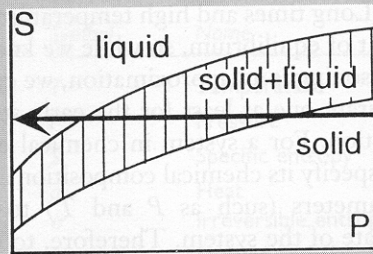


# MELTING THE MANTLE

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## GLOSSARY

- adiabatic** A thermodynamic process during which heat flow by conduction is negligible.
- batch melting** Melting process in a closed system in which liquid and solid remain in equilibrium.
- component** A chemical formula. The number of components in a system is the minimum needed to describe all possible chemical variations in the system, so a pure system with no chemical variation has one component.
- equilibrium** A state in which all natural processes have gone to conclusion and no further change is observed in any macroscopic variable.
- fractional melting** A melting and distillation process in which liquid is continuously and instantaneously removed from the system as it forms.
- isentropic** A thermodynamic process occurring at constant entropy; equivalent to *adiabatic* and *reversible*.

- liquidus** The temperature at which a liquid of given composition begins to crystallize at a given pressure.
- phase A** compositionally and structurally homogeneous material such as a mineral, liquid, or vapor.
- productivity** The increase in extent of partial melting per unit decrease in pressure, expressed in percent per gigapascal and usually a positive number.
- reversible** An idealized path that proceeds through a sequence of equilibrium states with no spontaneous processes occurring.
- solidus** The temperature at which a solid assemblage of given composition begins to melt at a given pressure.

## I. INTRODUCTION

The stony part of the Earth is solid under normal conditions at the present day. Volcanism is the eruption of molten or partially molten rock, that is, magma. The first stage of any volcanic process therefore must be melting: We have to produce a liquid by partial melting before it can migrate from the source region (see Migration of Melt chapter) and subsequently erupt (Part II). The locations of volcanoes on Earth are controlled primarily by where melting takes place. The volume, frequency, and style of eruptions are dependent on many factors but the first considerations are how much magma is supplied by melting, at what depth melting takes place,

and the distribution of melt production in the source region in space and time.

Before roughly 1960, igneous petrology was mostly concerned with the evolution and differentiation of magma after it left its source region. In the following two decades, attention shifted to the pressure ( $P$ ) and temperature ( $T$ ) at which primary melts were extracted from the mantle, without detailed understanding of how the melts were produced. In the last 20 years, however, the application of ideas from thermodynamics, experimental petrology, and fluid dynamics has provided a strong basis for understanding the basic physical processes underlying the melting of Earth's mantle. This chapter explores *how* and *why* the crust and mantle of the Earth melt, which together with information about tectonic processes and chemical composition (discussed in other chapters in this part) determines *where* and *when* melting takes place.

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## II. HOW TO MELT A ROCK

There are three ways to melt a rock. The most obvious way is to *raise the temperature* ( $T$ ). In the Earth, however, the other two methods are dominant: melting by *lowering the pressure* ( $P$ ) and by changing the chemical composition of the system to *lower its solidus temperature* below the current temperature. The *solidus* is the temperature at which a system of given composition begins to melt at a given pressure, whereas the *liquidus* is the temperature at which the system becomes completely liquid. Rocks melt over a range of temperatures so it is imprecise to speak of the melting point as we would for a pure substance. Melting by decompression and melting by addition of a flux (i.e., a component that lowers the solidus) dominate in the mantle because in the asthenosphere mass transport is much faster than heat transport and because internal heat generation by radioactivity is minor at the present era. In the lithosphere, which includes the crust, mass transport is slower and ambient temperatures are low enough that a temperature increase, whether by conductive heat flow or local heat generation, is required. For the earth as a whole, however, asthenospheric melting generates a much larger volume of magma every year than lithospheric melting.

Why three ways? The answer derives, like most of our reasoning about the physical and chemical behavior of rock systems, from thermodynamics and from the assumption of chemical *equilibrium*. Situations certainly occur, such as explosive eruptions, in which equilibrium

is a poor assumption, but at the source of melting temperatures are quite high and the processes considered are fairly slow. Long times and high temperatures favor the achievement of equilibrium, so while we know that equilibrium reasoning is an approximation, we consider it a rather accurate one at least for the early stages of magma production. For a system in chemical equilibrium, we must specify its chemical composition and two additional parameters (such as  $P$  and  $T$ ) to define uniquely the state of the system. Therefore, to cause a change of state such as melting, we see that we can group all the ways we can act on the system into changes in  $T$ , changes in  $P$ , and changes in composition. It is, as far as we know, universally true that high  $T$  favors the liquid state over the solid state, so that *raising the temperature* or *lowering the solidus temperature* (by changing composition) can lead to melting. On the other hand, while it is more common for low  $P$  to favor the liquid state over the solid state so that *lowering the pressure* can lead to melting, this is by no means universal (water is a well-known exception, and mantle rocks may experience the opposite behavior at very high pressures as well).

We can define the state of a system by specifying parameters other than  $P$  and  $T$  and in many cases it is quite convenient to do so because other parameters may be conserved or controlled (i.e., independent) during the physical process we are considering. Likewise,  $P$  or  $T$  may be dependent parameters, in which case we learn little by trying to specify them or by using them as axes on our diagrams. In an isochoric system such as an inclusion in a rigid mineral host, volume (or density) is held constant and  $P$  changes in a dependent manner as we approach equilibrium. For mantle melting, we are concerned with an approximately *adiabatic* system, that is, one in which heat flow by conduction is negligible. This is a reasonable approximation because the source regions of magmas in the mantle are large, thermal gradients are low, and the timescale for material to be transported across the source region by advection is shorter than the time for temperatures to equilibrate across the source region by thermal diffusion. If an adiabatic process is also *reversible*, then the independent variable is not  $T$  but entropy,  $S$  (see Table I). This follows directly from the second law of thermodynamics,

$$dS = \frac{dq}{T} + d\sigma, \quad (1)$$

which states that the change in entropy for any process is equal to the ratio of heat flow  $dq$  to temperature plus the irreversible entropy production  $d\sigma$ . If a process is adiabatic ( $dq = 0$ ) and reversible ( $d\sigma = 0$ ), then  $dS =$

TABLE I Symbols Used

Symbol	Name	Units
$T$	Temperature	°C, K
$P$	Pressure	GPa
$S$	Specific entropy	J/K/kg
$q$	Heat	J/kg
$\sigma$	Irreversible entropy production	J/K/kg
$H$	Specific enthalpy	J/kg
$F$	Extent of melting by mass	
$s$	Solid phase	
$l$	Liquid phase	
$\Delta S_{\text{fus}}$	Entropy of fusion (one-component system)	J/K/kg
$\Delta V_{\text{fus}}$	Volume of fusion (one-component system)	J/GPa/kg
$\alpha$	Coefficient of thermal expansion, $\frac{1}{V} \left( \frac{\partial V}{\partial T} \right)_P$	K <sup>-1</sup>
$\rho$	Density	kg m <sup>-3</sup>
$C_p$	Specific heat capacity at constant pressure	J/K/kg
$\left( \frac{\partial T}{\partial P} \right)_s$	Slope of an isentrope in $P$ - $T$ space	K/GPa
$\left( \frac{\partial T}{\partial P} \right)_{\text{solidus}}$	Slope of the solidus curve in $P$ - $T$ space	K/GPa
$-\left( \frac{\partial F}{\partial P} \right)_s$	Isentropic productivity	%/GPa
$\left( \frac{\partial T}{\partial P} \right)_F$	Slope of a constant $F$ contour in $P$ - $T$ space	K/GPa
$\left( \frac{\partial F}{\partial T} \right)_P$	Isobaric productivity	%K
$\left( \frac{\partial S}{\partial F} \right)_P^{\text{rxn}}$	Isobaric entropy of melt reaction	J/K/kg
$\left( \frac{\partial S_x}{\partial P} \right)_F$	Entropy change due to compositional effects per unit change in $P$ at constant $F$	J/K/GPa/kg
$\left( \frac{\partial S_x}{\partial F} \right)_P$	Entropy change due to compositional effects per unit increase in $F$ at constant $P$	J/K/kg
$G$	Gravitational constant ( $6.67 \times 10^{-11}$ )	m <sup>3</sup> kg <sup>-1</sup> s <sup>-2</sup>
$M_{\oplus}$	Mass of the Earth ( $6 \times 10^{24}$ )	kg
$R_{\oplus}$	Radius of the Earth (6371)	km
$X_Y^{\text{bulk}}$	Mole fraction of component $Y$ in bulk system	
$X_Y^l$	Mole fraction of component $Y$ in liquid phase	
$D_Y$	Partition coefficient for component $Y$ : ( $X_Y^l/X_Y^s$ )	

0, which is to say  $S$  is constant and we call the process *isentropic*. For our purposes, this property defines entropy. We can also imagine an adiabatic but irreversible process such as the Joule-Thomson experiment of forced flow through a nozzle in which enthalpy  $H$  is conserved. Some authors have suggested that adiabatic upwelling is an isenthalpic process (using a definition of enthalpy corrected for the presence of gravity), but although mantle melting is not strictly reversible, it is not clear why it should be isenthalpic. We find it easier to begin with the isentropic case and add known sources of entropy production (gravity and viscosity) to arrive at an accurate description of the natural process. In the section on decompression melting, we shall see why it is instructive to use the conserved quantity  $S$  in our thinking and on our diagrams rather than the dependent parameter  $T$ .

In the following sections, we review the essentials of decompression melting, flux melting, and melting by direct heating. Our principal focus is on decompression melting, since at least for the present-day terrestrial mantle, that is the dominant mechanism.

### III. DECOMPRESSION MELTING

Decompression melting is the best understood mechanism of melt production. It occurs at midocean ridges and ocean islands such as Hawaii and Iceland. The melting process underlying convergent margin volcanism is different and is discussed in a later section. The main points of this section are that from simple considerations we can show why the mantle melts when decompressed and how much melt we should expect to be produced during decompression.

#### A. One-Component System

We begin by considering a pure substance or one-component system (a *component* is a chemical formula, and the number of components in a system is the minimum needed to describe all possible chemical variations in the system, so a pure system with no chemical variation has one component). The mantle is a multicomponent system, but we can learn a great deal from the very simplest system that can then be generalized to systems that are more complex. First, let us imagine that this substance can only form two phases, a solid  $s$  and a liquid  $l$  (a *phase* is a compositionally and structurally

homogeneous material). Let our system be well behaved so that liquid is favored by high temperature and low pressure, so if we map the stable assemblage in  $P$ - $T$  space we will find a *phase diagram* like that shown in Fig. 1a. A phase diagram divides a plane or space according to what phases are stable at each point. There are three elements to the phase plane in Fig. 1a: a one-phase region where solid  $s$  is stable, a one-phase region where liquid  $l$  is stable, and a two-phase curve where  $s$  and  $l$  coexist. In this space, this two-phase curve is both solidus and liquidus. Thus if we take  $P$  and  $T$  as our independent variables, for example, along an isothermal decompression path, our one-component system in passing from one equilibrium state to another as we reversibly lower  $P$  would melt completely at a single point along the path.

Pressure and temperature have the special property that they are equal among all phases coexisting at equilibrium (except for surface tension effects), which is why the two phases  $s$  and  $l$  coexist along a curve in the  $P$ - $T$  plane. Entropy (as well as enthalpy and volume), on the other hand, differs between coexisting phases. If, therefore, we draw the phase diagram for our one-component system in the  $P$ - $S$  plane, we obtain something like Fig. 1b. This plot still has one-phase regions where  $s$  and  $l$  are stable alone, but now we see a two-phase region  $s + l$  crossed by vertical tie-lines showing  $s$  and  $l$  coexisting at the same pressure but different entropies (the length of the tie-line at each pressure is  $\Delta S_{\text{fus}}$ , the entropy of fusion). Now we recognize the boundary between the  $s$  and  $s + l$  regions as the solidus and the boundary between regions  $s + l$  and  $l$  as the liquidus. By taking entropy  $S$  as an independent variable, we have opened the two-phase curve up into a region and we see that in this case we no longer expect the system to melt or freeze completely at a single pressure. Instead, even in this one-component system, isentropic decompression leads to *partial melting*. An isentropic decompression path is shown in Figs. 1c and 1d. In the  $P$ - $S$  plane (Fig. 1d), this is a straight line, and it is obvious that the melting begins at the pressure where the solidus crosses the prescribed entropy and ends at the pressure where the liquidus crosses the prescribed entropy. On the other hand, in the more conventional  $P$ - $T$  plane (Fig. 1c), the isentropic decompression path kinks to follow the two-phase boundary for what seems an arbitrary interval: There is no simple way to predict the behavior of the system at fixed  $S$  by looking at a diagram that considers  $T$  independent.

During isentropic decompression melting, it is clear from Fig. 1c that the temperature actually decreases as the system melts, contrary to intuition. At constant

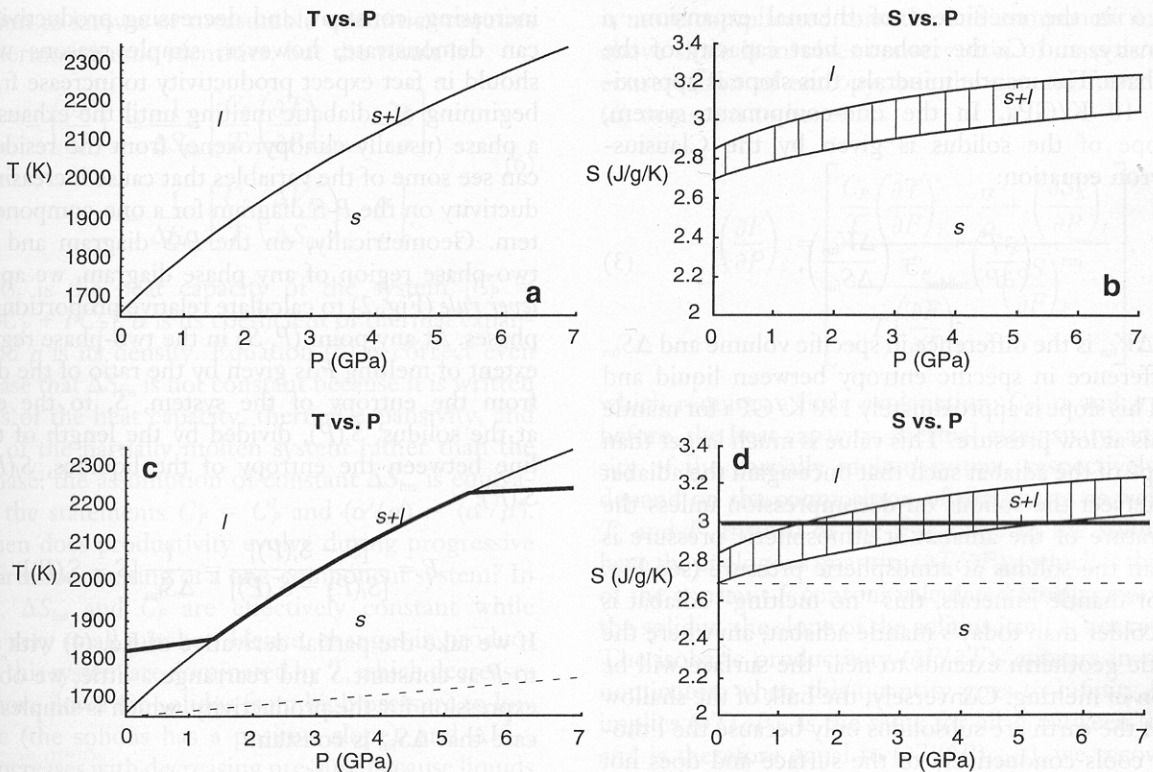


FIGURE 1 Phase relations and isentropic decompression melting of a one-component system. (a) A phase diagram in the pressure ( $P$ ) versus temperature ( $T$ ) plane is divided into a region where solid phase  $s$  is stable, a region where liquid phase  $l$  is stable, and a curved boundary  $s + l$  where the two phases coexist at equilibrium. (b) The same system plotted in the  $P$  versus entropy ( $S$ ) plane; here the two phases coexist over a region  $s + l$  bounded by the solidus and liquidus curves (which coincide in part a). The vertical tie-lines in the  $s + l$  field indicate that solids coexist with liquids at higher entropy but the same pressure. (c) The  $P$ - $T$  diagram with two isentropic decompressions path indicated; the slope of isentropes in the one-phase regions is about 10 K/GPa. The bold path shows a typical partial melting path. The light dashed path shows an adiabat so cold that melting would not occur even if adiabatic upwelling continued to the surface. During partial melting, the  $P$ - $T$  path in a one-component system is restricted to the coexistence curve  $s + l$ , but there is no way to read from this diagram how much melt is present at any point or how long the melting interval is. (d) The bold constant-entropy line superposed on the  $P$ - $S$  diagram shows that the size of the isentropic melting interval as well as the melt fraction at any point can be read directly from this diagram using the lever rule (see Fig. 2). Again, the light dashed line shows an isentrope cold enough to avoid melting.

total entropy, this behavior results because the entropy of fusion required to transform solid to liquid can only be obtained by decreasing the temperature of the system.

## B. Beginning of Melting

Isentropically lowering the pressure on a solid will generally cause it to intersect the solidus and begin partial melting. This is clear immediately on the  $P$ - $S$  diagram, where the isentrope is horizontal and the solidus has a positive slope, so that only isentropes with less entropy

than the intersection of the solidus with atmospheric pressure can avoid decompression melting altogether. On the  $P$ - $T$  diagram, this is not quite as obvious, but we can assign approximate numerical values to the slopes of the adiabat and the solidus in  $P$ - $T$  space and hence make the same argument. The slope of the subsolidus isentrope, that is, the change in temperature per unit change in pressure at constant entropy, is written in partial derivative notation  $(\partial T/\partial P)_S$  and is given by the formula

$$\left(\frac{\partial T}{\partial P}\right)_S = \frac{T\alpha}{\rho C_p}, \quad (2)$$

where  $\alpha$  is the coefficient of thermal expansion,  $\rho$  the density, and  $C_p$  the isobaric heat capacity of the solid phase. For mantle minerals, this slope is approximately 10 K/GPa. In the one-component system, the slope of the solidus is given by the Clausius-Clapeyron equation:

$$\left(\frac{\partial T}{\partial P}\right)_{\text{solidus}} = \left(\frac{\Delta V_{\text{fus}}}{\Delta S_{\text{fus}}}\right), \quad (3)$$

where  $\Delta V_{\text{fus}}$  is the difference in specific volume and  $\Delta S_{\text{fus}}$  the difference in specific entropy between liquid and solid. This slope is approximately 130 K/GPa for mantle minerals at low pressure. This value is much larger than the slope of the adiabat such that once again the adiabat will intersect the solidus on decompression unless the temperature of the adiabat at atmospheric pressure is less than the solidus at atmospheric pressure (see Fig. 1c). For mantle minerals, this “no melting” adiabat is much colder than today’s mantle adiabat; anywhere the adiabatic geotherm extends to near the surface will be a region of melting. Conversely, the bulk of the shallow parts of the earth are subsolidus only because the lithosphere cools conductively to the surface and does not lie on the mantle adiabat.

### C. Progress of Melting

To describe fully a system undergoing partial melting, we need to keep track of the composition of liquid and residue as well as the quantity of melt produced. The compositions of liquids produced by mantle melting are dealt with in later chapters. Here, we focus on the amount of melt generated. To track the amount of melt during progressive upwelling the quantity we most need to know is the *productivity*,  $-(\partial F/\partial P)_S$ , the increase in extent of partial melting  $F$  per unit decrease in pressure (expressed in percent per gigapascal and usually a positive number) and how the productivity evolves during progressive upwelling and melting. This variable is particularly hard to obtain by direct experimental measurement or from  $P$ - $T$  phase diagrams, but many authors have estimated that productivity should be about 10%/GPa for adiabatic mantle melting, which is an adequate average value.

For many purposes, the average estimate of 10%/GPa is sufficient. Going beyond this average, we may ask whether we should expect any systematic behavior in the variations of productivity during the progress of melting. Different authors have argued, variously, for

increasing, constant, and decreasing productivity. We can demonstrate, however, simple reasons why we should in fact expect productivity to increase from the beginning of adiabatic melting until the exhaustion of a phase (usually clinopyroxene) from the residue. We can see some of the variables that cause increasing productivity on the  $P$ - $S$  diagram for a one-component system. Geometrically, on the  $P$ - $S$  diagram and in any two-phase region of any phase diagram, we apply the *lever rule* (Fig. 2) to calculate relative proportions of the phases. At any point  $(P, S)$  in the two-phase region the extent of melting  $F$  is given by the ratio of the distance from the entropy of the system,  $S$ , to the entropy at the solidus,  $S_s(P)$ , divided by the length of the tie-line between the entropy of the liquidus,  $S_l(P)$ , and  $S_s(P)$ :

$$F = \frac{[S - S_s(P)]}{[S_l(P) - S_s(P)]} = \frac{1}{\Delta S_{\text{fus}}} [S - S_s(P)]. \quad (4)$$

If we take the partial derivative of Eq. (4) with respect to  $P$  at constant  $S$  and rearrange a little, we obtain an expression for the productivity, which is simplest in the case that  $\Delta S_{\text{fus}}$  is constant:

$$-\left(\frac{\partial F}{\partial P}\right)_S = \frac{1}{\Delta S_{\text{fus}}} \left(\frac{\partial S_s}{\partial P}\right)_{\text{solidus}}, \quad (5)$$

that is, productivity is given by the slope of the solidus divided by the entropy of fusion  $\Delta S_{\text{fus}}$ . To relate this

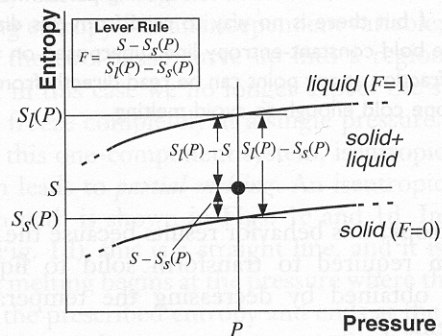


FIGURE 2 Graphical illustration of the lever rule. The lever rule is just a statement of conservation of an intensive property in a two-phase system. On the  $(P, S)$  plane, if the system has total specific entropy  $S$  then at equilibrium and at pressure  $P$ , the system will be totally solid ( $F = 0$ ) if  $S < S_s(P)$ , the specific entropy of the solid phase at the solidus. The system will be totally molten ( $F = 1$ ) at  $S > S_l(P)$ , the specific entropy of the liquid phase at the liquidus. If  $S_s(P) < S < S_l(P)$ , however, the system will be partially molten and the mass fraction of liquid  $F$  will be given by the lever rule as shown.

expression to known or measurable quantities requires some thermodynamic identities, but the result is

$$-\left(\frac{\partial F}{\partial P}\right)_S = \frac{1}{\Delta S_{\text{fus}}} \left[ \frac{C_p}{T} \left(\frac{\partial T}{\partial P}\right)_{\text{solidus}} - \frac{\alpha}{\rho} \right] \quad (6)$$

$$= \frac{1}{\Delta S_{\text{fus}}} \left[ \frac{C_p}{T} \left(\frac{\Delta V_{\text{fus}}}{\Delta S_{\text{fus}}}\right) - \frac{\alpha}{\rho} \right],$$

where  $C_p$  is the heat capacity of the system [ $C_p = (1 - F)C_p^s + FC_p^l$ ],  $\alpha$  is its coefficient of thermal expansion, and  $\rho$  is its density. Equation (6) is correct even in the case that  $\Delta S_{\text{fus}}$  is not constant because it is written in terms of the heat capacity, thermal expansivity, and density of the partially molten system rather than the solid phase; the assumption of constant  $\Delta S_{\text{fus}}$  is equivalent to the statements  $C_p^l = C_p^s$  and  $(\alpha^l/\rho^l) = (\alpha^s/\rho^s)$ . How then does productivity evolve during progressive decompression melting of a one-component system? In general,  $\Delta S_{\text{fus}}$  and  $C_p$  are effectively constant while  $\alpha/\rho$  is a very small number. Hence, changes in productivity in this system are dominated by  $T$ , which decreases as we move down the solidus from high pressure to low pressure (the solidus has a positive slope), and  $\Delta V_{\text{fus}}$ , which increases with decreasing pressure because liquids are typically much more compressible than solids (hence the solidus has negative curvature, i.e.,  $(\partial T/\partial P)_{\text{solidus}}$  increases with decreasing pressure). Both of these effects contribute to *increasing productivity* with decreasing pressure or progressive decompression melting from a given intersection with the solidus. For melting of pure diopside ( $\text{CaMgSi}_2\text{O}_6$ ) beginning at 7 GPa, productivity increases from 6%/GPa at 7 GPa to 35%/GPa when it reaches the surface at  $F = 0.9$ .

#### D. Multicomponent Systems

Earth's mantle, of course, is not a pure substance. It contains at least 10 different oxide components in amounts greater than 0.1 wt%. Here we attempt to generalize the simple insights apparent in the one-component example to systems more like the mantle. In systems of more than one component, two or more phases that differ in composition can often coexist anywhere in a region of the  $P$ - $T$  plane, so that the solidus and the liquidus can be distinct in  $P$ - $T$  space and partial melting is to be expected even when temperature is the independent variable. In other words, the isobaric productivity  $(\partial F/\partial T)_P$ , which is infinite for one-component systems, generally has a finite value. The derivation of a productivity equation for multicomponent systems

is more complicated than the one-component example above since it involves conservation of mass as well as conservation of entropy, but the general form of Eq. (6) turns out to be:

$$-\left(\frac{\partial F}{\partial P}\right)_S = \left[ \frac{\frac{C_p}{T} \left(\frac{\partial T}{\partial P}\right)_F - \frac{\alpha}{\rho} + \left(\frac{\partial S_X}{\partial P}\right)_F}{\frac{C_p}{T} \left(\frac{\partial F}{\partial T}\right)_P + \left(\frac{\partial S}{\partial F}\right)_P^{\text{rxn}}} \right], \quad (7)$$

which requires a little explanation:  $C_p$ ,  $\alpha$  and  $\rho$  are, as before, the heat capacity, thermal expansivity, and density of the partially molten system, respectively; they depend on the composition of the phases as well as  $P$ ,  $T$ , and  $F$ . Instead of the  $P$ - $T$  slope of the solidus, we have the analogous quantity  $(\partial T/\partial P)_F$ , which is the slope of the constant  $F$  contour calculated locally; away from the solidus, the slope of the solidus itself is not relevant. The isobaric productivity  $(\partial F/\partial T)_P$  appears in the denominator; when this quantity goes to infinity [which implies  $(\partial T/\partial P)_F$  is the same for all  $F$  between 0 and 1 and is therefore equal to  $(\partial T/\partial P)_{\text{solidus}}$ ], we recover the one-component version, Eq. (6). Instead of the entropy of fusion (which is ill-defined for multicomponent systems) we use  $(\partial S/\partial F)_P^{\text{rxn}} = S^l - S^s + (\partial S_X/\partial F)_P$ , which we call the *isobaric entropy of melt reaction*. The terms  $(\partial S_X/\partial P)_F$  and  $(\partial S_X/\partial F)_P$  involve a subtle point in the entropy budget, but for multicomponent systems they must be included both to obtain an accurate productivity equation and to define precisely the quantity that acts like the entropy of fusion. In a closed system where the coexisting phases differ in composition (e.g., basalt and residual peridotite), moving an increment of mass from solid to liquid necessarily involves changes in the composition of at least one (and generally all) of the phases. Similarly, changing pressure at constant melt fraction involves changes in the equilibrium composition of various phases because phase equilibria and partition coefficients depend on pressure and temperature. In turn, the specific entropies of the phases depend on their composition and the specific entropy of the system depends on which phases the components reside in. Therefore an equation such as Eq. (7) that expresses conservation of mass and conservation of entropy needs to include sums over all components and phases of the partial specific entropy of each component in each phase times the changes in mass fractions of the components and phases per increment of melting at constant pressure,  $(\partial S_X/\partial F)_P$ , and per increment of pressure change at constant melt fraction,  $(\partial S_X/\partial P)_F$ .

With Eq. (7) in hand, we can discuss the variables that determine productivity and how it changes during melting of various simple and natural multicomponent systems. We need to know all the variables in Eq. (7) to calculate the actual value of productivity, but here we focus on variations in productivity during progressive upwelling. As in the one-component case,  $C_p$ ,  $\alpha$ , and  $\rho$  are not important sources of variation. It also turns out that  $(\partial S_X/\partial P)_F$  is small and that, although according to some models  $(\partial S/\partial F)_{P,T}^{\text{sm}}$  may vary by nearly a factor of 2 during melting of fertile peridotite, the first term in the denominator is larger in magnitude and dominates the variations in productivity. Also,  $T$  varies systematically, but we find that its influence is generally swamped by the variations in  $(\partial T/\partial P)_F$  and especially  $(\partial F/\partial T)_P$ . To a reasonable approximation then, we can understand the behavior of productivity during isentropic upwelling of multicomponent systems just by considering the behavior of  $(\partial T/\partial P)_F$  and  $(\partial F/\partial T)_P$ . We can state with considerable certainty that except in the vicinity of changes in the solid phase assemblage  $(\partial T/\partial P)_F$  increases with decreasing pressure like the slope of the solidus in the one-component system, simply because silicate liquids are more compressible than minerals. This contributes to increasing productivity with decreasing pressure, just as in the one-component system.

The behavior of  $(\partial F/\partial T)_P$  has been the subject of considerable disagreement among researchers. We present here a simple argument that leads us to expect that  $(\partial F/\partial T)_P$  increases with  $F$  away from the solidus in natural peridotites, but experimental data are equally consistent with nearly constant  $(\partial F/\partial T)_P$  and many authors have constructed models in which  $(\partial F/\partial T)_P$  decreases with  $F$ . Indeed isobaric productivity is sometimes taken as infinite up to some  $F$ , which is the case for univariant melting as in the one-component system; this behavior is sometimes misleadingly called *eutectic*. The idea that melting of peridotite is invariant descends from the behavior simple systems such as CaO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>, which are thought to be suitable analogues of the mantle because these oxides constitute roughly 90% of the mass of mantle rocks. It turns out, however, that minor components and solid solutions dominate the melting behavior at low melt fractions. The simplest model system that captures the effect of a minor component such as Na<sub>2</sub>O on productivity is a binary system with one major component  $Z$  that forms a nearly pure solid phase and one minor component  $Y$  that partitions preferentially into the liquid phase [that is, its partition coefficient  $D_Y \equiv (X_Y^l/X_Y^s)$  is less than one, where  $X_Y^s$  and  $X_Y^l$  are the mole fractions of component  $Y$  in the solid and liquid phases, respectively]. At low concentrations of  $Y$ , the liquidus temperature of such a system is described

approximately by the cryoscopic or freezing point depression relation

$$X_Y^l = k(T_Z^{\text{fus}} - T), \quad (8)$$

where  $T_Z^{\text{fus}}$  is the melting point of pure solid  $Z$  and  $k$  is a constant that depends only on the properties of  $Z$ . To solve for the behavior of  $F$  and  $(\partial F/\partial T)_P$ , we only need the standard equations for the concentration of a trace element in the liquid as a function of  $F$  during batch and fractional melting (*batch melting* is a closed system in which liquid and solid remain in equilibrium; *fractional melting* is a distillation process in which liquid is continuously and instantaneously removed from the system as it forms):

$$X_Y^l = \frac{X_Y^{\text{bulk}}}{F + (1-F)D_Y} \text{ (batch)} \quad (9)$$

$$X_Y^l = \frac{X_Y^{\text{bulk}}}{D_Y} (1-F)^{(1/D_Y-1)} \text{ (fractional)}. \quad (10)$$

The phase diagram and isobaric melting relations of this system are shown in Fig. 3. For any given bulk composition, the productivity  $(\partial F/\partial T)_P$  is lowest at the solidus, where the concentration of the incompatible element in the liquid and hence the temperature [these are linearly related per Eq. (8)] is changing most quickly per unit change in  $F$ . At the solidus the productivities of batch and fractional melting are equal, but the productivity of batch melting increases more quickly at low  $F$ . At higher  $F$ , where the concentration of  $Y$  in the liquid during fractional melting becomes nearly constant at zero, the productivity of fractional melting becomes larger than that for batch melting. Furthermore, at equal  $F$ ,  $(\partial F/\partial T)_P$  is *lower* in instances of this system with higher bulk concentrations of the incompatible component  $Y$ , although the opposite is true at equal  $T$ .

The productivity behavior of the systems described by Eq. (8) through (10) is much like that of systems like forsterite-fayalite or diopside-hedenbergite, that is, a nearly ideal solid solution in equilibrium with a liquid solution. Using the solidus and thermodynamic properties of the minerals diopside and hedenbergite, the isentropic melting behavior of such a system can be calculated using Eq. (7) and is compared with the isentropic melting behavior of the diopside-anorthite eutectic binary and with the results of a thermodynamic model of natural peridotite melting in Fig. 4. The eutectic and peridotite systems all show a prominent drop in productivity at the exhaustion of clinopyroxene from the residue, but as long as the residual phase assemblage remains the same, all of these systems show very similar behavior, namely, monotonically increasing productivity as a function of increasing  $F$  or decreasing  $P$ . Figures 4e and

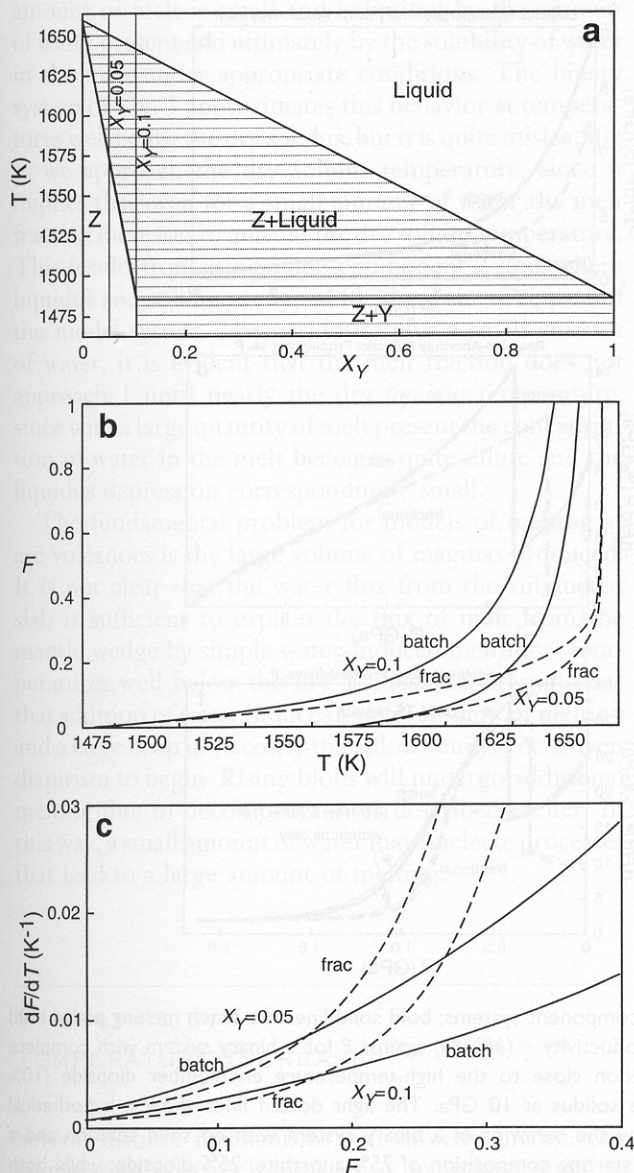


FIGURE 3 Isobaric melting behavior of a simple binary system containing a solid phase Z in which the component Y is incompatible. The partition coefficient  $D_Y$  has the constant value 0.1 in this example. (a) A phase diagram in the  $T - X_Y$  plane is divided into a one-phase region where solid Z exists alone, a one-phase region where liquid solution exists alone, a two-phase region where Z coexists with liquid 10 times richer in Y, and a two-phase region where Z coexists with pure solid Y. Two bulk compositions are marked by vertical lines. For any bulk composition  $X_Y$  and temperature  $T$  in the two-phase Z + liquid region, the equilibrium melt fraction  $F$  can be read from this diagram using the lever rule. (b) The isobaric melting behavior of the two bulk compositions marked in part (a), plotted as melt fraction  $F$  versus  $T$ . (c) Isobaric productivity  $(\partial F/\partial T)_P$  versus  $F$  along the melting paths plotted in part (b).

4f show that, for natural peridotite, a calculation of productivity that holds all variables constant except  $(\partial T/\partial P)_P$  and  $(\partial F/\partial T)_P$  provides an adequate approximation to the results of the full thermodynamic calculation of all the variables in Eq. (7).

The results of all of these considerations are that adiabatically upwelling mantle will produce melt with an average productivity of about 10%/GPa, but this average probably conceals variations from as little as 2%/GPa to as much as 30%/GPa. Melting proceeds from the intersection of the adiabat and the solidus up to the base of the lithosphere. Specifically, melting is stopped either by the mechanical lithosphere (which arrests upwelling) or the thermal lithosphere (where conduction cools the system), whichever is thicker. A thick lithosphere, such as we expect at oceanic islands like Hawaii, stops melting without regard to considerations of productivity. Where lithosphere is thin, for example, at a fast-spreading midocean ridge, the large productivity approaching exhaustion of clinopyroxene and the decreased productivity thereafter imply that over a fairly wide range of mantle temperatures, the peak extent of melting will be near that required to exhaust clinopyroxene from the source. This corresponds to observations of many oceanic peridotite samples and to the typical melt fractions inferred from normal midocean ridge basalt. Under most circumstances, the low productivity "tail" at the beginning of melting contributes a small volume of high-pressure melts, often with distinctive chemistry, to the aggregate liquid.

#### IV. MELTING BY CHANGING COMPOSITION

After decompression melting, melting induced by changes in composition is the next most important melting process in the earth. It is mostly responsible for island-arc volcanism, whose often explosive manifestations dominate the rest of this encyclopedia. The origins of arc volcanism are only imperfectly understood; tracing contributions from the subducting slab, subducted sediment, mantle wedge, and overriding plate represents a major area of contemporary research. Still, the basic phenomenon of subduction puts a cold slab into the mantle, generating low temperatures and downward flow. We know from midocean ridges and hot spots that decompression melting requires upward flow and generates more magma given higher temperature. The only reasonable process to induce melting in such a setting is the addition of components such as water that

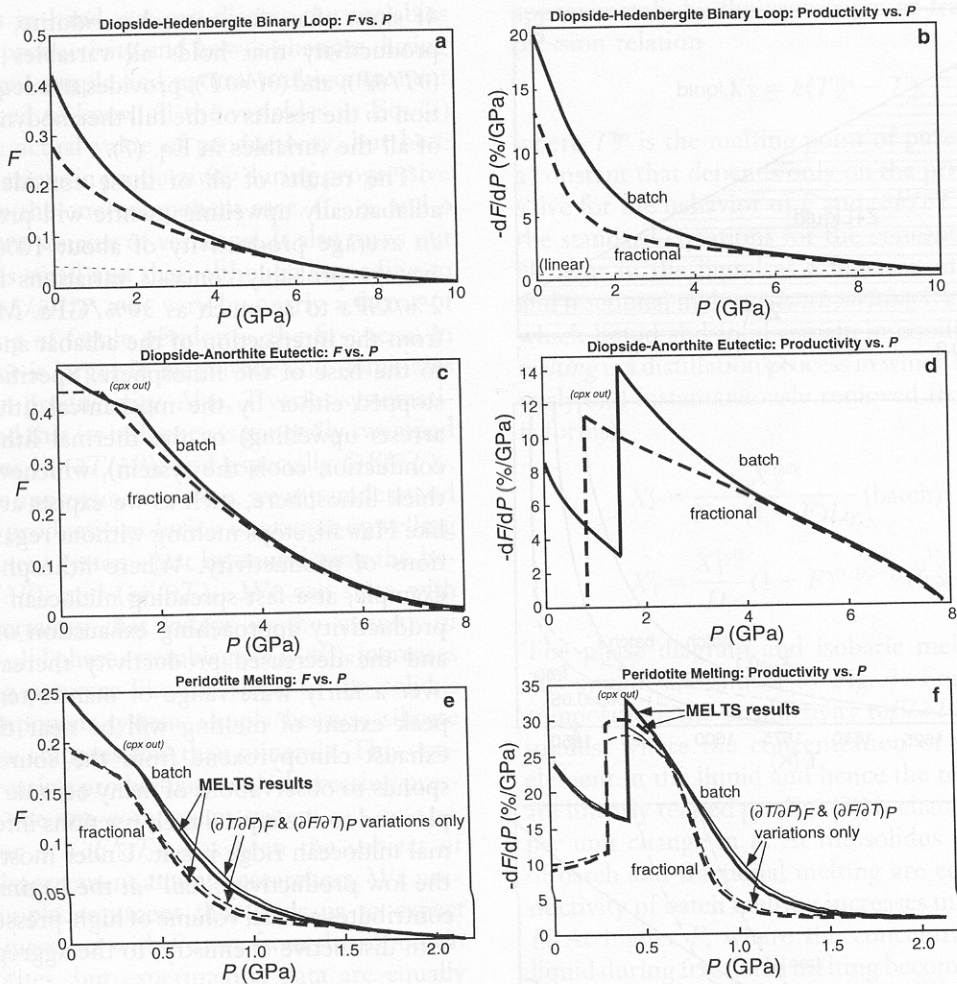


FIGURE 4 Isentropic, polybaric melting behavior of three multicomponent systems; bold solid lines are batch melting paths, bold dashed lines are fractional melting paths. (a) and (b) show  $F$  and productivity  $-\frac{dF}{dP}$  against  $P$  for a binary system with complete solid-solution (diopside-hedenbergite), plotted for a bulk composition close to the high-temperature endmember diopside (10% hedenbergite, 90% diopside) along an isentrope that intersects the solidus at 10 GPa. The light dotted lines show a hypothetical constant-productivity melting path for comparison. (c) and (d) show the behavior of a binary system with a eutectic point, diopside-anorthite. The particular case shown has a starting composition of 75% anorthite, 25% diopside; while both phases and liquid are present, the system is restricted to the eutectic curve and behaves much like the one-component system. When diopside is exhausted from the residue, there is a large drop in productivity of batch melting and a total cessation of fractional melting. (e) and (f) show a thermodynamic model of natural peridotite melting that combines aspects of the behavior of both of the preceding systems. Within regions where the residual phase assemblage does not change, productivity increases during progressive melting. The thin curves show an approximate calculation in which all variables are held constant except  $(\partial T/\partial P)_F$  and  $(\partial F/\partial T)_P$  (see text); this approximation captures most of the productivity variations in this system. As in the binary case, there is a large drop at the exhaustion of clinopyroxene from the residue, but it does not stop melting altogether, even for the fractional case.

drastically lower the solidus temperature of peridotite. That the magmas produced and erupted at arc volcanoes contain abundant water is evident from their explosive behavior.

We can use the binary system illustrated in Fig. 3, with some caution, to discuss the effect of addition of water to peridotite. Like any component that preferen-

tially partitions into silicate melts rather than solids, it expands the stability field of liquid and lowers the solidus temperature. Hence, a water-bearing system can be partially molten at temperatures where the dry system would be entirely solid. The amount of melt produced is a function of the amount of water added as well as temperature. At temperatures below the dry solidus, the

amount of melt is small and is limited by the amount of water present and ultimately by the solubility of water in the melt under appropriate conditions. The binary system of Fig. 3 approximates this behavior at temperatures well below the dry solidus, but it is quite misleading as we approach the dry solidus temperature, since it implies that even for a small amount of water the melt fraction increases to unity at the dry solidus temperature. This results from using a one-component system where liquidus and solidus are equal for the water-free part of the model system. For peridotite plus a small amount of water, it is evident that the melt fraction does not approach 1 until nearly the dry *liquidus* temperature, since with a large quantity of melt present the concentration of water in the melt becomes quite dilute and the liquidus depression correspondingly small.

The fundamental problem for models of melting at arc volcanoes is the large volume of magmas produced. It is not clear that the water flux from the subducted slab is sufficient to explain the flux of melt from the mantle wedge by simple water-induced melting at temperatures well below the dry solidus. An alternative is that addition of water induces a small amount of melting and a large drop in viscosity that allows buoyancy-driven diapirism to begin. Rising blobs will undergo additional melting due to decompression as described earlier. In this way, a small amount of water may nucleate processes that lead to a large amount of melting.

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## V. MELTING BY TEMPERATURE INCREASE

Several distinct mechanisms can directly increase the temperature of a rock mass so as to induce melting. None is volumetrically significant on today's Earth, but each may have been important at some time in Earth's history or on other planets.

### A. Impact Melting

Impacts of extraterrestrial objects large enough to induce substantial amounts of melting are rare at present. When earth was forming, however, it was constantly being bombarded by large objects, including possibly a few impacts large enough to melt the entire mantle and induce a magma ocean. We can easily calculate the amount of energy carried by the impact of an object of given mass and hence to place some bounds on the amount of melting that may occur. In detail, though,

large impacts are complex and only imperfectly understood and the efficiency of melt production is difficult to calculate with any precision.

For concreteness, let us consider impacts on Earth after the time when it had essentially achieved its modern mass ( $M_{\oplus} = 6 \times 10^{24}$  kg) and radius ( $R_{\oplus} = 6371$  km). The minimum kinetic energy per unit mass of an incoming object is equal to the gravitational potential at the Earth's surface relative to infinite separation,  $GM_{\oplus}/2R_{\oplus} = 3.14 \times 10^4$  kJ/kg, corresponding to a minimum velocity  $\sqrt{2GM_{\oplus}/R_{\oplus}} = 11.2$  km/s. The enthalpy of fusion of the Earth's mantle is about 750 kJ/kg. Hence if all the kinetic energy of the impact went into melting (requiring the target to start at the melting point, and totally unphysical compression behavior) an impactor can melt at most itself plus 40 times its own mass of the target planet. The impact of a Mars-sized planetesimal with  $\sim 10\%$  the mass of the Earth (such as the event which is thought to have created the Moon) seems likely to result in complete melting of the mantle.

In practice, hypersonic collisions heat and compress the target material along an irreversible thermodynamic path and accelerate much of the target material out of the crater. Thus, much of the impact energy goes into dissipation and kinetic energy. For large cratering events, the work against gravity involved in excavating the crater and lofting the ejecta consumes more of the impact energy. Also the target is generally well below the melting point and so much of the energy goes into heating without inducing melting. Furthermore, the energy of impact ends up very unevenly distributed in the target. Closest to the impact, materials will receive so much energy as to be vaporized, which consumes a great deal of energy and leaves less available to melt rocks that see lower shock pressures.

When large craters on Earth are examined, there typically is a thin (a few meters) sheet of impact melt lining the crater and underlying the ballistic ejecta blanket. Likewise, roughly 1% of the lunar soil consists of glass droplets that presumably originated as molten ejecta from nearby craters. These observations demonstrate that impacts do produce some melt, but at the same time support the argument that impact is an inefficient process for generating magma. On Venus, radar observations taken by the *Magellan* spacecraft reveal evidently fluid outpourings from large impacts, but it is not at all clear whether these flows consist of impact melt or fluidized debris. Finally, although the largest impact basins on the Moon are filled with voluminous floods of Mare basalt, this volcanism was not a direct product of the impact but a later process that preferentially filled the topographic depressions of the impact basins with basaltic magma denser than the lunar crust.

### B. Radioactive Heat Generation

Natural radioactive heat sources are too weak at the present day to bring any rock mass to its melting point or power its melting. Furthermore, the Earth almost certainly began very hot, not because of radioactivity, but rather from the release of gravitational potential energy involved in assembling a planet (unless it accreted so slowly that radiation to space maintained a cool starting Earth, but this scenario is not favored by dynamic models or radiometric studies of accretion). Nevertheless, radioactivity bears mention as a source of heat for melting for two reasons. First, it has certainly contributed to keeping the Earth hot and volcanically active for longer than would have been possible with only the heat of accretion. Second, in the very early solar system it was the principal source of heat for the melting of small planetesimals that formed the igneous meteorites.

The principal heat-producing isotopes at present are  $^{40}\text{K}$ ,  $^{235}\text{U}$ ,  $^{238}\text{U}$ , and  $^{232}\text{Th}$ . In material of bulk-Earth composition these presently contribute a total of  $\sim 1.5 \times 10^{-4}$  J/kg/year in radiogenic heat. These elements are all strongly partitioned into liquids during melting, however, and so are highly concentrated in the continental crust. Average continental crust generates about 100 times more radiogenic heat than bulk-Earth material. While the excess heat flow from the continental crust is clearly measurable, it is still insufficient to heat crust of normal thickness to its melting point or provide the latent heat of fusion on reasonable timescales. Only in the very early solar system, when the heat flux from K, U, and Th was  $\sim 5$  times higher but more importantly short-lived radionuclides such as  $^{26}\text{Al}$  and  $^{129}\text{I}$  were active, could radioactivity by itself heat an asteroid-sized body enough to cause differentiation (melting and core formation).

### C. Heating by Conduction

In continental regions, away from plate boundaries and hot spots, volcanoes do occur that require a mechanism other than decompression melting. Their source materials are embedded in the lithosphere, which, aside from rapid and large-scale tectonic extension, prevents drops in pressure fast enough to overwhelm thermal conduction. Conduction therefore is the most likely mechanism for bringing these sources into their melting range by direct heating. This begs the question of why conduction should deliver more heat in one place than another, and the answer generally goes back to decompression melting in the underlying mantle. Geochemical and geophysical evidence typically shows that, although the

principal source of intraplate volcanism lies within the crust, there is often a mantle component. Ponding of basaltic magma and subsequent underplating of basaltic rocks at the base of the crust is the most efficient way to focus a large heat flow into a specific region of the crust. Because basalts crystallize at temperatures above  $1000^\circ\text{C}$  and the rocks of the continental crust can begin melting (in the presence of water) near  $700^\circ\text{C}$ , it is clear that crustal melting is a likely consequence of the arrival of a large mass of basalt at the base of the crust.

### D. Frictional Heating and Viscous Dissipation

Neither of these mechanisms is likely to generate a sufficient mass of magma to migrate to the surface and form a volcano. Exhumed fault zones do sometimes contain a narrow zone of glassy rock called pseudotachylyte that appears to represent melt produced by frictional heating, and this process has recently been invoked to explain the faulting mechanism of some very large deep-focus earthquakes. Furthermore, exposed low-angle normal faults and metamorphic core complexes are spatially associated with syntectonic plutonism and some volcanism, but the mechanisms more likely include decompression by isostatic rebound of the footwall and melting of the hanging wall by direct juxtaposition against a hot footwall.

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## VI. THE MANTLE ADIABAT

In the preceding discussion of decompression melting, we noted that decompression is likely to cause mantle rocks to intersect their solidus and begin melting. Very cold mantle, however, could reach atmospheric pressure without intersecting its solidus, and very hot mantle might be partially molten all the way to the core-mantle boundary. The Earth's mantle happens to be in between these two states such that its typical adiabat intersects the solidus in the range of 50 to at most a few hundred kilometers (Fig. 5). This depth range corresponds to a temperature range (measured at equal depth below the solidus) of  $200\text{--}300^\circ\text{C}$  sampled along midocean ridges, ranging from the deepest, thinnest crust ridge segments up to ridge-centered hot spots such as Iceland. Away from ridges, it is more difficult to map the temperature variation in the mantle because the depth of melting is not generally available as a probe. Instead, we rely on models of thermal convection and on other indirect

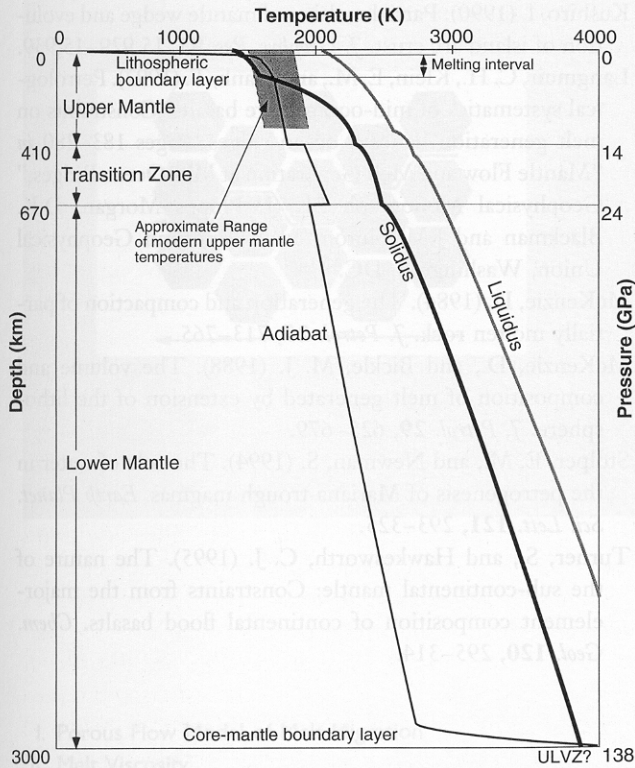


FIGURE 5 Schematic illustration of the temperature structure of the Earth's mantle at the present era. The mantle is conventionally divided into upper mantle, transition zone, and lower mantle, with the boundaries set at the 410- and 670-km seismic discontinuities. The solidus (heavy black curve) and liquidus (heavy shaded curve) are inferred from melting experiments on probable mantle compositions; these curves bracket the range where partial melting can occur. A probable typical temperature structure of the mantle is shown assuming whole-mantle convection with a cold upper boundary layer (the lithosphere) and a hot lower boundary layer where heat is conducted out of the core. In between these boundary layers the temperature profile is adiabatic (see text and glossary). The shaded band shows the likely global range of temperatures of oceanic upper mantle; anywhere the adiabatic part of any of these geotherms extends to shallow pressure, melting will occur. Except possibly for the recently discovered ultra-low-velocity zone (ULVZ) at the core-mantle boundary, melting is restricted to the upper 100 or 200 km of the mantle.

indicators of temperature such as seismic velocity, seafloor depth, and gravity. All of these lines of evidence are broadly consistent with a mantle that contains a very cold upper thermal boundary layer that is subducted into the interior at trenches as well as hot thermal plumes rising from a lower boundary layer that transport  $\sim 10\%$  of total heat flow through the upper mantle. The magnitude and origin of temperature variations away from slabs and plumes remain controversial.

Considering that the mantle is 2900 km deep, the depths at which mantle adiabats intersect the solidus are remarkably close to the surface. It is legitimate to inquire whether this is a coincidence or the result of some regulating mechanism, and hence whether this depth range is particular to the present era and steadily declining or else maintained at a quasi-steady value over long periods of geologic history. Because the viscosity of rocks is exponentially dependent on inverse temperature, it is clear that the temperature of a convecting system such as the mantle does regulate itself so that it can flow fast enough to deliver to the surface all the heat that enters the bottom and is produced internally. There is, however, no necessary relationship between the temperature that provides sufficiently low viscosity for mantle convection and the temperature at which melting occurs. Standard plate tectonic theory assigns no particular role to basaltic volcanism in the heat engine of convection; the mantle cools itself primarily by forming and subducting oceanic lithosphere (not crust). One may speculate, however, that the mantle adiabat dwells in the range where basaltic volcanism occurs because basaltic volcanism is somehow a necessary part of the plate tectonic cycle. Perhaps the formation of basalt, which converts to the very dense rock eclogite when subducted, is necessary to sustain subduction. If so, it seems likely that the mantle temperature will remain high enough to support ocean ridge volcanism until its heat flow is low enough to be transported by a qualitatively different and less efficient mechanism than plate tectonics. This remains a question for future work.

## VII. SUMMARY

Decompression melting occurs anywhere that upward flow of mantle reaches shallow depths. Such upward flow occurs at midocean ridges due to plate spreading and at hot spots due to thermal plumes in the mantle. Melting begins at the intersection of the adiabat with the solidus curve of the mantle source composition. Melting proceeds at an increasing productivity with an average value near  $10\%/GPa$  until the base of the lithosphere stops upwelling or cools the system off the adiabat. The addition of components such as water that substantially lower the solidus temperature of mantle rocks induces melting above subducting slabs in spite of low temperatures and downward flow. Occasional small volumes of magma may be produced by other means such as conductive heating by underlying magma intrusions, but decompression and fluxing are the domi-

nant sources of the magmatic liquids that supply the world's volcanoes.

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#### SEE ALSO THE FOLLOWING ARTICLES

Geothermal Systems • Mantle of the Earth • Migration of Melt • Physical Properties of Magmas • Volcanism on the Moon • Volcanism on Venus

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