Chapter 8

Subduction and arc formation

The age-bathymetry-heatflow relationship in the ocean basins reflects the time dependent cooling and densification associated with lithospheric thickening (and, moreover, provides strong support for the notion that the oceanic lithosphere is thermally stabilised). Since the great proportion of old oceanic lithosphere is cold mantle peridotite, after some critical time the oceanic lithosphere must become more dense than the hotter peridotite of the subjacent convective asthenosphere thus allowing the possibility of subduction. The gravitational body forces acting on old dense ocean lithosphere give rise to the second fundamental driving force for the motion and deformation of the lithosphere, termed slab pull. The subduction process leads to melting of both the subducting slab and the overlying mantle wedge; the resultant melts segregate to form magmatic arcs (island arcs when they are built on oceanic lithosphere and *Andean* or *Cordilleran* arcs when they are built on continental lithosphere).

8.1 Buoyancy of the ocean lithosphere

The condition required for the onset of subduction is the negative buoyancy of the ocean lithosphere, i.e.:

$$\frac{1}{z_l}\int_0^{z_l} \rho_z \geq \rho_m \tag{8.1}$$

where ρ_z is the density at depth z within the lithosphere of thickness z_l and ρ_z is the density of the convective mantle. Assuming an appropriate density structure for the ocean lithosphere we can solve for the critical lithospheric thickness, z_{lc}, for which Eqn 8.1 is true. We begin by assuming the simplest of possible density structures in the ocean lithosphere, as shown in Figure 8.1:

$$\rho_z = \rho_c \qquad 0 < z < z_c$$
$$\rho_z = \rho_m \left(1 + \alpha\, T_l\, \frac{z}{z_l}\right) \qquad z_c < z < z_l \tag{8.2}$$

where ρ_c is the density of the crust. The appropriate equality for z_{lc} is:

$$\frac{1}{z_{lc}} \int_0^{z_{lc}} \rho_z \, dz = \rho_m \tag{8.3}$$

Integration of Eqn. 8.3 gives:

$$\frac{1}{z_{lc}} \left(\rho_c\, z_c + (z_{lc} - z_c)\left(\frac{\rho_m\, \alpha\, (T_l - T_c)}{2}\right)\right) = \rho_m \tag{8.4}$$

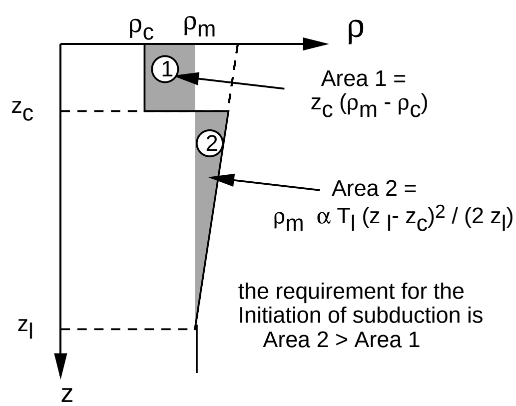

Figure 8.1: Density structure assumed for solution of Eqn 8.1 where T_c is the temperature at the Moho.

For a linear geotherm T_c is given by:

$$T_c = T_l \left(\frac{z_c}{z_{lc}}\right)$$

An identical solution is readily obtained graphically by simply making the equivalence between the two regions in Figure 8.1:

$$z_c\,(\rho_m - \rho_c) = \rho_m\, \alpha\, T_l \frac{(z_{lc} - z_c)^2}{z_l\, 2} \tag{8.5}$$

Substituting in the following values appropriate to the ocean lithosphere : $z_c = 6000$ m, $\rho_c = 3000$ kg m^{-3}, $\rho_m = 3300$ kg m^{-3}, $T_m - T_s = 1250^o$C, $\alpha = 3$ x $10^{-5\,o}$K^{-1}, gives the critical lithosphere thickness of 40 km. Substituting z_{lc} into Eqn 8.3 yields the critical age of 10 Ma, for the onset of negative buoyancy. In this calculation we have assumed that the only compositional changes associated with the production of oceanic lithosphere are within the oceanic crust due to the basalt extruded at the ridge axis. However, the extraction of basalt from the underlying peridotite must leave a depleted "harzbugitic" residual mantle peridotite, the density of which is somewhat less than the primary mantle peridotite by an amount relating to the proportion of melt extracted (50 kg m^{-3} for 25% extraction). This "depleted" zone extends beneath the ridges to the base of melting (about 40 km for normal mantle with a potential temperature of 1280^oC, Figure 7.3), and is eventually frozen onto the base of the cooling lithosphere. Calculations which account for this "depleted" nature of the upper 40 km of oceanic mantle lithosphere show that the age for onset of negative buoyancy is significantly greater than calculated assuming the density structure as in Eqn 8.2 and Figure 8.1. For example, Oxburgh & Parmentier (1977) calculate a critical age of approximately 50 Ma.

The above analysis shows that much of the old oceanic lithosphere will have the potential for initiating subduction, with the precise age dependent on the average density of the oceanic lithosphere. Negatively buoyant lithosphere may however be held up by the elastic strength of surrounding buoyant lithosphere, and the actual mechanism which triggers the onset of subduction is poorly understood. Moreover, younger positively buoyant oceanic lithosphere may be subducted if it is dragged into pre-existing subduction zones by the pull of the older subducting slab and the push of the existing ridge. Indeed, some workers have suggested it is possible that ocean ridges can be subducted, even though they are necessarily positively buoyant.

8.2 Therml structure of subducted slabs

The penetration of the mantle by the subducting slab must disturb the thermal regime of the mantle, providing the slab does not heat up by diffusion at a rate fast compared to subduction. The question, then, is whether the advective or the diffusive terms in Eqn 4.3 dominate the thermal evolution of the downgoing slab. Simple application of the dimensional analysis outined in section 4.2 allows us to resolve this problem by determining the thermal Peclet number appropriate to subduction (if $Pe_T > 10$ then advection dominates weheras if $Pe_T < 1$ diffusion dominates). The thermal Peclet number is given by :

$$Pe_T = \frac{v\,l}{\kappa}$$

v is given by subduction velocities (> 1 cm/yr), l by the approrpiate length-scale across which diffusion must act, which is here defined by the thickness of thelithosphere (100 x 10^3 m) and κ by the thermal diffusivity (10^{-6} m^2 s^{-1}. The Pe_T appropriate to subduction is therefore about 30, thus implying that subduction advects the thermal structure of the ocean lithosphere into the deep mantle.

8.3 The magnitude of slab pull

Subduction occurs due to the gravitational body forces acting on the descending slab, which are proportional to the density contrast between the slab and the surrounding hotter mantle. This density contrast, which drives subduction, arises because of (Figure 8.2):

- the temperature differences between the cool descending slab and the surrounding mantle which gives rise to a *thermal density defect* per unit length of slab of order 10^{13} N m^{-1}, and

- density changes accompanying metamorphism of the descending slab such as eclogitizaton of the basaltic crust and shallowing of the olivine-spinel transition, which give rise to a *compositional density defect* also of order 10^{13} N m^{-1}.

The force balance on the descending slab must also include a term related to the viscous resistance of the mantle to the downward movement of the slab, which will be proportional to subduction velocity. Because the rheological properties of rocks at high temperatures are poorly known, this viscous term in the force balance cannot yet be quantified to any reasonable levels of confidence. An analytical formulation of the slab pull is therefore far more complicated than for ridge push, and here we treat the problem in a qualitative way only.

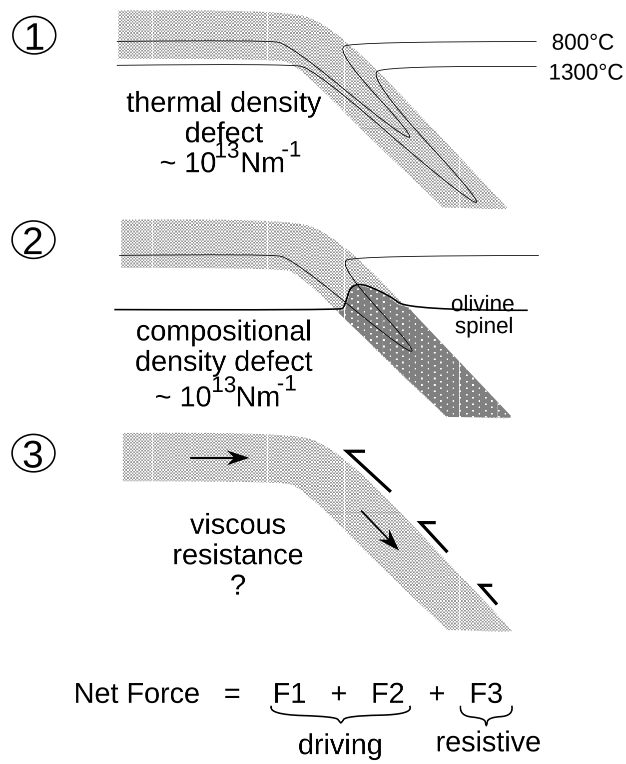

Figure 8.2: Scematic force balance on a subducting.

That the upper parts of subducting slabs are in general descending under the influence of a gravitational body force is indicated by the existence of extensional stresses in the descending slab as witnessed by focal mechanisms of earthquakes generated in the upper 100 - 200 km of the slab. This implies that for most subduction environments the net gravitational body force minus the viscous resistance (which combine to give slab pull) exceed the magnitude of the ridge push) for normal velocities of subduction (typically in the range 0 -10 cm yr^{-1}). The lack of subduction at rates significantly greater than 10 cm yr^{-1} may be understood by the increase in the viscous resistence for increased velocities: the viscous resistance providing a negative feedback on the descent velocity.

8.4 Arc dynamics

Kinematics

From the point of view of lithospheric kinematics, subduction represents one of two mechanisms accommodating convergence between plates; the alternative mechanism being the internal deformation of the plates. The relative importance of subduction and internal deformation can be illlstrated by considering the plate convergence velocity V_{con} relative to subduction velocity V_{sub} as in Figures 8.3 & 4.

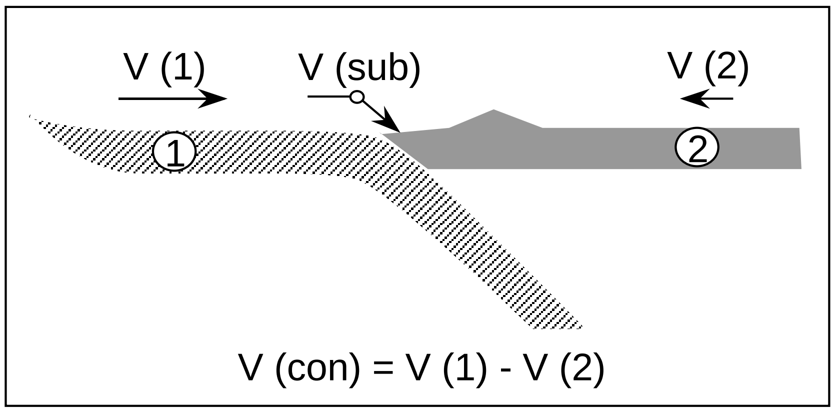

Figure 8.3: Convergence between plates, $V_{\mathrm{con}} = V_2 - V_2$, may be taken up either by subduction V_{sub} or by internal deformation of the plates as illustrated in Fig. 8.4.

Mechanics

While the subduction environment is one of convergence, it may not necessarily involve horizontal compression. We have shown that the driving force for subduction is provided by the negative buoyancy of old oceanic lithosphere which imparts tensional stresses in the descending slab. The state of stress at the trench will depend primarily on the buoyancy of the oceanic lithosphere that is being subducted. We have already commented that it is possible to subduct relatively young buoyant lithosphere. Moreover, there are regions of elevated oceanic lithosphere where the crust is anomalously thick due to, for instance, the accretion of Ocean Island Basalts (OIB). Examples of subduction of anomalously bouyant oceanic lithosphere are provided by the Juan Fernandez and Nazca ridges presently being subducted

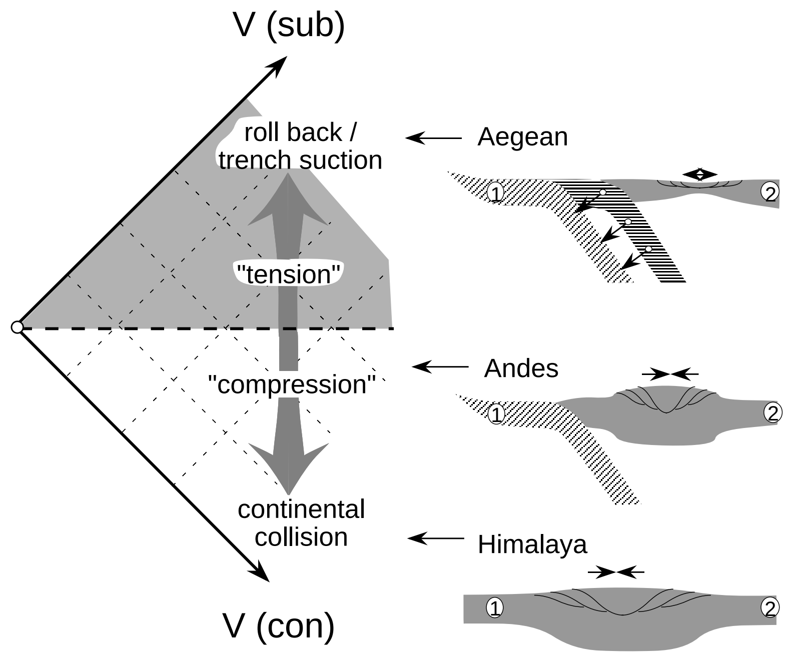

Figure 8.4: Alternative kinematic scenarios for convergent orogens expressed terms of the velocity of subduction relsative to velocity of plate convergence

beneath the amagmatic zones of the South American Andes. Siesmicity from the slab descending beneath these regions shows that this anomalously buoyant oceanic lithosphere is subducting at a much shallower angles than adjacent 'normal' oceanic lithosphere. The state of stress at the trench may therefore be compressional if young or anomalously buoyant lithosphere is being subducted or it may be tensional if old negatively buoyant lithosphere is being subducted.

The force applied at the trench (F_t) provides a boundary condition for deformation in the neibouring arc, which is also subject to buoyancy forces (F_b) arising form the topographic gradients on the arc itself (Chapter 3). The resultant force balance on an arc (F_a) can be formulated as :

$$F_a = F_t - F_b \tag{8.6}$$

where F_t and F_a are positive for compression and F_b is the bouyancy force exerted by the arc on the surrounding lithosphere.

Because of difficulties in formulating F_t we do not provide a complete solution here (one way of treating F_t is in a completely ad hoc fashion by considering it as variable within the bounds of slab pull and ridge push). However, substantial horizontal extensional forces can arise from topography of the magnitude observed in arcs. Extensional failure of an arc gives rise to the development of small ocean basins, termed back-arc basins, between ruptured arc fragments. The foundering of old dense oceanic lithosphere in front of an active arc will lead to a "roll back" effect of the trench, which will greatly enhance the tendency of an arc to spread (Figure 8.4.).

8.5 Arcs and crustal growth

The magmatic island arcs and continental marginal arcs represent the most important sub-aerial terrestrial volcanic provinces. In these settings the visible volcanic activity is volumetrically minor in comparison with contemporary intrusive activity. Compositionally the characteristic lavas of these terrains are andesitic (or quartz diorite, tonalite, granodiorite or quartz monzonite intrusives), though basalts (gabbros) are also common. These zones of convergence between either continental - (e.g. the Andes) or oceanic lithosphere (island arcs) and oceanic lithospheric plates, are sites which are probably critical to the evolution of the continental crust (and subcrustal lithosphere). The "andesite" model of Taylor (1977) embodies this concept, suggesting that the average composition of at least the upper continental crust is andesitic in bulk geochemical terms and that it aquired this composition because the main mode of continental growth has been by marginal accretion of magmatic arcs.

As we have seen in the previous chapters, the oceanic crust has been developed by extensive and repeated fusion of the asthenosphere over a substantial portion of the Earth's history (at least the past 2 Ga). This source of MORB has very large-scale depletions of lithophile (incompatible) trace- and minor elements in comparison with the pristine or primitive (chondritic) mantle. The low $^{87}Sr/^{86}Sr$, $^{207}Pb/^{204}Pb$, $^{206}Pb/^{204}Pb$ and high $^{143}Nd/^{144}Nd$ values of MORB imply that its source region has had low Rb/Sr, Nd/Sm and U/Pb ratios for substantial periods of Earth history. The fact that the continental crust has complementary isotopic and geochemical characteristics with those of MORB suggests that the crust represents the "missing-part" of this asthenospheric source of present-day MORB. The question of what role arcs play in mediating this transition of ocenaic crust to continental crust is fundamental to understanding crustal growth.

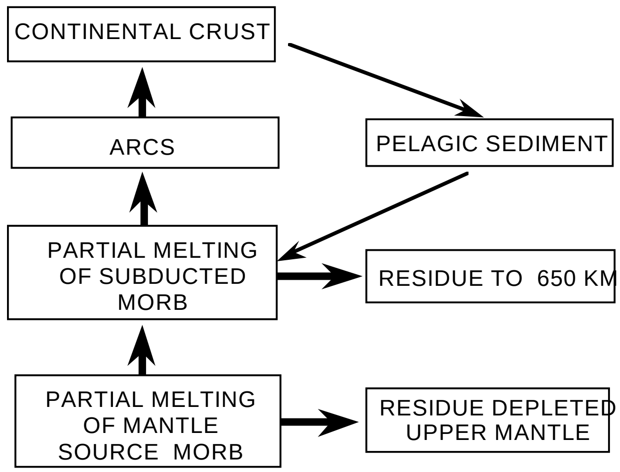

Figure 8.5: A possible relationship between MORB and continental growth. If the important events which influence the complementary geochemical /isotopic evolution of the sub-oceanic upper mantle and the continental crust do in deed take place at plate boundaries (and there is some argument against this), then the critical boundary must be the zones of lithospheric plate convergence (subduction zones). It is clear that although the primary extraction of material from the asthenosphere takes place at the ridges the basalt produced here does not resemble the continental crust , the upper portion of which is certainly much more felsic. The following section will discuss how the quite special circumstances that exist in subduction zones combine to produce and emplace relatively felsic magmas in the crust , and why these magmas probably contain a large part of the incompatible element content of the oceanic crust (and hence of the sub-oceanic upper mantle).

The sources of arc magmas

One complicating factor in the study of arc magmatism is the question of the source of the magmas. It is easy to show that the covariation of isotope and geochemical data for many suites of arc volcanic rocks must be due to mixing of components from more than one source and in fact many studies can demonstrate up to four or five seperate components. The possible source components which may be available include:

1. The basaltic (crustal) portion of the subducting slab.

2. The pelagic sediments which may be carried with the subducting slab.

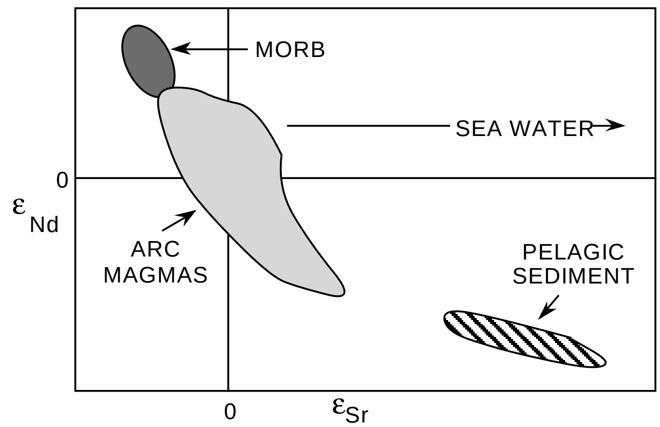

Figure 8.6: Nd and Sr isotopic composition of arc magmas and their potential sources.

3. The depleted peridotite wedge above the subducting slab.

4. The lower crust of the arc itself.

5. An enriched (or none-depleted) mantle source (like the OIB source).

Added to this list is the important fact that the oceanic crust undergoing subduction is not pristine, but rather has undergone very extensive alteration, hydration and metamorphism since the time of its formation at the ridge. The upper portion of the subducting slab will therefore undergo dehydration at depth generating fluids which themselves carry considerable concentrations of solutes. These will escape from the subduction zone and their role in fluxing the overlying mantle will have a pivotal role in the generation of arc magmas. The hydrothermal alteration is promoted by the initially high thermal gradient at the ridge crest which promotes circulation of sea water in fracture systems that might extend through the thickness of the whole crust. The oceanic crust is geochemically and mineralogically modified by exchange with sea-water which changes (amongst other things) its O- and Sr-isotopic composition towards those of sea water. The primary igneous minerals and glass are altered to hydrates including talc, actinolite-tremolite , epidote ,chlorite, serpentine and brucite and these minerals carry their bound water to be released at depth from the subducting slab.

The thermal structure of arcs

The heat-flow along an across arc transect from the oceanwards-side of the trench to the back-arc region varies considerably. There is low heat flow above the old oceanic crust, extremely low heat-flow from the trench and significantly higher flow from the arc and backarc regions. The extremely low heat flow at the trenches is an expression of the persistent coldness of the the downgoing slab to depth. This has the importance of delaying the onset of those dehydration reactions that are mainly temperature dependent to much greater depths than under the regime of normal geothermal gradients. Particularly significant amongst these is the the reaction (Figure 8.5):

amphibole $->$ Na - clinopyroxene (omphacite) + garnet + H_2O

which defines the amphibolite - eclogite transition and at pressures less than 25-30 kbars. this takes place at close to 1000^{o}C. Above 25 kbars the reaction is much more pressure dependant (Figure 8.5).

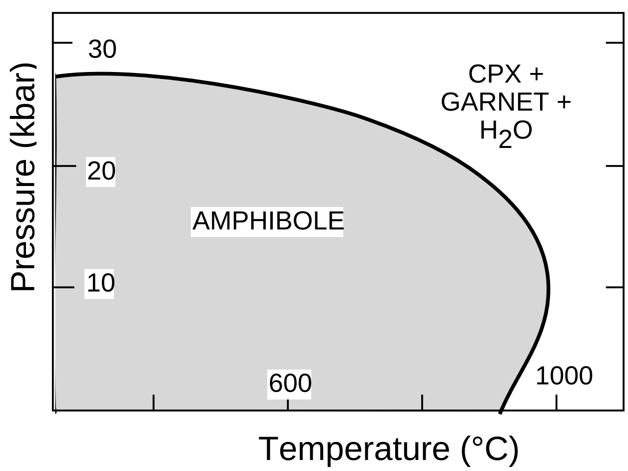

Figure 8.7: The stability field of the hydrate amphibole as a function of temperature and pressure.

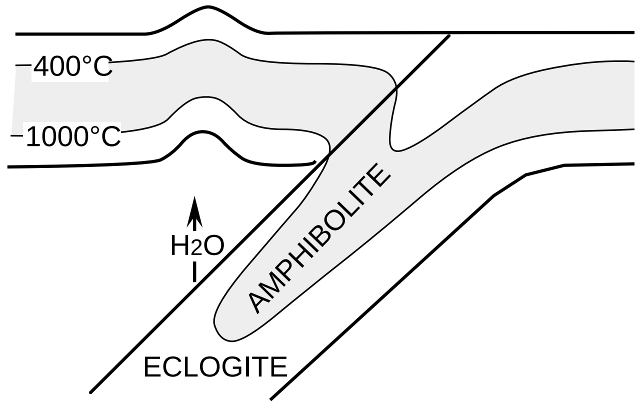

Figure 8.8: A simplified view of the thermal structure of the subducting slab and the sub-arc mantle. The stippled region illustrates the stability field of amphibole. The break- down of amphibole when the the slab reaches about 100 km releases H_2O and produces the dense rock eclogite consisting of garnet and omphacite ($\rho = 3600$ kg m^{-3}).

Melting in the Subducting Slab

The crustal part of the downgoing slab has the potential capacity to produce felsic melts with quite high lithophile element contents, high $^{87}Sr/^{86}Sr$, $^{207}Pb/^{204}Pb$, $^{206}Pb/^{204}Pb$ and low $^{143}Nd/^{144}Nd$. This is partly due to the pelagic sediment component and partly to the hydrothermal alteration of the basaltic oceanic crust. A popular theory when subduction was first recognised was that arcs are dominated by intermediate (andesitic) magmas because they were derived from melting of mafic sources rather than the ultra-mafic sources that generate basaltic magmas, i.e. the upper part of the down-going slab. However, the common presence of basaltic magmas in arcs implies that the slab is not the major source of these melts . Rather it seems more likely that the wedge of peridotite mantle above the slab is the source of the bulk of the arc magmas and this melts at relatively low temperatures because of the fluxing by water -rich fluids driven off the slab. The mantle wedge is also "enriched" with respect to normal mantle peridotite in the lithophile (incompatible) element component derived from the slab. The hydrous, mafic, primary magmas thus produced then undergo variable fractionation en-route to the upper part of the crust to produce andesite, dacite and rhyolite.

The fate of the slab after it has suffered the extraction of small volumes of melts and fluids is of interest. Ringwood (1974) has determined the density variation of the main earth materials in the subduction system. As illustrated in Figure 8.5, the subduction zone can be considered as a two layerd slab composed of MORB crust and depleted peridotite mantle lithosphere (harzburgite) which is being thrust into asthenospheric mantle. Under conditions of thermal equilibrium, at pressures equivalent to depths up to 650 km, MORB is more dense than both asthenospheric and harzburgititic (lithospheric) peridotite, but is less dense than both types of peridotite in the depth interval 650- 720 km. Depleted, harzburgitic peridotite is less dense than less refractory peridotite (asthenosphere) at all depths except in the 650 - 720 km interval. On this basis, Ringwood

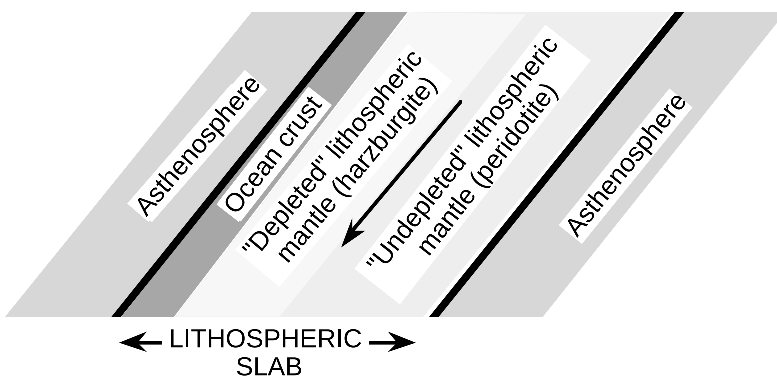

Figure 8.9: Compositional structure of the descending slab.

(1974) suggests that the subducting slab, reaches a neutral buoyancy level at about 650 km where it is preserved as a layer composed of harzburgite with remnant blobs of the basaltic (MORB) portion of the slab. He further suggests that this accumulated layer would provide a physical barrier to promote the development and maintenance of a two - layered convection system in the mantle.

The role of water in arc magmatism

Allegre et al (1987) calculate that over the past 4 Ga. subduction has recycled about 4 times the mass of the present oceans back into the mantle together with considerable amounts of the atmospheric rare gas contents. It is clear that the mantle residence time of this massive flux is essentially zero and that subduction zone volcanism is the obvious site of outgassing of this system. The direct evidence for the role of water in arc magmatism comes from many observations. These include:

1. Direct discharge of water vapour (with other fugitive species) during volcanic eruption, leading to explosive and pyroclastic volcanism.

2. High plagioclase content of intermediate magmas resulting from its latent enrichment during fractionation of hydrous magmas. Water delaying the normal appearance of plagioclase to lower temperatures.

3. The expansion of olivine crystallisation to relatively silicic magmas.

4. The occurence of hydrous minerals such as amphibole and biotite.

5. The discharge of magmatic hydrous fluids from decompressing and crystallising magmas at upper crustal pressures to contribute to hydrothermal alteration halos around some plutons (porphyry copper deposits).

6. The delayed, late, or none crystallisation of quartz in many quatrz-normative magma systems.

7. The complex oscilliatory zonation/ resorbtion patterns of many phenocrysts in arc magmas suggesting growth variable P_{H_2O} regimes in crustal magma chambers.

8. Vapor-bearing fluid inclusions in many phenocrysts.

It is probable that melting in the peridotite wedge above the subducting slab takes place under conditions like those of the water undersaturated solidus in Figure 8.5. Melting is promoted by the high activity of water from the dehydration of the slab and the initiation of melting will promote the rise of the geotherms in the strip of asthenosphere above the main locus of dehydration in the slab. This in turn will contribute to the rise of the asthenosphere beneath the

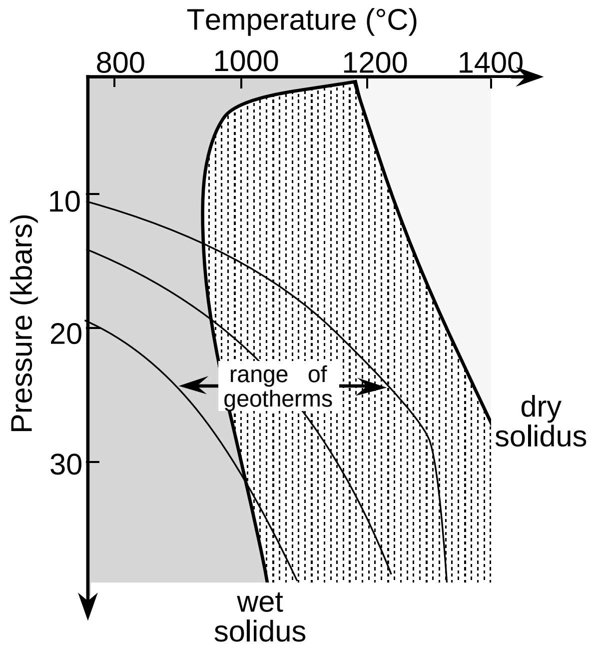

Figure 8.10: Peridotite solidi for dry, H_2O-undersaturated and H_2O -saturated (wet) conditions. The stippled area indicates the possible range of geothermal gradients.

arc, leading to thinning of the mantle lithosphere beneath arc and further decompressive melting in the asthenosphere. At this stage the situation is not unlike that at the mid ocean ridges and arc splitting and spreading may take place. The combination of the effects of the removal of the mantle lithosphere from beneath the arc and the addition of large volumes magmatic rocks to the crust will tend to depress the base of the arc's crust through rising geotherms, to the effect that the the solidus of mafic rocks may be intersected at lower crustal depths. This will lead to short-timescale magmatic recycling within the arc itself.