Figure 3.13. Cooling of the Earth assuming heat loss is purely by conduction ($\kappa = 10$)

Figure 3.3. Isobaric (1 atm = 0.1 MPa) T-X phase diagram in the system FeO-MgO-SiO$_2$ exhibiting complete solid solution of Fe and Mg in olivine and in the melt. From [Bowen and Schairer, 1935].
Figure 3.17  Same as in Figure 3-16, except that temperature-depth paths of upwelling mantle plumes are shown.
Figure 6.16. Melting of the mantle as it rises along an adiabat.

normal gradients of the lithosphere, and the rocks are cold and rigid on insulating the mantle. The boundary of the lithosphere is steep. 7.5 is a fairly adjacent would, however, the liquid state solidus, for the adiabat and the variation of the liquid if the solidus would be thinner, where the temperature is lower.

melting, Figure 7.6(b) illustrates the effect of rising hot, low-density mantle 'burning' into the base of the lithosphere until it induces partial melting by intersecting the dry solidus at 100 km depth. With time, a large shallow melt zone (shaded in Fig. 7.6(b)), extending almost to the surface, could develop; this type of upwelling may occur beneath some 'hot spots' and ridge zones, such as in Iceland (discussed further in Chapters 8 and 9). Finally, there are circumstances (Fig. 7.6(c)) in which mantle at normal temperatures can be caused to melt; this requires that the thermal boundary layer is thinned dramatically, bringing about the necessary decompression. This will occur where the lithosphere is under strong extensional stress – for example at ocean ridges. This model of decompression melting is developed further in Section 7.4.

Figure 7.6 Development of a zone of possible melt production (shaded) under three different scenarios. (a) Lherzolite solidus is depressed by the presence of volatile species in the mantle, principally CO₂ and H₂O produced by the breakdown of minerals such as mafic amphibole above the exhumed solidus; intersection with normal geotherm (from Fig. 7.5) dictates the depth limits of the possible melt zone. (b) Rising high-temperature material in the mantle with a potential temperature of 1500°C thins the lithosphere from beneath and causes melting at the dry solidus (from Fig. 7.5) in the upper mantle. (c) Similar to (b) but with 1280°C potential temperature (the normal mantle adiabat) brought closer to the surface due to decompression (depressing) of the shallow mantle. (Sources: McKenzie and Bickle, 1988; Wyllie, 1992, with adaptations.)
**Mantle and oceanic crust**

Figure 7.9 Summary of the thermal structure beneath a passive spreading ridge with a mantle potential temperature of 1280°C. Mantle material rises (joint arrows) to form an extensive melt zone beneath the point of plate separation, due to strong extensional stresses (heavy arrows) that cause the upper mantle to become compressed. (After McKenzie and Hickey, 1988.)

**Figure 6.13.** Simplified columnar section of the Samail ophiolite (Hopson et al., 1981).

7.10 provide a summary at two scales of the melting regime and convective development of oceanic crust at spreading centres.

[Signature: Brown, M. R. 1993]
BASALT TETRAHEDRON

PLANE OF SILICA SATURATION

PLANE OF SILICA UNDERSATURATION

Di (Cpx)

Fo (Ol)

Ne

Ab (Flm)

En (Opx)

Qz

Cpx

Ne

Pl

Ol

Ol

Olivine Basalt

Alkali Basalt Group

Cpx

Olivine Tholeiite

Olivine Tholeiite Group

Cpx

Hz

Opx

Qz

Pl

Olivine Tholeiite

Quartz Tholeiite Group

Cpx

Opx

Qz

Gabbro/Basalt

Cpx
Figure 4-4 Compositional fields of volcanic rocks in terms of total alkalis and silica. Reprinted fromBas et al., 1986; published with permission of Oxford Univer-
Figure 8. Variations in the maximum MgO content of picrites and komatiites plotted against time. $T_p$ gives the approximate mantle potential temperature.

(Campbell and Griffiths, 1992)