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Constraints on thermal models of the Basin and Range province

JEAN-CLAUDE MARESCHAL [1] and GEORGE BERGANTZ [2]

[1] *Département des Sciences de la Terre and GEOTOP, Université du Québec à Montréal, Montréal, Que. H3C 3P8, (Canada)*
[2] *Department of Earth and Planetary Sciences, Johns Hopkins University, Baltimore, MD 21218 (U.S.A.)* *

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Abstract

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Several mechanisms that have been proposed to explain the high heat flow in regions of tectonic extension are examined. The transient thermal regime is computed and the results are discussed. Conductive heating of the lithosphere from below requires a time of at least 75 Ma to affect the surface heat flow. Advective heating by the injection of magmas into the lithosphere does not produce steady-state conditions before 75 to 300 Ma (depending on boundary conditions). Only mechanical stretching of at least the shallow lithosphere can explain the increase in surface heat flow in a time of the order of 30 to 50 Ma. The heat flow data are compatible with several mechanisms for the extension of the lower lithosphere (underplating or dike injection) and other geologic and geophysical data are needed to constrain the models.

Introduction

The heat flow data collected in the southwestern United States (e.g., Sass et al., 1971; Reiter et al., 1979; Bodell and Chapman, 1982) have established clear patterns in the thermal regime of the lithosphere. The correlation proposed between heat flow and tectonic provinces (Roy et al., 1968) has been established but, at the same time, additional complexities have been revealed in the thermal regime. The mean heat flow in the Basin and Range province and in the Rio Grande Rift, which have undergone extension during the past 50 Ma, is about 40 mW/m^2 higher than in stable continental regions; on the other hand, with the exception of localized anomalies, the heat flow is not significantly higher in the central Colorado plateau (Bodell and Chapman, 1982) although it was uplifted, perhaps in several episodes, during the past 30 Ma. Within the Basin and Range, the heat flow exhibits extreme local variations and it is 100 mW/m^2 higher than average in the Battle Mountain subprovince (Lachenbruch and Sass, 1978).

The interpretation of the heat flow data in recently active provinces is extremely ambiguous; downward continuation of the heat flow (Mareschal, 1989) is model dependent (i.e., heat sources distribution, variations in thermal conductivity) and, most importantly, it requires hypotheses, such as conductive equilibrium that are unlikely to be verified in a tectonically active and environment.

Crough and Thompson (1976) and Pollack and Chapman (1977) have used the reduced heat flow to infer the lithospheric thickness; they assumed that the anomalous surface heat flow is equal to the flow conducted at the base of the lithosphere. These assumptions are appropriate for the authors'

* Present address: Department of Geological Sciences, University of Washington, Seattle, WA 98195 (U.S.A.).

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purpose, but they cannot be used to determine the thermal regime in regions that have been active recently and have been affected by mechanical extension and heat transport by magmas. In addition, the amplitude of the thermal perturbations and the time required to reach equilibrium is too long for these conductive models to be applicable. Lachenbruch and Sass (1977, 1978; see also Lachenbruch, 1978) have proposed several mechanisms that could explain the heat flow anomaly in the context of an extending lithosphere, and they derived steady-state solutions for the thermal regime of a plate that is either (1) uniformly stretched, or (2) stretched and underplated by solidifying basalt and (3) extended and intruded by dikes. They discussed the feasibility of these three mechanisms and concluded that these three mechanisms, individually or as a combination, could explain the Basin and Range heat flow.

The purpose of the present paper is to examine the evolution of heat flow predicted by these models. Analytical solutions are used to describe the effect of the accretion of hotter material of the base of the lithosphere and of the advection by magma intrusions. The effect of mechanical stretching (combined or not with intrusions and underplating) is determined numerically. The study supports the main conclusions of Lachenbruch and Sass (1978) and demonstrates that a steady-state regime is not reached before at least 80 and that stretching of the lithosphere at a high rate is the most efficient mechanism to produce a rapid increase in heat flow, and that it could be combined with dike intrusions in the lithospheric mantle or with underplating.

Conductive models of the lithosphere

Several physical mechanisms have been proposed to explain the change in the thermal regime of the lithosphere and the increased surface heat flow. The lithosphere could move over hotter asthenosphere or magmas could rise in the asthenosphere, settle and solidify at the base of the lithosphere; in both cases, additional heat would be conducted into the lithosphere. It is also conceivable that the source of heat is shallower

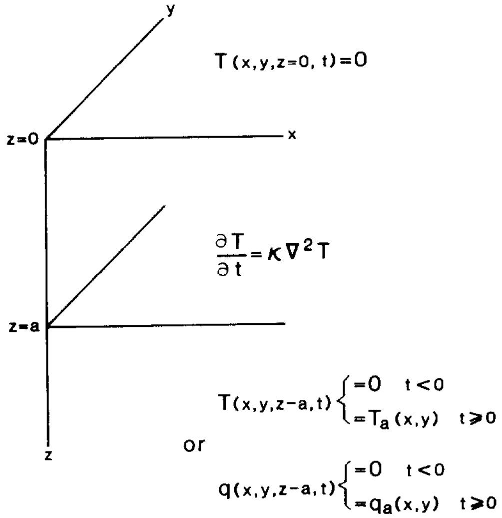

Fig. 1. The geometry, equations, and boundary conditions for the conductive heating model.

than the LAB (lithosphere–asthenosphere boundary). For instance, the magmas could rise through the lower lithosphere and settle at the base of the crust, or the asthenosphere could replace the lithospheric mantle by delamination (Bird, 1979) or diapiric uprise (Neugebauer, 1983; Mareschal, 1983a). All these mechanisms imply the diffusion of heat from the base of a slab to its surface, with different boundary conditions.

In order to determine the effect of a thermal perturbation at the base of a slab, is considered a three-dimensional model of the lithosphere (Fig. 1). A Cartesian coordinate system x, y, z is used; z is vertical, positive downward, $z=0$ is the surface, and $z=a$ is the LAB.

The temperature perturbation, $T(x, y, z, t)$, in the slab is determined by the heat equation (Carslaw and Jaeger, 1959):

$$\frac{\partial T}{\partial t} = \kappa \nabla^2 T + \frac{H}{\rho C} \qquad (1)$$

where ρ is the density, κ is the thermal diffusivity, H is the distribution of heat sources, and C is the heat capacity.

The temperature is constant at the surface $z=0$:

$$T(x, y, z=0, t) = 0 \qquad (2a)$$

and either the temperature or the heat flow changes at the LAB $z=a$:

$$T(x, y, z=a, t) = \Delta T_a(x, y, t) \qquad (2b')$$

or:

$$q(x, y, z=a, t) = \Delta q_a(x, y, t) \qquad (2b'')$$

Initially, there is no temperature perturbation:

$$T(x, y, z, t=0) = 0 \qquad (3)$$

It is convenient to use integral transforms to solve this system of equations, boundary and inital conditions (see Mareschal, 1981, for a complete derivation). If we consider first a stepwise change in heat flow at the lower boundary; i.e.:

$$\Delta q_a = 0 \qquad t<0$$
$$= \Delta q_a(x, y) \qquad t>0$$

It can be decomposed into its Fourier components and the surface heat flow can be determined for each of the Fourier components as (Mareschal, 1981):

$$\Delta q_0(k_x, k_y, t) = \Delta q_a(k_x, k_y)\left\{\frac{1}{\sinh ka} - 4\exp(-\kappa k^2 t) \times \sum_{n=0}^{\infty} \frac{(-1)^n(2n+1)\pi}{4k^2a^2 + (2n+1)^2\pi^2} \times \exp\left[-(2n+1)^2\pi^2\kappa t/4a^2\right]\right\} \qquad (4)$$

where $k=\sqrt{k_x^2+k_y^2}$ is the wavenumber.

Likewise the surface heat flow following a stepwise change in temperature at the lower boundary can also be determined for each wavevector (k_x, k_y) of the perturbation (Mareschal, 1987).

$$\Delta q_0(k_x, k_y, t) = \frac{K\Delta T_a(k_x, k_y)}{a}\left\{\frac{ka}{\sinh ka} - 2\exp(-\kappa k^2 t) \times \sum_{n=1}^{\infty} (-1)^{(n+1)} \frac{n^2\pi^2}{k^2a^2 + n^2\pi^2} \times \exp(-n^2\pi^2\kappa t/a^2)\right\} \qquad (5)$$

where K is the thermal conductivity.

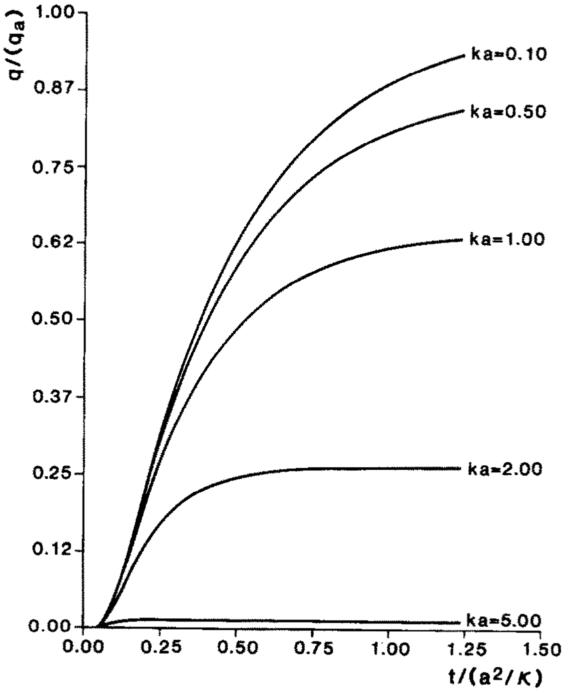

Fig. 2. The surface heat flow at time t after a stepwise change in heat flow at the lower boundary for different dimensionless wavenumbers (i.e., wavenumber times slab thickness). The time unit is a^2/κ, (300 Ma for the lithosphere, 50 for the crust); the heat flow unit is the amplitude of the lower boundary perturbation.

Figure 2 shows the change in surface heat flow as a function of time after a stepwise unit change in heat flow at the lower boundary for perturbations with varying wavenumber. The time unit is a^2/κ; this is the order of the time constant for a thermal perturbation at the base of the lithosphere to reach the surface; the actual value of the time constant depends on boundary conditions.

Figure 3 shows the change in surface heat flow after a stepwise change of temperature at the lower boundary. The heat flow is divided by KT_a/a; this is the steady state heat flow perturbation that would be observed if the temperature were raised uniformly by T_a on the plane $z=a$.

It can be observed that the perturbations with large wavenumbers (i.e., with a wavelength short compared to the thickness of the slab) are seriously attenuated. In other words, the lithosphere acts like a filter that attenuates short wavelength perturbations. In general, the perturbations of geophysical interest contain more than one wave-

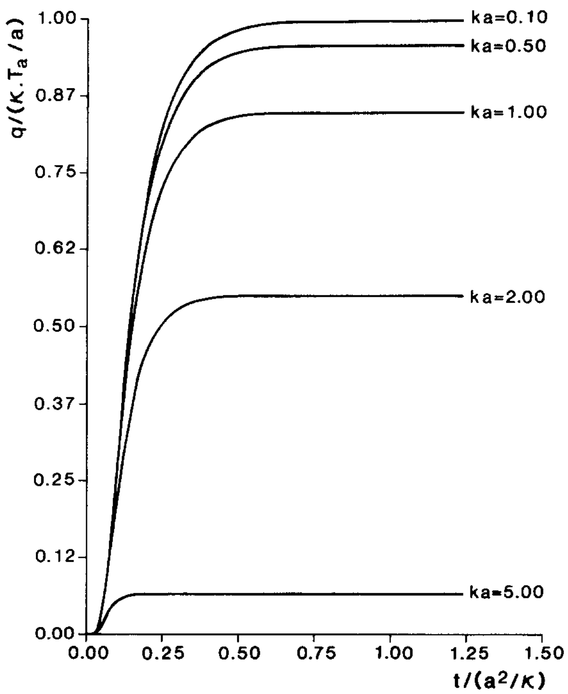

Fig. 3. The surface heat flow at time t after a stepwise change in temperature at the lower boundary for different dimensionless wavenumbers. The heat flow unit is the amplitude of the heat flow perturbation at the lower boundary; all the other conventions are the same as used in Fig. 2.

number and are everywhere positive; the boundary conditions can still be decomposed into their Fourier components and the surface heat flow is computed by superposing the individual components; filtering by the slab will produce a heat flow anomaly that is wider and has a smaller amplitude at the surface than at the base of the lithosphere. Therefore, the source of a narrow and large amplitude thermal anomaly must be shallow.

Another difficulty of the conductive model is that a large amplitude anomaly always requires a large temperature perturbation at the base of the slab (whether the boundary condition is flow or temperature). An order of magnitude of the temperature perturbation is $\Delta q_a \ a/K$: for $K = 2$ Wm^{-1} $°K^{-1}$ and $a = 100$ km, a 40 mW/m^2 anomaly requires $\Delta T_a = 2000°C$; if the source of the thermal perturbation is at Moho depth (40 km), it still requires $\Delta T_a = 800°C$. Both numbers are too high because they imply melting of the lower crust and mantle.

Finally, equilibrium is reached very slowly; for a heat flow perturbation, the time required is of the order of a^2/κ, i.e., 300 Ma for $a = 100$ km or 50 Ma for $z = 40$ km. The time required to reach equilibrium is significantly lower for a temperature perturbation; comparing the first terms in the expressions (4) and (5) suggests that only 1/4 of this time is needed (75 Ma for the lithosphere or 12 Ma for the crust), and this is demonstrated by direct comparison between Figs. 2 and 3. This point was also made by Lachenbruch and Sass (1978). Instantly after the stepwise increase in temperature at the base of the lithosphere, the heat flow becomes infinite; therefore, the amount of heat flowing into the lithosphere initially is much larger after a change in temperature than after a change in the heat flow at the lower boundary. Conductive heating could be an acceptable mechanism only if the source of the thermal perturbation is not deeper than the Moho, if the temperature is suddenly raised, and the termal perturbation is maintained during the whole episode.

Injection of magmas into the lithosphere

Lachenbruch and Sass (1978) have proposed that extension of the lithosphere could be accompanied by the injection of magma from the asthenosphere. These magmas, which are hotter and release latent heat as they solidify, are the source of the thermal anomaly. Lachenbruch and Sass (1978) computed analytical solutions for the thermal perturbation under steady-state conditions. Time-dependent solutions for a similar problem were derived by Mareschal (1983b). The thermal effect of the intruding magmas can be simulated by uniformly distributed heat sources; indeed, when magmas rise into the colder lithosphere, they transport latent and specific heat; if the ascension and freezing time of the magmas is short compared to the heat conduction time, the thermal effect will be identical to that of heat sources with intensity proportional to the amount of heat transported.

The geometry of this problem is shown on Fig. 4. With the same conventions as above and assuming that the magmas intrude only the lower frac-

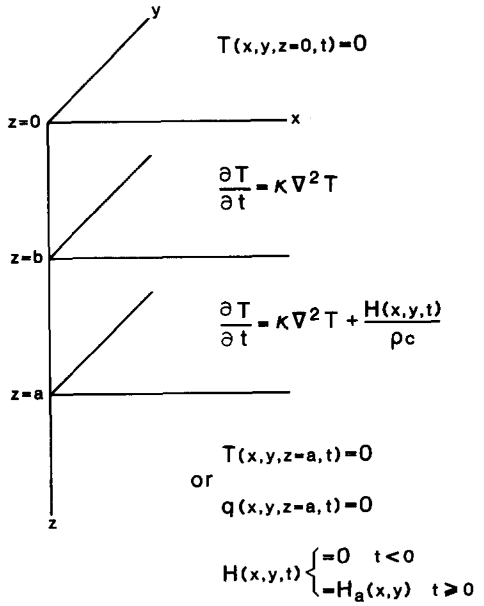

Fig. 4. The geometry, equations, and boundary conditions for the magma injection model.

tion of the lithosphere (the region $b < z < a$); the temperature perturbation is the solution of the heat equations:

$$\frac{\partial T}{\partial t} = \kappa \nabla^2 T \qquad 0 < z < b \tag{6a}$$

$$\frac{\partial T}{\partial t} = \kappa \nabla^2 T + \frac{H}{\rho C} \qquad b < z < a \tag{6b}$$

The surface $z = 0$ is isothermal:

$$T(x,\ y,\ z = 0,\ t) = 0 \tag{7a}$$

and the heat flow or the temperature is constant at z = a

$$T(x,\ y,\ z = a,\ t) = 0 \tag{7b$'$}$$

or

$$q(x,\ y,\ z = a,\ t) = 0 \tag{7b$''$}$$

The conditions at the lower boundary are introduced only for convenience. A stepwise change in heat flow or in temperature must take place at the lower boundary. This boundary condition can easily be included by adding the temperature field for no sources and inhomogeneous boundary conditions given in eqns. (4) and (5) to the solution with sources and the corresponding homogeneous boundary condition.

Initially, there is no temperature perturbation:

$$T(x,\ y,\ z,\ t = 0) = 0 \tag{8}$$

Again, the source distribution in (6b) can be decomposed into its Fourier components and a solution determined for each wavenumber individually. It can be shown (Mareschal, 1983b) that, for the heat flow boundary condition (7b″), the surface heat flow perturbation following a stepwise increase in heat sources is given by:

$$q(k_x,\ k_y,\ t) = aH(k_x,\ K_y)\left\{\frac{\sinh k(a-b)}{ka \cosh ka} - 8 \exp(-\kappa k^2 t) \times \sum_{n=0}^{\infty}\left\{(-1)^n \exp\left[-(2n+1)^2\pi^2\kappa t/4a^2\right] \times \frac{\sin\left[(2n+1)\pi(a-b)/2a\right]}{\left[(2n+1)^2\pi^2 + 4k^2a^2\right]}\right\}\right\} \tag{9}$$

If the temperature is kept constant at the lower boundary, the heat flow perturbation is given by (Mareschal, 1987):

$$q(k_x,\ k_y,\ t) = aH(k_x,\ k_y)\left(\frac{\cosh k(b-a) - 1}{ka \sinh ka} - 2 \exp(-\kappa k^2 t) \times \sum_{n=1}^{\infty}\left[(-1)^{n+1} \exp(-n^2\pi^2\kappa t/a^2) \times \frac{\cos[n\pi(b-a)/a] - 1}{k^2a^2 + n^2\pi^2}\right]\right) \tag{10}$$

The heat flow perturbations following the intrusion of magmas throughtout the lithosphere are shown on Fig. 4 and 5 for constant flux and temperature conditions, respectively. The time unit is (a^2/κ); the heat flow amplitude is aH (which is the steady state heat flow anomaly that would be observed if the whole lithosphere were uniformly intruded with sources of intensity H). When an isothermal condition is assumed, the amplitude of the surface heat flow is only half the amount of

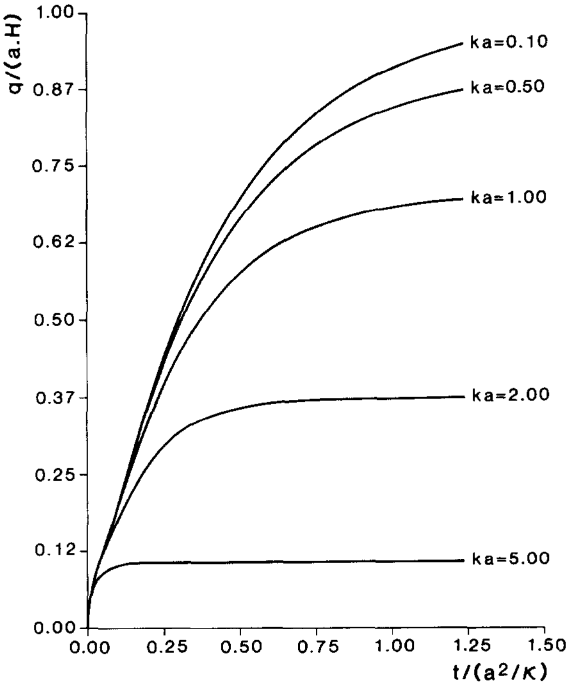

Fig. 5. The heat flow perturbation at time t after the onset of magma injection in the whole lithosphere at constant rate when the flow is constant at the lower boundary. The heat flow amplitude is aH. All the other conventions are the same as in Figs. 2 and 3.

heat produced by the extra sources; this is because half the extra heat flows downward and half flows upward. All the extra heat will flow upward, only when the lower boundary temperature increases and is equal to the perturbation caused by the sources.

When a 100 km thick lithosphere is intruded, the intensity of heat sources required to produce a 40 mW/m^2 anomaly is 0.4 $\mu W/m^3$. The latent heat of the magmas is of the order of 400 kJ/kg and the specific heat of mantle rock is 0.7 kJ kg^{-1} $°K^{-1}$; if the excess temperature of the magma is 300°C, the amount of heat carried will be 600 kJ/kg or 2 GJ/m^3; the rate of intrusion required is therefore $0.2 \cdot 10^{-15}$ s^{-1}, i.e., 0.006 Ma^{-1}. For the Basin and Range province, an extension of 30% during the past 50 Ma is not excessive and is much less than what has been suggested by geophysical and geological data (e.g., Eaton, 1980). The temperature perturbation at the base of the lithosphere in steady state would be about half the temperature change required by the conductive models (1000°C or more, depending on how large a fraction of the lithosphere is invaded by magmas). This is large and implies melting in the lower lithosphere.

It is at first surprising that the time of thermal equilibrium is not much smaller for this mechanism where heat is convected than for the conductive model discussed above. This is demonstrated by a direct comparison of Figs. 5 and 6 with Figs. 2 and 3 as well as by comparing the analytical solutions which contain the same type of exponential functions of the time in the series. The reason is that, initially, the additional heat is absorbed by the surrounding lithosphere and raises its temperature. Only after the lithosphere has been heated will the extra heat flow through the surface. The heat transported in the lithosphere depends only on the extension rate. With the above assumptions on latent heat and magma temperature, the heat flow (in W/m^2) will be of the order of $2 \cdot 10^{14}$ times the extension rate (in s^{-1}).

The choice of the appropriate condition at the lower boundary is important because it changes

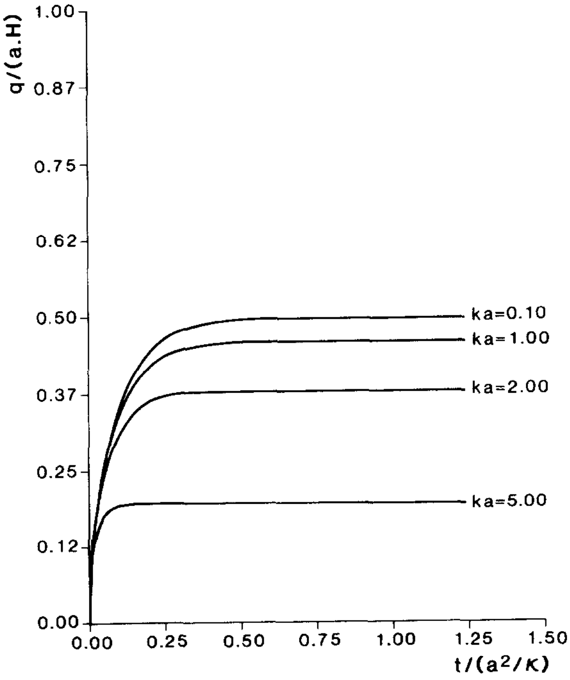

Fig. 6. The heat flow perturbation caused by magma injection when the temperature is constant at the lower boundary. $(a-b)/a = 1.00$. All the conventions are the same as above.

by a factor of 4 the time needed to reach steady state. With the isothermal condition at the lower boundary, half of the additional heat flows upward and half flows downward; and the heat flux vanishes at midplate. In other words, if the whole lithosphere is intruded, an isothermal condition at the lower boundary is equivalent to a flux condition at midplate. None of the conditions brings equilibrium in 50 Ma; the isothermal condition requires 75 Ma if the lithosphere is 100 km thick. The choice of the isothermal condition seems reasonable since the source of the magma at the LAB has a higher temperature, and therefore the rise of magma is accompanied by a jump in temperature rather than a jump in heat flow at the lower boundary.

Regardless of the boundary condition, steady-state conditions will not be reached in less than 75 Ma at best; the time required to reach equilibrium is the same regardless of how large a fraction of the lithosphere is intruded (Mareschal, 1983b). Dike injections during extension would account for the large heat flow anomaly in the Basin and Range province only if either the extension episode had been longer than what is commonly recognized or the extension rate is larger than required by steady-state conditions. Such an extension (100 to 150% during the last 50 Ma) is probably not incompatible with geologic data, but it implies that about $5 \cdot 10^7$ km^3 have been injected as dikes in the lithosphere in the past 50 Ma. A continuous rate of magma injection of the order of 10 km^3/yr is large; it is the rate at which new crust is produced by all the midoceanic ridges. Whether such a rate could have been sustained in the Basin and Range during 50 Ma is debatable.

Lithosphere stretching

If only the lower fraction of the lithosphere in extension is invaded by dikes, mechanical deformation of the upper part must take place to accommodate the extension. This solid-state stretching (elongation and thinning) of the whole or of a fraction of the lithosphere will raise the isotherms and therefore cause a transient increase in heat flow. An upper bound on the change in heat flow can be obtained by assuming that stretching is instantaneous and neglecting the redistribution of radioactive heat sources; the heat flux after stretching, q_f, is related to the flux before stretching q_1:

$$q_f = q_1(1 + r) \tag{11}$$

where r is the ratio of initial to final thickness of the extended lithosphere. It does not depend on the thickness of the stretched slab. However, if stretching takes place at a finite rate, the heat flow anomaly will be smaller because of heat diffusion; the time for the transient to decay depends on the thickness of the slab that is stretched.

The transient heat flow perturbation can be determined by solving the heat equation with a transport term that includes the effect of mechanical stretching:

$$\frac{\partial T}{\partial t} + v \cdot \nabla T = \kappa \nabla^2 T + \frac{H}{\rho C} \tag{12}$$

where H is the source distribution (radioactive sources as well as magma injection effects) and v is a velocity distribution compatible with the mass conservation condition; for instance:

$$\nabla \cdot v = 0 \tag{13}$$

if there are no sources or sinks of matter.

Let us consider a two-dimensional model of the lithosphere, with a coordinate system (x, z) and using the same conventions as above. A velocity field compatible with (13) would be:

$$v_x = -v_0\left(\frac{z}{a}\right) \quad 0 < z < a \tag{14a}$$

$$v_x = v_0\left(\frac{x}{a}\right) \quad 0 < z < a \tag{14b}$$

Continuity of the velocity and mass conservation at $z = a$ can be achieved in many ways; the two models that will be considered are sketched in Fig. 7. The first model (7a) is similar to the underplating model of Lachenbruch and Sass (1978); a sill of new material is accreted at the base of the slab. In the second model (7b), extension continues but mass conservation is achieved by the injection of dikes below the level $z = a$. These two models require different boundary conditions at $z = a$.

If there is no horizontal temperature gradient, the heat equation for the flow defined by (14a)

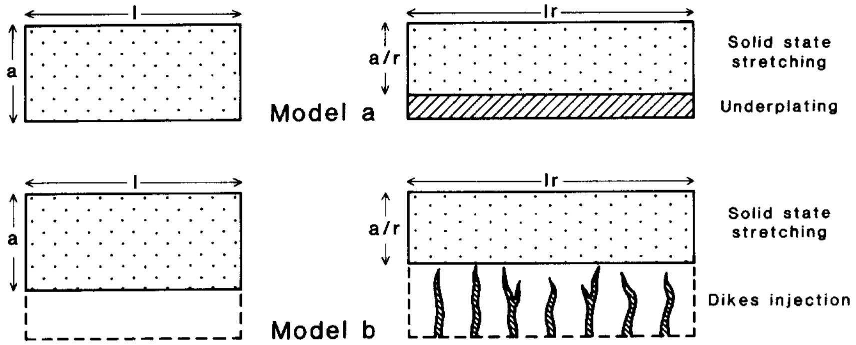

Fig. 7. Two different models of the lithospheric stretching. In model (a), underplating, magma is accreting and solidifying at the base of the lithosphere. In model (b), stretching and injection, extension takes place by solid-state stretching in the upper part of the lithosphere and by dike injection in the lower part.

and (14b) reduces to:

$$\frac{\partial T}{\partial t} - \frac{v_0 z}{a}\frac{\partial T}{\partial z} = \kappa\frac{\partial^2 T}{\partial z^2} + \frac{H}{\rho C} \tag{15}$$

The temperature stays constant at the surface $z = 0$. Initially, the lithosphere is in thermal equilibrium and the surface heat flow is q_0.

$$T(z, t=0) = q_0\frac{z}{K} - \int_0^z dz' \int_0^z dz'' \frac{H(z'', t=0)}{K} \tag{16}$$

Equation (15) was solved numerically with different conditions at the lower boundary corresponding to underplating or dike injection. Figure 8 shows the surface heat flow as a function of time for crustal stretching and dike injection with heat flux constant at the LAB. Three different extension rates have been considered (0.01, 0.015, and 0.02 Ma^{-1}); the latent heat of the injected magma is assumed to be 400 kJ/kg, and the magma is 300°K hotter than the surrounding lithosphere. The heat flow at the lower boundary is assumed to be 20 mW/m^2.

Figure 9 shows the surface heat flow with constant temperature at the LAB. All the other parameters are the same as in Fig. 8. Figures 10

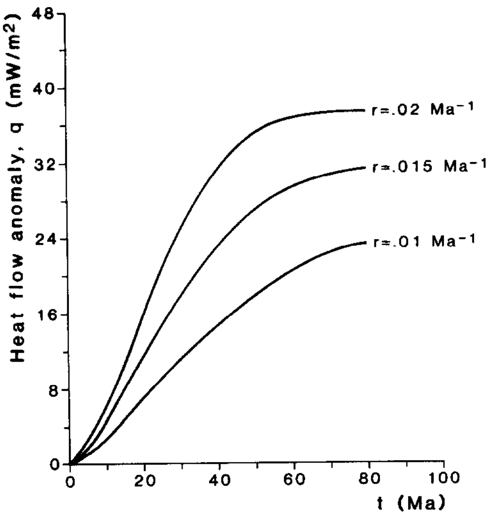

Fig. 8. The surface heat flow as a function of time for the combined stretching and dike injection model, when the heat flux at the LAB is constant. The rate of extension varies from 0.01 to 0.02 Ma^{-1}. The latent heat of intruding magma is 40 kJ/kg, and the magma is 300°C hotter than the surrounding lithosphere. The reduced heat flow (q_m) is 20mW/m^2. Crustol thicknes = 40 km, lithospheric thickness = 100 km.

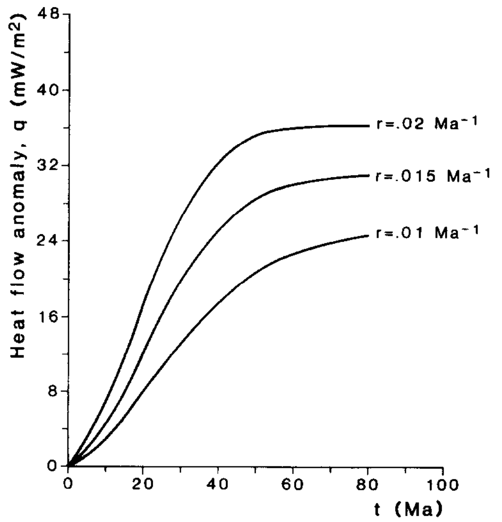

Fig. 9. The surface heat flow as a function of time for the same models when the temperature is constant at the LAB. $q_m = 20$ mW/m^2.

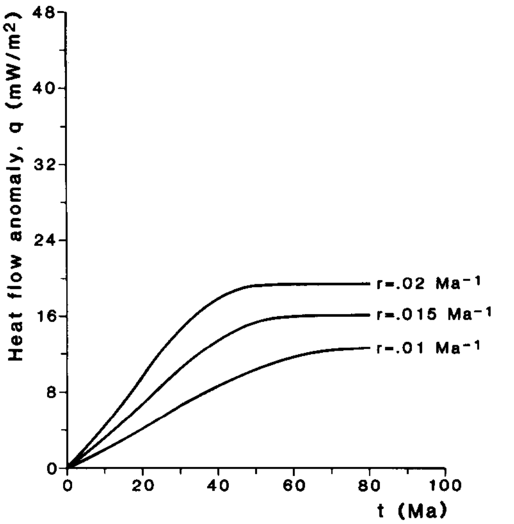

Fig. 10. The surface heat flow for solid-state stretching with constant flux at the LAB. $q_m = 20$ mW/m^2. Initial lithospheric thickness = 100 km.

and 11 show the effect of extension with constant flux and constant temperature boundary conditions at the LAB, respectively. The constant flux condition models the effect of stretching; the constant temperature simulates simple stretching with underplating.

The evolution of the surface heat flux is clearly different for each mechanism. For extension with constant temperature at the lower boundary, the heat flux will increase as long as extension continues while in all the other models steady-state conditions are approached after about 50 to 100 Ma, depending on boundary conditions. Constant flux at the LAB produces a smaller heat flow anomaly. The two models of extension combined with intrusion give rise to approximately the same heat flow anomaly in steady state, but a steady-state regime is established more rapidly when isothermal conditions are assumed at the LAB. Because dike injection brings heat sources in the mantle, the thermal perturbation will be larger than for the underplating model by about 25% (after 30 Ma).

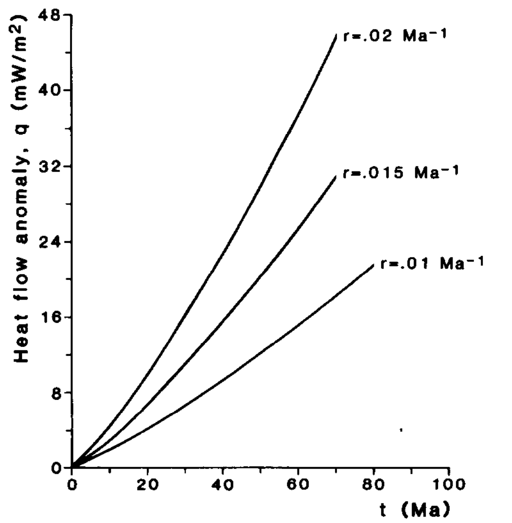

Fig. 11. The surface heat flow for solid-state stretching with constant temperature at the LAB. $q_m = 20$ mW/m^2. Initial lithospheric thickness = 100 km.

Discussion and conclusions

Different mechanisms that could explain the higher heat flow in tectonically active regions have been considered, and the transient behavior of the surface heat flow has been determined.

Because a long time is required to reach equilibrium and the large temperature perturbations imply melting of the crust and mantle, conductive heating from below the lithosphere or the crust is not a feasible mechanism except perhaps for some local anomalies with shallow origin.

Heating of the lithosphere by the injection of magmas in more or less uniformly distributed dikes does meet the requirements of a feasible model. Steady-state conditions are not be established in less than 75 Ma at best; therefore, the rate of extension and of magma injection would be large. The geological data could support extension rates of 0.02 Ma^{-1} (i.e., 100% in the past 50 Ma); but if extension is accompanied by magma injection, the rate of magma intrusion in the lithosphere of the Basin and Range province was at least 10 km^3/yr. This is the rate at which new crust is being generated at all the midoceanic ridges, and it is at least questionable that such a rate could have been sustained over 50 Ma in the Basin and Range province.

Stretching of the entire lithosphere with constant temperature at the LAB is the most efficient mechanism to increase continuously the surface heat flow. However, before quasi steady-state conditions are established (80 Ma), stretching of the crust with dike injection in the mantle causes a heat flow anomaly of the same magnitude, or even slightly larger.

In the Basin and Range, an extension of 100% caused by lithospheric stretching without intrusion would give rise to a heat flow anomaly equal to the initial reduced heat flow. One hundred percent (100%) extension is compatible with geologic data, but it requires the initial crustal thickness to have been double the present crustal thickness (i.e., 60 km), unless new crust was created; in this situation, the stretching model is no longer applicable. The hypothesis, that the crustal thickness in the Basin and Range province was 60 km and that 100% stretching took place in the past 50 Ma, cannot be ruled out altogether. Also, the addition of new crust by some form of underplating should not be ruled out, particularly in view of the subhorizontal Moho reflections obtained by COCORP which suggest that Moho is a recent feature (Hauser et al., 1987). Alternately, the same heat flow anomaly could have been produced by a slightly slower extension (75%) of a thinner crust accompanied by dike injection in the mantle. The one-dimensional calculations presented here do not permit discrimination between these hypotheses. Modeling the lateral variations of the thermal regime could perhaps provide more insight on the mechanism of extension in the Basin and Range province.

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