{5.3}$$

where in general

$$\rho = \rho_0\,(1 - \alpha_v \Delta T)$$

and $\bar{\rho}$ is the average density corresponding to the average temperature of each layer, and the temperature at the base of the stretched lithosphere

$$T_b = T_a\,\frac{\beta_1}{\beta}$$

as a consequence of the plastic distension. After the extension, the thinned lithosphere warms up while $T_b$ tends to $T_a$, the temperature of the asthenosphere. The temperature is

$$T(z, t) = T_0 + \frac{T_a - T_0}{h_l/\beta}\,z + \frac{2}{\pi}(T_a - T_b)\sum_{n=1}^{\infty}\frac{(-1)^n}{n}\cdot$$

$$\cdot \sin\frac{n\pi z}{h_l/\beta}\exp\left[-\frac{kn^2\pi^2 t}{(h_l/\beta)^2}\right] \tag{5.4}$$

and the surface temperature gradient is

$$\left(\frac{\partial T}{\partial z}\right)_{z=0} = \frac{T_a - T_0}{h_l/\beta} + \frac{2}{h_l/\beta}(T_a - T_b)\sum_{n=1}^{\infty}(-1)^n\cdot$$

$$\cdot \exp\left[-\frac{kn^2\pi^2 t}{(h_l/\beta)^2}\right] \tag{5.5}$$

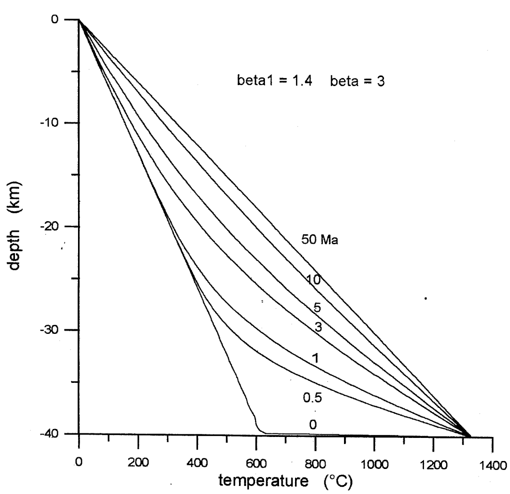


**Fig. 14.** Isotherms in the stretched lithosphere at different times after instantaneous stretching (after Mongelli and Zito, 1997).

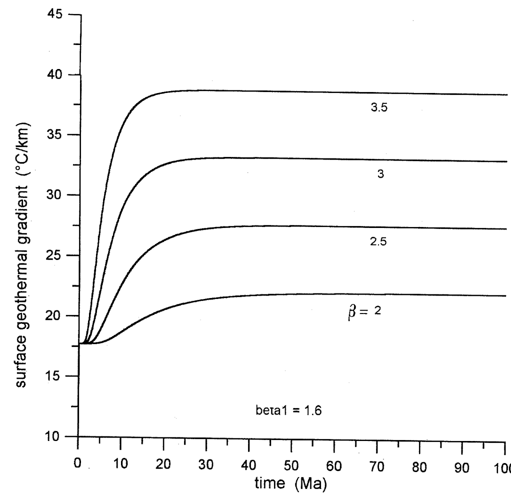


**Fig. 15.** Variation with time of the geothermal gradient in the stretched lithosphere for different values of $\beta$ (after Mongelli and Zito, 1997).

where $T_0$ is the surface temperature; $h_l$ is the thickness of the lithosphere; $\beta$ is the stretching factor of the lithosphere; $\beta_1$ is the stretching factor of the crust.

Equations (5.4) and (5.5) are represented in figs. 14 and 15; equilibrium is reached after 15-20 Ma, then cooling begins according to the McKenzie eqs. (5.1) and (5.2).

This model allows a calculus of the surface uplift by dilatation (fig. 16), that is

$$\Delta l(t) = \alpha_v \int_0^{h_l/\beta} \Delta T(z, t) \cdot dz =$$

$$= \frac{1}{2}\alpha_v h_l (T_a - T_0)\left(\frac{\beta - 1}{\beta^2}\right) + \qquad (5.6)$$

$$+ \frac{2}{\pi^2}\alpha_v T_a \left(\frac{1-\beta_1}{\beta^2}\right) h_l \sum_{n=1}^{\infty} \frac{[-1+(-1)^n]}{n^2} \exp\left[-\frac{k n^2 \pi t}{(h_l/\beta)^2}\right].$$

In particular the application of this model to the Tuscan region explains a tectonic subsidence of about 1.2 km and an uplift of about 400 m.

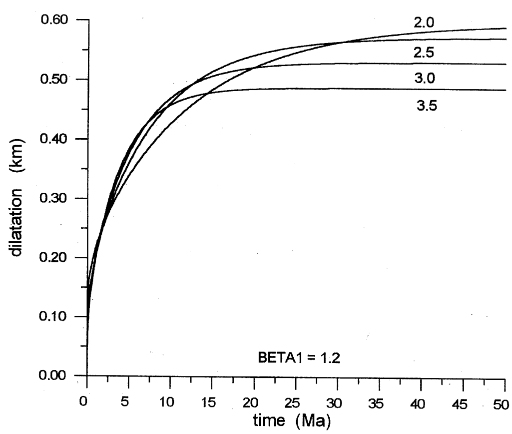


**Fig. 16.** Surface uplift by dilatation of the lithosphere for different values of $\beta$ and different times after stretching (after Mongelli and Zito, 1997).

## 6. Conclusions

The main guidelines of how the geothermal research in Italy should develop are:

a) A study of rock density variations with pressure and temperature to better evaluate the slab weight.

b) A world-wide analysis to validate the existence of «warm» accretionary prisms in westward subductions because of the implication on seismicity and oil maturation.

c) The understanding of the relationship between rheological properties of the lithosphere and any given geodynamic-geothermic setting, as preliminary done by Dragoni *et al.* (1996, 1997).

d) The 2 and 3D modelling of symmetric (pure shear) and asymmetric (simple shear) back-arc basins.

The final remark is relative to the fact that most of the geodynamic phenomena involved are episodic; this will require new modelling approaches.

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