# The Earth's thermal gradient

Modern theories emphasize the importance of mantle convection and recognize the variation in temperature distribution beneath different regions of the globe.

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Vast amounts of heat are stored in the Earth, often breaking dramatically through the surface at volcanoes and hot springs such as that shown in figure 1. Just how this heat is distributed within the Earth has concerned geophysicists for years. Recently their theories of the temperature distribution have been influenced in several ways by the revolutionary plate-tectonics model of the Earth. This model[1] depicts the relative motions and mutual interactions of several large surface plates, driven by material welling up in some parts of the globe (ridges) and descending in others (trenches), as illustrated in figure 2 (a map of the worldwide plate boundaries is shown in the accompanying article by Carl Kisslinger). Such a model has forced the abandonment of a cherished geophysical boundary condition—spherical symmetry in the temperature profile. Overwhelming evidence has established that temperature distribution beneath the ridges is very different from that at the convergent plate boundaries; the temperature profile in the upper portion of the globe depends very much on whether it is below a zone of spreading plates, a zone of colliding plates or a region of plate interiors.

The plate-tectonics model has also underscored the importance of mantle convection in determining the temperature distribution. Little attention is now given to conductive heat transport within the Earth or to radiative heat transfer in the upper part of the Earth. Although these effects must, of course, be incorporated into any theory of the Earth's thermal structure, they are being displaced by considerations of heat advection by vertical and horizontal movements of mantle material. Geophysicists are also generally agreed that finding the temperature profile does not require cosmological considerations such as the hypothesis of changes in the universal gravitational constant or the assumptions of an expanding or contracting planet that have occasionally been advanced.

Of course we must be careful not to be dogmatic about this particular viewpoint because of a lesson learned from the century-old debate over the age of the Earth. Earth scientists, led by Charles Darwin, estimated that the Earth was over 300 million years old on the basis of empirical geological data. On the other hand, physicists, led by Lord Kelvin, claimed that the Earth was only a few million years old on the basis of deductions from classical heat-transfer and radiation theories. With the discovery of radioactivity, new evidence emerged that proved Kelvin's ideas to be wrong. Geophysicists now routinely report granitic ages in excess of $1.8 \times 10^9$ years. Nevertheless, without being dogmatic in the present case, we can summon the support of the modern concept of plate tectonics together with a large number of extraordinary experiments conducted on a grand scale to back our assertion of the importance of mantle convection to the temperature gradient.

## Methods of estimating the temperature

Geophysicists use three general methods to estimate the temperature distribution:

▸ They may deduce the temperature from direct measurements of surface heat flux, radioactivity and other physical properties such as thermal conductivity. Or they may infer the temperature from measured variations in physical properties, such as electrical conductivity, that depend on temperature, pressure, composition and other factors in a known way. For instance, simple

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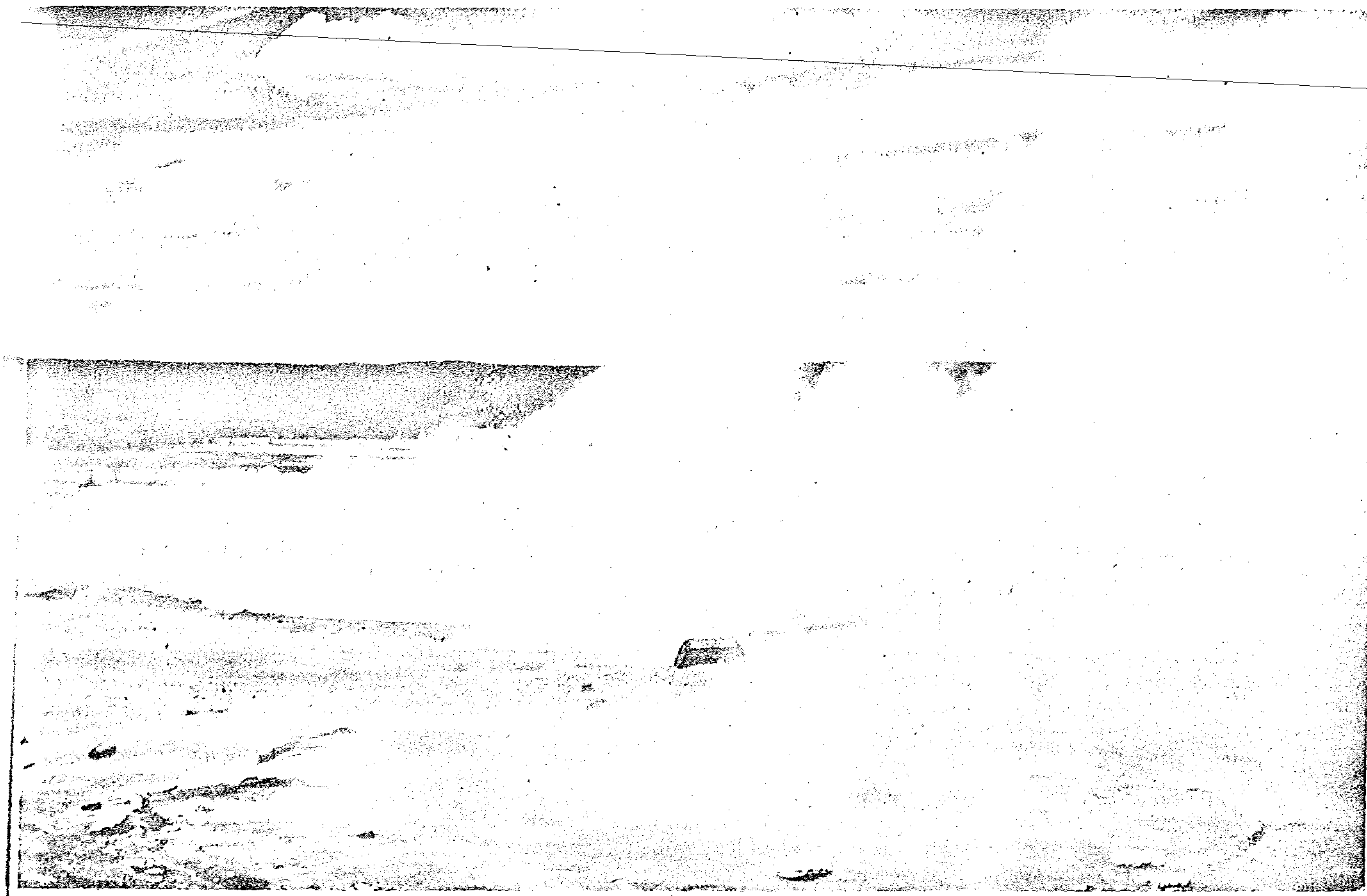


**Hot springs** present a vivid demonstration of the high temperatures present within the Earth, at points where there is a local aberration in the thermal gradient—which itself varies from point to point in different parts of the Earth's surface. This photograph shows Clepsydra Geyser (*clepsydra* is Greek for "water clock") in Yellowstone National Park, Wyoming. Figure 1

lattice-theory concepts can be used to invert measured sound-velocity profiles to temperature profiles. Unfortunately, these latter methods do not provide unique interpretations and are often not very reliable because of the many additional unknowns, such as composition.

▸ They may base their estimates on geochemical and petrological data taken from experiments at high temperatures and high pressures, coupled with inferences from thermodynamics. As one example, assumed petrological models of the Earth can be used to establish the relative contribution of thermal conduction, radiation, and internal radioactive heat sources to the geotherm at various depths.

▸ They may seek analytical and numerical solutions to fluid-dynamical and thermal equations, using special boundary conditions and measured values of physical properties. For example, measured values of heat flow from an ocean ridge (which is producing hot rocks for the new crust and lithosphere) can be compared with the theoretical predictions from mantle-flow models that assume certain properties of the deep rocks.

In all these methods, surface heat-flux measurements are among the most important data. They provide an overview that suggests a certain picture of the temperature distribution below the surface and they also constitute the only experimental boundary condition that all models must satisfy. In the past decade a large amount of heat-flux data taken from widespread locations over the Earth have become available.[2-5] In addition, laboratory data on thermal and radioactive properties are now available for many geological materials. These data, together with knowledge of present and past motions of plates, are leading to a new and, at least quantitatively, consistent understanding of the Earth's thermal state in depths of a few hundred kilometers. Let us start with a global look at the terrestrial heat flow before focusing on the heat flow in the specific regions of continents, oceanic ridges and basins and oceanic trenches.

## Surface heat flux

When some of the most recent heat-flux data[4] are organized according to geographical regions and major tectonic regions on continents and oceans, as in Table 1, a striking feature is the near equality of the average heat flux for continents and oceans. This feature is, at first, rather surprising in view of the much larger total and specific content of radioactive heat-producing elements in the thicker "sialic" (rich in silicon and aluminum) continental crust as compared with the thinner "mafic" (rich in magnesium and iron) oceanic crust. Uranium, thorium and potassium are from 5 to 10 times more abundant in continental crust rocks than in oceanic basalts; thus, heat production per unit volume is roughly an order of magnitude larger for the continental granitic-type rocks. In fact, much of the heat flux at the surface of continents can be accounted for by the total heat produced within the continental crust by radioactive decay. By contrast, radioactive decay in the oceanic crust can account for only a few percent of the oceanic heat flux.

One interpretation of this equality of oceanic and continental heat flux was given by Edward C. Bullard[6] after Roger Revelle and Arthur E. Maxwell's first measurement of oceanic heat flux.[7] Bullard postulated that the heat flux from the suboceanic mantle must exceed that from the mantle beneath the continents, implying deep-seated differences in radioactivity in the mantle underlying these two major topographical units. However, in the light of continental drift and sea-floor spreading these differences would of necessity be confined to the moving

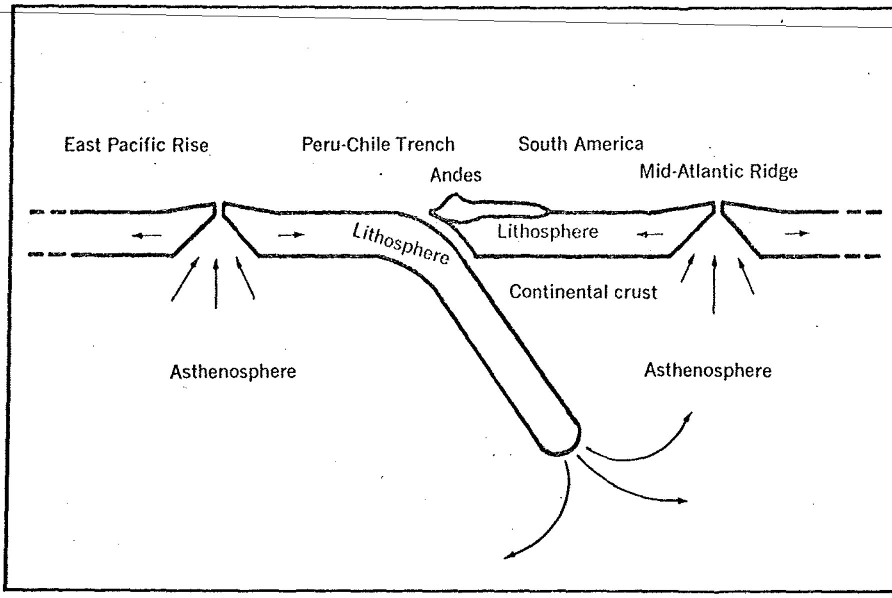



**Plate-tectonics** theory describes the relative motion of large surface plates. Material upwells at the mid-oceanic ridges, forcing the plates to move outward and finally to descend into the trenches. (Redrawn from a figure by F. M. Richter, "Dynamical Models for Sea Floor Spreading," in Revs. Geophys. Space Phys. **11**, 1973.) Figure 2

lithospheric plates. An alternative view of the heat-flow equality calls for a thicker lithosphere beneath continents and an isothermal "asthenosphere," (a region wherein deformation or creep of material occurs more readily than in the relatively rigid overlying lithosphere) at the base of both continental and oceanic lithospheres.[8] Precisely how this comes about is not specified, but presumably mantle convection must play an essential role in maintaining such an isothermal asthenosphere.

Mantle convective motions may also contribute significantly in other ways to the heat-flow equality. For example, Gerald Schubert and Donald L. Turcotte[9] have used a model in which the mass motion of lithospheric plates is balanced by a shallow return flow in the upper mantle to show that viscous dissipation can contribute 10–20% of the observed oceanic heat flux. This production of heat by viscous dissipation was remarkably insensitive to variations in the plate velocity and viscosity function. The flow is apparently self-adjusting, with heating by viscous dissipation providing a self-lubricating mechanism. Other calculations have confirmed this self-adjusting or lubricating nature of the flow.[10]

Tabel 1 also shows that, within continents and ocean basins, heat flow correlates with geologic provinces and major tectonic boundaries. Heat flow is low and rather uniform in Precambrian-shield areas, which have been geologically stable since the Precambrian, while more recent orogenic (that is, "mountain forming") areas have both higher and lower, and more scattered, values of heat flux.[2-4] Over ocean basins heat flux is moderate and uniform as compared with the higher and more scattered values of heat flux over oceanic ridges, and the relatively lower values at oceanic trenches. Let us now take a closer look at observations and theories of heat flow over specific regions of the globe.

### Continental heat flow

Variations in heat flow over continental areas are typified by those for the United States. Heat-flow values in the eastern US[11] are rather uniform, with high values occurring at locations of plutonic rocks of high radioactivity. Heat flow in the western US[12] is characterized by regions of high (Basin and Range) and low (Sierra Nevada) values, with significant variations being attributable to variations in heat flow from the mantle.

A major step foward in understanding the significance of these variations in continental heat flow was the realization that, to a large extent, such variations could be attributed to regional differences in upper crustal radioactivity.[13] It has been found that surface heat flux $q_s$ is linearly related to the concentration $Q_s$ of radioactivity in surface plutonic rocks:[14,15]

$$q_s = q_m + Q_s\delta \tag{1}$$

where $\delta$ is a measure of the depth over which upper crustal radioactive heat sources are important and $q_m$ is a measure of the heat flux into the crust from the underlying mantle. In the eastern US, $q_m$ = 0.8 microcal/cm$^2$ sec and $\delta$ = 7.5 km; in the Basin and Range, $q_m$ = 1.4 microcal/cm$^2$ sec and $\delta$ = 9.4 km, and in the Sierra Nevada, $q_m$ = 0.4 microcal/cm$^2$ sec and $\delta$ = 10 km.[15] Presumably, from the equation above, with appropriate values of $\delta$ and $Q_s$, the contribution of crustal radioactivity to the surface heat flux can be removed to reveal the heat flux $q_m$ emerging from the mantle. The high values of heat flow in the Basin and Range and the low values in the Sierra Nevada can be attributed to variations in heat flux from the mantle. Several plate-tectonic models have been suggested to explain qualitatively the mantle heat-flow variations beneath the western US.[11]

This method of discovering the mantle heat flow from beneath the continents has involved only a knowledge of the integrated effect of crustal radioactivity. To extrapolate the Earth's temperature to depths within the continental crust we require information on the depth dependence of crustal radioactivity. However, the equation relating surface values of heat flux and radioactivity does not uniquely determine the depth distribution of crustal radioactive heat sources. Nevertheless, the hypothesis of a distribution that decreases exponentially with depth appears reasonable from considerations of geochemistry,[16] of thermodynamics[17] and from direct empirical evidence.[18] Thus we assume that the concentration $Q$ of radioactive heat sources in the crust decreases with depth $d$ as given by

$$Q = Q_s e^{-d/\delta} \tag{2}$$

Integration of this depth-dependent radioactivity concentration leads to the empirically determined relationship between surface heat flux and surface radioactivity noted above. The two equations are thus mutually consistent.[19]

The temperature $T$ in a layer of radioactivity that decreases exponentially with depth is given by

$$T - T_s = \frac{\delta^2 Q_s}{k}(1 - e^{-d/\delta}) + d\left(\frac{q_s - \delta Q_s}{k}\right) \tag{3}$$

where $T_s$ is the surface temperature and $k$ is the thermal conductivity. Thus a knowledge of surface heat flux, surface radioactivity, the scale height of radioactivity distribution and the thermal conductivity suffices to determine the temperature as a function of depth within the constraints of the assumed depth distribution of radioactives. Given the empirically determined values of $q_m$ and $\delta$ mentioned earlier, and using $k = 6 \times 10^{-3}$ cal/cm sec deg C, Arthur H. Lachenbruch[19] has calculated crustal temperature distributions for a number of possible values of $Q_s$. Calculated temperatures at the base of the crust in the Basin and Range province are sufficiently high to allow for some partial melting at this depth. Although there is still some uncertainty in the quantitative character of the crustal geotherm for each continental region, there can be no doubt about the qualitative validity of the temperature distribution for a particular province. Furthermore, comparisons of geotherms between provinces can be made with some confidence. Continental heat-flow values are also influenced by the time and nature of the most recent orogenic event. Heat flows for tectonic provinces are correlated with the age of such provinces, the older ones having the lower heat flows.[20]

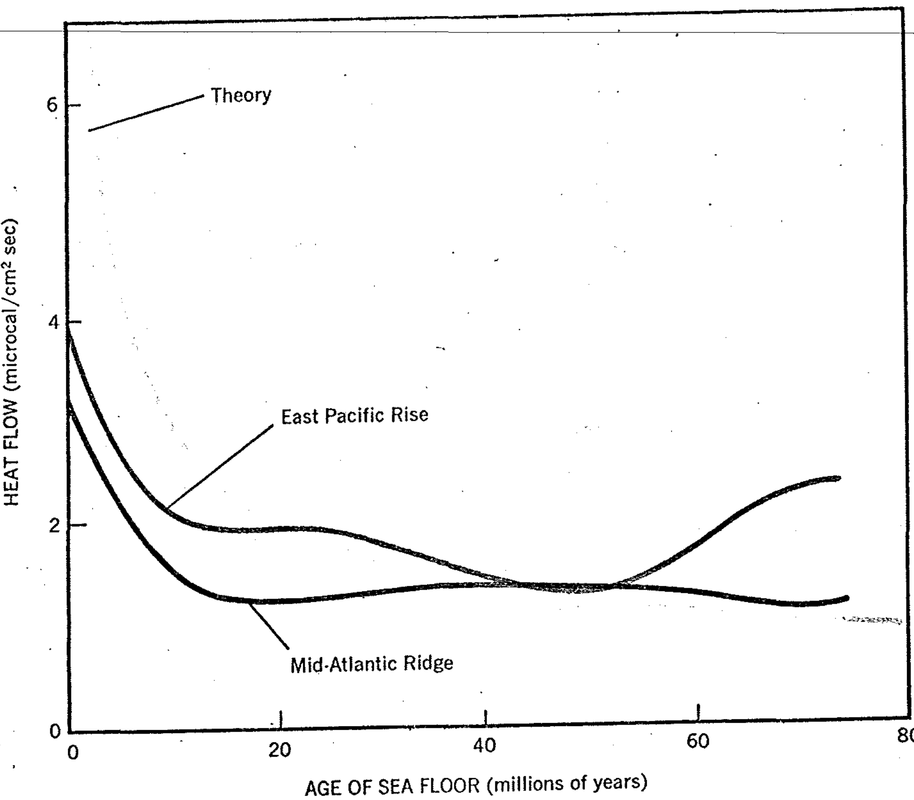
### Oceanic heat flows

Although there is considerable scatter in the measurements, the heat-flow profiles for mid-ocean ridges are qualitatively similar. Typically, they exhibit reasonably high values near the ridge axis and tend to decrease to lower values with increasing sea-floor age or distance from the ridge axis, possibly approaching a rather constant value far from the ridge. These features are seen in general discussions and summaries of heat-flow data for the floor of the world oceans[21] and in some recent reports of newly acquired data.[22]

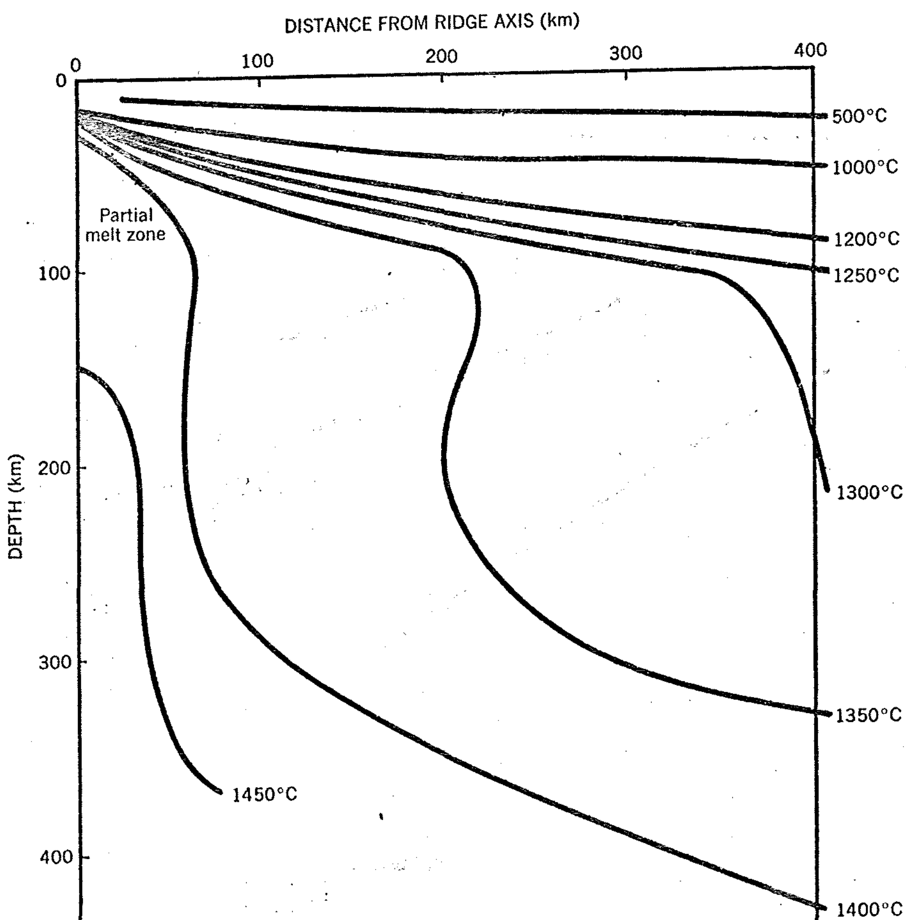
A relatively simple plate-tectonic model can qualitatively explain the variation in heat flow with distance from a mid-ocean ridge. This model of a spreading ridge, discussed by Dan P. McKenzie,[23] consists of a lithospheric slab of constant thickness moving with a constant velocity. The temperatures at the bottom of the slab, that is, deep under the surface, and at the ridge axis are held fixed at the same constant value while the temperature at the top of the slab is also held fixed. The variation of heat flow with distance from the ridge axis is thus explained simply by the cooling of the spreading lithosphere. Additional calculations with this cooling-slab model of spreading at a mid-ocean ridge have been reported.[4,24] Mark G. Langseth and his colleagues[25] had earlier presented a numerical analysis of a similar model for the thermal state at oceanic spreading centers.

An equally simple and perhaps more realistic model of the thermal environment in the vicinity of mid-ocean ridges views the spreading oceanic lithosphere as a cooling thermal boundary layer in a global system of mantle convection (see a summary of this point of view by Turcotte and E. Ronald Oxburgh[26] and a number of papers written by these authors developing this boundary-layer model of mantle convection.[27]) The cooling of the thermal boundary layer is governed by the simple differential equation

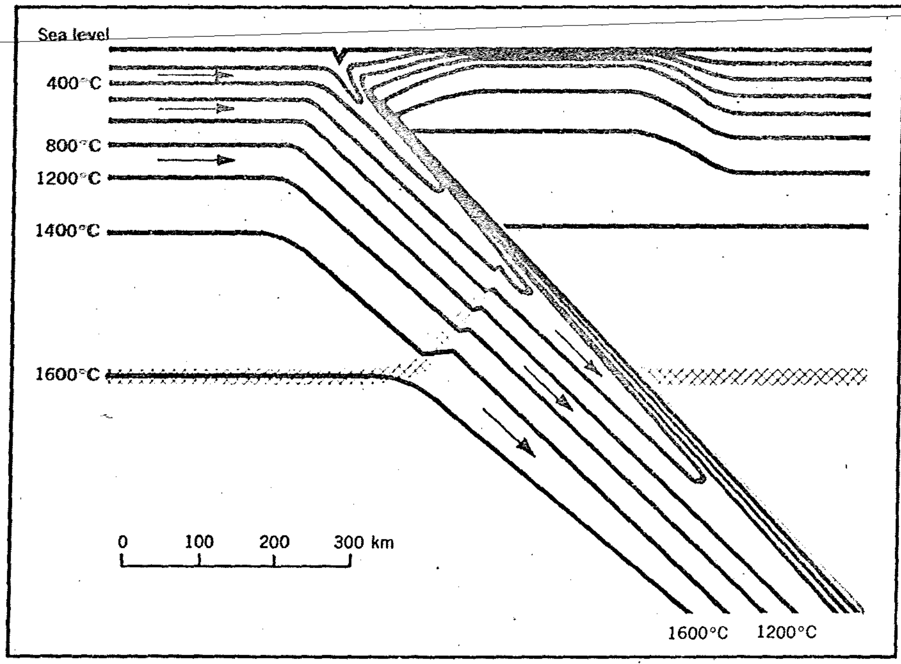
$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \qquad (4)$$




**Heat-flow data** for mid-Atlantic ridge (black) and east Pacific rise (grey) agree qualitatively with theoretical curve (color) which is based on a model of the spreading lithosphere as a cooling thermal boundary layer. (From D. L. Turcotte, E. R. Oxburgh, ref. 26.) Figure 3




**Dynamical and thermal state** near a mid-ocean ridge are depicted by streamlines (color) and isotherms (black). Darker areas indicate higher temperatures. (From D. L. Turcotte, E. R. Oxburgh, reference 26.) Figure 4




**At oceanic trenches,** the cold descending plate lowers the iostherms near the slip zone (colored line) and elevates the olivine-spinel phase boundary (cross-hatched line). (From D. L. Turcotte, G. Schubert, reference 37.) Figure 5

where $T$ is the temperature, $z$ is the depth and $\kappa$ is the thermal diffusivity. At $z = 0$ the temperature must equal $T_0$, the temperature of the ocean floor, and as $z \to \infty$, $T \to T_m$, the isothermal core temperature of the convective cell. $T = T_m$ at time $t = 0$, that is, along the axis of the spreading ridge. The solution[26] is

$$\frac{T - T_0}{T_m - T_0} = \mathrm{erf}\left\{\frac{z}{2}\left(\frac{1}{\kappa t}\right)^{1/2}\right\} \qquad (5)$$

which, by virtue of Fourier's law of heat conduction,

$$q_s = k\left(\frac{\partial T}{\partial z}\right)_{z=0}$$

gives

$$q_s = \frac{k(T_m - T_0)}{\sqrt{\pi \kappa t}} \qquad (6)$$

This result, evaluated for $T_0 = 0$ deg C, $T_m = 1200$ deg C (based on materi-al properties and magma te[...] tures), $k = 10^{-2}$ cal/cm sec deg, $\kappa = 10^{-2}$ cm$^2$/sec, is plotted in t[...] together with heat-flow data fr[...] Mid-Atlantic ridge and East P[...] rise. Theory and observations a[...] qualitative agreement, except pa[...] very close to the ridge axis, whe[...] ficulties encountered in making [...] flux measurements, and the poss[...] that substantial amounts of hea[...] lost directly to the ocean throug[...] drothermal circulation, may suff[...] explain some of the discrepancy. Illustrative streamlines and isoth[...] depicting the dynamical and ther[...] state at a mid-ocean ridge[26] are sh[...] in figure 4.

The model of a spreading and c[...] ing lithosphere is consistent with [...] topography of oceanic ridges, whic[...] characterized by an elevated region [...] the ridge axis and a decreasing el[...] tion with increasing distance from [...] axis. This topography can be lar[...] understood in terms of the ther[...] contraction of the spreading litho[...] ere as it cools and moves away fr[...] the ridge.[24,25,29] John G. Sclater a[...] others[30] have shown that the empir[...] data on ridge elevation and the age [...] the sea floor for the Pacific, Indian a[...] Atlantic oceans support this conc[...] Surface-pressure variations caused [...] mantle convective motions can a[...] contribute to mid-ocean ridge topog[...] phy.[26] Finally, we note that the si[...] ple model of lithospheric spreading a[...] cooling from an ocean ridge has be[...] used to construct petrologic models [...] the lithosphere.[31]

## Table I. Summary of Terrestrial Heat Flux Measurements

| Tectonic and geographical regions | Number of data | Mean heat flux (microcal/ cm²sec) | Standard deviation from the mean (microcal/ cm²sec) |
|---|---|---|---|
| Global* | 3127 (673) | 1.63 (1.47) | 1.07 (0.74) |
| Continents* | 597 (95) | 1.45 (1.46) | 0.57 (0.46) |
| Precambrian shields | 214 | 0.98 | 0.24 |
| Post-Precambrian non-orogenic areas | 96 | 1.49 | 0.41 |
| Palaeozoic orogenic areas | 88 | 1.43 | 0.40 |
| Mesozoic–Cenozic orogenic areas | 159 | 1.76 | 0.58 |
| Oceans* | 2530 (591) | 1.67 (1.47) | 1.15 (0.78) |
| Atlantic Ocean* | 436 (126) | 1.47 (1.34) | 1.14 (0.57) |
| Indian Ocean* | 358 (108) | 1.36 (1.32) | 0.95 (0.52) |
| Pacific Ocean* | 1308 (310) | 1.70 (1.50) | 1.24 (0.84) |
| Ocean basins | 683 | 1.27 | 0.53 |
| Mid-oceanic ridges | 1065 | 1.90 | 1.48 |
| Oceanic trenches | 78 | 1.16 | 0.70 |
| Continental margins | 642 | 1.80 | 0.93 |

**Means and standard deviations are based on arithmetic averages of individual data. Numbers in parentheses, however, are based on averages over equal-area elements measuring 5 degrees × 5 degrees at the Equator, to account for uneven geographical distribution of data for these large areas.*

*Data from W.H.K. Lee, reference 4.*

### Oceanic trenches

Generally speaking, the heat-fl[...] distribution near oceanic trenches [...] consistent with the concept from plat[...] tectonics of relatively cold surfac[...] plates or lithosphere descending in[...] the mantle at the trenches.

On the ocean sides of trenches hea[...] flow is resonably uniform with an ave[...] age value of about 1.3 microcal/cm[...] sec.[4] Directly behind trenches th[...] heat flow generally falls below 1 micro[...] cal/cm$^2$ sec, while farther from th[...] trenches, over inland or marginal sea[...] it rises to values of about 2 microcal[...] cm$^2$ sec.[32] The low heat flow directly behind the trench may be understood in terms of a depression of the iso[...] therms resulting from the presence of the colder descending slab. Farther from the trench, the high surface hea[...] flow and volcanism may be attributed to frictional heating on the slip zone between the descending plate and the overriding mantle[33,34] or to intrusion of oceanic crust caused by tensional stresses.[35]

Among the calculations of tempera[...] ture distributions in the descending lithosphere[36] is a recent and inter[...]

esting proposal by Turcotte and Schubert.[34] They suggest that the linear chains of active volcanoes usually associated with oceanic trenches lie above the point on the slip zone between the descending lithosphere and the overriding mantle where partial melting of the oceanic crust takes place. This idea, coupled with an assumption regarding the frictional-stress distribution along the slip zone (such as one of constant stress), leads to the following simple result for the increase of temperature $T_{sz}$ along the slip zone due to frictional heating[34]

$$T_{sz} = 2\left(\frac{\kappa x}{\pi u \cos\phi}\right)^{-1/2} \frac{u\tau_0}{k} \qquad (7)$$

where $x$ is distance along the slip zone, $\kappa$ is the thermal diffusivity, $\tau_0$ is the assumed constant frictional shear stress on the slip zone, $u$ is the velocity of the surface plate, $k$ is the thermal conductivity and $\varphi$ is the angle between the surface plate velocity and the normal to the trench axis. The above equation applies only up to the point on the slip zone where partial melting begins at the temperature $T_{sz} = T_m$ (the solidus temperature of basalt) at depth $d_v$. Rearranging the above equation, we find that the constant shear stress on the slip zone is given by

$$\tau_0 = \frac{kT_m}{2}\left(\frac{\pi \sin\theta \cos\phi}{u\kappa d_v}\right)^{1/2} \qquad (8)$$

where $\theta$ is the angle the subducting (that is, descending) plate makes with the horizontal. From the solidus temperature, the velocity and geometry of the descending plate, the thermal properties of the plate and the depth to the plate beneath the volcanoes (from seismic data defining the Benioff zone), the shear stress on the slip zone may be determined.

The temperatures inside the descending lithosphere have been computed by extending the analytic model discussed above with numerical calculations.[34] The temperature on the slip zone at depths greater than that which corresponds to the onset of partial melting is assumed to be given by the basalt solidus temperature. Calculations included the effects of adiabatic heating and heating due to the olivine–spinel phase transformation. This latter is a solid–solid transformation from olivine material, that is, magnesium or iron silicate, with varying percentages of iron or magnesium, to spinel material, which is chemically the same but more densely structured. Figure 5 shows an example of isotherms in the descending plate and surrounding mantle.[37] The elevation of the olivine–spinel phase boundary within the descending plate is also shown.

## Geothermometry

A promising new technique to infer the Earth's temperature gradient is being developed by a branch of geochemistry called experimental petrology. Rocks that arise from the deep parts of the Earth often are quenched in their ascent, so that the mineral distribution corresponding to some depth is fixed. Certain minerals are in equilibrium along a given locus of pressure $P$ and temperature $T$ and it sometimes happens that different pairs of minerals, in the same rock, have intersecting loci of $P$–$T$ equilibrium conditions so that the temperature and depth of origin of the rock can be inferred. A sequence of these rocks can be used to infer the temperature–depth profile. This technique appears to be most promising for rocks associated with natural diamonds, because these rocks are ejected cold from great depths in the Earth (in excess of 150 km).[38]

Only a few years ago, it was thought that if the initial conditions of formation of the Earth could be defined and the disposition of radioactive materials in the Earth described, then the past and present temperature distributions of the Earth could be computed. It was thought that this goal would be realizable with numerical techniques and, once it was accomplished, decisions could be made about the chemical-composition distribution by comparing the theoretically determined temperature with measured physical properties. Now it is appreciated, as a result of the abandonment of the spherical-symmetry boundary condition and the importance of advective transport of energy by mantle motions, that this goal is a long way off.

The determination of the temperature distribution of the upper portion of the globe depends, it is now believed, upon the solution of very complicated nonlinear fluid dynamical and heat-transfer equations for the highly viscous mantle. The constitutive parameters in the equations are dependent upon temperature and pressure, and the Newtonian stress–rate-of-strain relationship may not be a valid approximation. The picture is further complicated by transport of material through solid–solid phase boundaries and processes such as partial melting and diffusion. In some regions there is great heterogeneity, and the driving force for plate motions may involve the gradual consumption of heterogeneous materials into a deeper homogeneous mantle.

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