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Some like it hot: The temperature structure of the Earth

Thomas S. Duffy and Russell J. Hemley

Geophysical Laboratory and Center for High-Pressure Research, Carnegie Institution of Washington Washington, D.C.

Introduction

The Earth's temperature distribution is intimately connected to problems of the structure, composition, dynamic state, and evolution of the planet. The high temperatures in the interior provide the driving force for the Earth's convective engine and are ultimately responsible for the vigorous geological activity on the surface of the planet. The past quadrennium has witnessed major advances in our understanding of the thermal structure of the Earth. In this short review of primarily U. S. work, only a few highlights of recent advances can be summarized. While considerable progress has been made, a number of aspects of the Earth's temperature structure remain highly controversial.

Temperature of the Core

Constraints on temperatures in the deep interior can be obtained from measurements of the pressure (P) - temperature (T) slopes of the phase transitions responsible for the major seismic discontinuities. For example, the inner core-outer core boundary (IOB) at 5150 km depth (329 gigapascals (GPa)) separates the liquid outer core from the solid inner core. Measurements of the melting curve of iron, when extrapolated to IOB pressures, establish an upper bound to the Earth's temperature at this depth.

A number of studies have reported measurements of the melting curves of iron and iron alloys in the laser-heated diamond anvil cell [*Williams et al.*, 1991; *Knittle and Jeanloz*, 1991; *Boehler* 1992, 1993; *Saxena et al.*, 1994]. For pure iron, the melting curve measured by *Williams et al.* [1991] to 105 GPa is in strong disagreement with those of *Boehler* [1992, 1993] to 200 GPa and *Saxena et al.* [1994] to 150 GPa, with the melting temperature of the former being ~1100 K higher at 100 GPa. By extrapolating his measurements to core pressures, *Boehler* [1993] estimated the temperature of the Earth's IOB is 4850 ± 200 K and the temperature at the core-mantle boundary (CMB) at 2891 km depth (135 GPa) is about 4000 K. Recent shock compression experiments to 340 GPa [*Yoo et al.*, 1993], however, yield a melting temperature of iron at the IOB about 2000 K higher (6830 ± 500 K), and in general agreement with earlier shock melting experiments as well as static measurements of *Williams et al.* [1991].

An additional problem that has not yet been explored in detail is the effect on melting temperature of the wide variety of possible alloying components in the core (e. g., H, O, C, Si, S, and Ni). There has been sharp disagreement over the melting behavior of FeO. *Knittle and Jeanloz* [1991] measured the melting curve of FeO to 102 GPa, and inferred that the melting temperature of FeO exceeds that of Fe by 1000-2000 K at the CMB pressure. Assuming oxygen is a major alloying component in the outer core, a CMB temperature of 4800 ± 500 K was inferred from this data. *Boehler* [1992, 1993] and *Shen et al.* [1993] obtained significantly lower melting temperatures (by ~800 K at 50 GPa) for FeO than *Knittle and Jeanloz* [1991], and *Boehler* [1993] found no measurable difference in the melting temperature of Fe and Fe-FeO alloys above 60 GPa. *Boehler's* [1992] measurements of the melting curves of FeS and FeS_2 to 50 GPa show strong melting point depression relative to Fe. Shock temperature measurements on an Fe-19%Cr-9%Ni alloy reveal that this material melts at 5800 ± 300 K at 250 GPa, providing constraints on the effect of transition metal alloys on iron melting temperatures [*Gallagher et al.*, 1994].

Using dislocation theory, *Poirier and Shankland* [1993] calculated a melting temperature of 6160 ± 250 K for pure iron at 330 GPa, and estimated that the temperature of the IOB is 5160-5660 K. In a review of experimental data that includes a possible new high-temperature phase of iron above 200 GPa, *Anderson* [1993] concluded that an upper bound to the melting temperature of iron at the IOB is 6500 K, and taking account of plausible depression due to light elements, an upper bound for the central temperature of the Earth is 5700 K. Further discussion of the iron phase diagram can be found in papers collected in *Schmidt et al.* [1994].

Temperature Structure of the Mantle

Melting Curves of Lower Mantle Phases

As silicate perovskite is widely believed to be the dominant phase of the Earth's lower mantle, the perovskite melting curve represents an upper bound to the temperature of this region, and places important constraints on the thermodynamic and rheological properties of the lower mantle. Again, experimental results reported over the last four years disagree. *Sweeney and Heinz* [1993] reported spatially averaged melt temperatures for $(Mg_{0.86}, Fe_{0.14})SiO_3$ perovskite between 30 and

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94 GPa in a Nd:YAG laser-heated diamond cell. Their results support earlier studies which concluded that the melting temperature of perovskite is around 3000 K at 60 GPa, and the extrapolated melting temperature at the CMB pressure is 4500 ± 500 K.

In contrast, *Zerr and Boehler* [1993] measured a melting temperature of 5000 ± 200 K in their CO_2 laser heating experiments at 62.5 GPa. Extrapolation to the CMB pressure yields a melting temperature between 7000 and 8500 K. A number of discussions of the relative merits of these experiments have been reported [*Heinz et al.,* 1991; *Heinz et al.,* 1994; *Boehler and Zerr,* 1994]. The potential implications of a high melting temperature for perovskite on mantle dynamics have just begun to be addressed [*van Keken et al.,* 1994]. Extrapolation of recently measured melting curves (to 32 GPa) of MgO and (Mg,Fe)O indicates these materials melt at 3500-5000 K at lower mantle pressures [*Zerr and Boehler,* 1994], but molecular dynamics calculations of MgO melting temperatures are higher by several thousand Kelvin [*Cohen and Gong,* 1994]. The effect of minor components, including volatiles, on the melting behavior of lower mantle materials has not yet been studied.

Temperature Structure of the Lower Mantle

The most fundamental question for evaluating the thermal regime of the lower mantle is the degree to which the upper and lower mantle are chemically homogeneous. If the mantle is compositionally layered, a thermal boundary layer must exist at or near the 660-km discontinuity that will exert a profound influence on temperatures in the lower mantle. Due to the overriding importance of this question, a large number of studies have been devoted to this issue, of which only a small sampling can be incorporated into this review.

On the basis of thermoelasticity data for magnesiowüstite and silicate perovskite, *Stixrude et al.* [1992] and *Hemley et al.* [1992] concluded that an upper mantle composition does not completely match seismic data for the lower mantle, and specficially, there could be an enrichment of silicon in the lower mantle (see also *Zhao and Anderson* [1994]). The tradeoff between temperature and iron content has also been explored by these authors who found that iron enrichment of the lower mantle is required for the higher temperatures that would be expected in the deep interior of a chemically stratified planet. On the other hand, *Wang et al.* [1994] concluded from an analysis of lower pressure equation of state data on silicate perovskite that a uniform mantle composition is consistent with seismic data. A thermodynamic analysis of data on the thermal expansivity of silicate perovskite has been carried out by *Anderson and Masuda* [1994]. The degree to which compositional differences can develop as a result of the coupling of a chemical change and a phase transition in a multivariate system has been examined by *Bina and Kumazawa* [1993].

This question of compositional stratification has also been approached using numerical simulations of mantle flow. *Glatzmaier and Schubert* [1993] used a three-dimensional, spherical convection model to investigate how the style of mantle convection is influenced by layering. They concluded that whole mantle convection models better reproduce geophysical observations. *Morgan and Shearer* [1993] combined seismic velocity heterogeneities, discontinuity topography, and radial viscosity models to compute radial mantle flow distributions that showed no reduction across the 660-km discontinuity, consistent with whole mantle convection.

Numerical simulations of convection have raised the possibility of a new solution to the above controversy: namely, that the mantle is partially stratified, with the degree of stratification varying in space and time. These results stem from the development of numerical mantle models that include the effects of phase boundaries [*Machetel and Weber,* 1991]. Two-dimensional and axisymmetric calculations showed that the presence of the endothermic 660-km boundary promotes layered or intermittently layered convection [*Zhao et al.,* 1992; *Weinstein,* 1992; *Peltier and Solheim,* 1992; *Solheim and Peltier,* 1994]. The sensitivity of these results to variation of a range of parameters was investigated by *Ita and King* [1994]. Three-dimensional calculations also show complex, time-dependent behavior, including avalanching of cold material across the boundary layer [*Honda et al.,* 1993; *Tackley et al.,* 1994]. When the Clapeyron (P-T) slope of the phase transition is included, downwellings are found to be impeded at the boundary while rising plumes are unaffected [*Liu,* 1994].

Recent comparisons between convective flow calculations and seismic tomography are suggestive of a convective style that is dominantly whole mantle but allows for localized layering [*Jordan et al.,* 1993; *Woodward et al.,* 1994]. *Weinstein* [1993] has argued that episodic flow across the 660-km discontinuity could produce temperatures in the transition region that are ~250 K cooler than the surrounding mantle due to the temporary stagnation of cold subducting material in this region. Additional features that need to be incorporated into numerical models before more definitive conclusions can be drawn include temperature-dependent viscosity, rigid plates, and three-dimensional calculations at higher (more Earth-like) Rayleigh numbers.

Lateral Temperature Variations

Significant progress has been made in recent years in mapping lateral heterogeneities in the seismic velocity structure of the mantle [*Woodward and Masters,* 1991; *Pulliam et al.,* 1993; *Forte et al.* 1993; *Su et al.* 1994; *Vasco et al.,* 1994]. The velocity variations may arise from a combination of temperature, composition, and phase, with temperature believed to play the dominant role. Determining the temperature variations associated with the velocity heterogeneities places constraints on the temperature differences driving mantle convection and enables more direct comparison with dynamic flow models and geophysical observables such as the geoid, heat flow, and plate motions. At ambient pressure, the scaling coefficients between between seismic velocity and temperature have now been measured in a variety of mantle minerals [*Anderson et al.,* 1992;

Isaak, 1992]. Sound velocity measurements at high P and T under shock compression have shown that the compressional velocity-temperature scaling relation at deep mantle pressures (~100 GPa) is ~5 times smaller than ambient pressure values [*Duffy and Ahrens*, 1992, 1994]. From this work, it was estimated that long-wavelength velocity heterogeneities in the deep lower mantle correspond to root-mean-square thermal anomalies of ~150 K. The potentially important role of anelasticity in velocity-temperature scaling relations has been discussed by *Karato* [1993].

In the most detailed studies relating thermal anomalies to seismic tomography results to date, *Yuen et al.* [1993] and *Čadek et al.* [1994] estimated that the large-scale anomalies found in the lower mantle have outer temperatures of 400 K above the surrounding mantle and thermal anomalies in excess of 1000 K near the plume center. Extremely cold anomalies, which could be related to avalanche events from the upper mantle, are also found in the lower mantle. The possible role of radiative heat transfer in the formation of large-scale lower mantle anomalies has been examined [*Matyska et al.*, 1994]. In studies of diffracted and reflected waves in the D″ layer near the base of the mantle, *Wysession et al.* [1993, 1994] argue that thermal anomalies as large as 400-1000 K might explain seismic velocity perturbations near the CMB. Thus, in the deep lower mantle, it is becoming increasingly apparent that large thermal anomalies may exist, despite comparatively low levels of seismic heterogeneity in much of the region.

Temperature Structure of the Upper Mantle

Advances in understanding the thermal structure of the upper mantle have been led by seismological studies that, for the first time, have produced detailed maps of the global and regional variation of the depths to the 660- and 410-km seismic discontinuities. In conjunction with the Clapeyron slopes of the relevant phase transitions (see *Bina and Helffrich* [1994] for a recent analysis of the experimental phase equilibria data), the temperature differences required to produce such topography can be deduced. *Revenaugh and Jordan* [1991] used shear wave reverberation data to measure topography variations on the 410- and 660-km discontinuities and inferred that the lateral temperature variations implied by this topography are ±200 K. Short-period array data were also used to determine that the 660-km discontinuity is depressed by 20-30 km and the 410-km discontinuity is elevated by ~15 km beneath subduction zones, which together imply 300-400 K average temperature difference between subduction zones and normal mantle [*Vidale and Benz*, 1992]. In a high-resolution study of the Izu-Bonin subduction zone, the 660-km discontinuity was found to be depressed by 60 km, suggesting a thermal anomaly of 1000 K [*Wicks and Richards*, 1993]. A global study utilizing underside shear wave reflections from the 660-km discontinuity found regional variations of ±30 km with depressions correlated with subduction zones [*Shearer and Masters*, 1992]. The possibility that temperatures in parts of the upper mantle may be sufficiently high to produce partial melts at depths greater than 300 km has received support from a shear wave reverberation study sampling the mantle beneath the Sea of Japan [*Revenaugh and Sipkin*, 1994]. As with the lower mantle, it now appears that thermal anomalies on the order of several hundred K exist in the upper mantle at the length scales (100-1000 km) sampled by these seismic techniques.

Tomographic studies of upper mantle heterogeneity generally show good correlation with surface tectonic features and can be used to infer the depth extent of the thermal anomalies that are associated with mid-ocean ridges and hotspots. In a high-resolution global surface wave study, *Zhang and Tanimoto* [1993] found low velocities under hot spots at 100-200 km depth while ridges showed very slow anomalies only in the upper 100 km, with the low-velocity regions shifting away from the ridge at greater depth. In contrast, *Su et al.* [1992] found that very slow anomalies under mid-ocean ridges extend continuously to at least 300 km, and hotspots are not underlain by low-velocity anomalies. Fast anomalies under continental shields were found to extend to 300-400 km depth [*Su et al.*, 1992, 1994]. Correlations between S-wave velocity, bathymetry, and basalt chemistry were found beneath the Mid-Atlantic Ridge at depths of 100-200 km, and the temperature variations at these depths were estimated to be 100-300 K [*Zhang et al.*, 1994].

Temperatures in the Crust and Lithosphere

Heat flow measurements place important constraints on the temperature profile of the crust and lithosphere. *Pollack et al.* [1993] reported a global synthesis of 24,744 heat flow measurements that constrain the global heat loss to be 44.2 x 10^{12} W/yr, some 4-8% higher than earlier estimates. The global heat flow pattern correlates well with seismic tomographic maps [*Zhang and Tanimoto*, 1993] at shallow depths, implying that seismic shear velocity variations at these depths are strongly correlated with temperature. *Stein and Stein* [1992] used improved oceanic heat flow data to construct a new model for oceanic lithosphere that is both hotter (1400°C at its base) and thinner (95 km) than previous models. New oceanic lithospheric temperature profiles have also been constructed that are consistent with heat flow and bathymetry constraints and incorporate temperature-dependent mineral properties [*Denlinger*, 1992]. General steady-state geotherms for the continental crust that satisfy heat flow data and use thermal parameters appropriate for crustal compositions have been developed as well [*Chapman and Furlong*, 1992].

Summary

The past four years have seen important developments in understanding the thermal structure of the Earth. Temperatures in the Earth's core are still uncertain because of large differences in extrapolated melting temperatures from different laboratories and because of the uncertain effects of alloying components. For the mantle, major progress has been made in determining lateral temperature differences from improved seismic imaging of velocity variations and discontinuity depths. Numerical simulations of mantle convection are increasingly approaching the complexity of the Earth's mantle and suggest that the mantle can exist in intermediate states between whole mantle and layered convection.

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T. S. Duffy and R. J. Hemley, Geophysical Laboratory and Center for High-Pressure Research, Carnegie Institution of Washington, 5251 Broad Branch Rd., NW, Washington, D.C. 20015. (e-mail:duffy@gl.ciw.edu)

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