PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Thermal Processes in the Formation of Continental Lithosphere [and Discussion]

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Phil. Trans. R. Soc. Lond. A 1978 **288**, 415-429
doi: 10.1098/rsta.1978.0025

Phil. Trans. R. Soc. Lond. A. 288, 415–429 (1978) [415]
Printed in Great Britain

Thermal processes in the formation of continental lithosphere

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The thermal evolution of continental lithosphere in atectonic regions has been interpreted in terms of (1) conductive cooling, in the same way as oceanic lithosphere, but over much longer periods; (2) conductive cooling accelerated by erosion; (3) erosional removal of near-surface concentrations of heat-producing elements; and (4) various special temperature conditions assumed for its base. Although all of these factors influence lithospheric temperatures, particularly early in the development of continents, for times greater than 10^9 a, the thickness of the lithosphere and the processes by which it forms are of overriding importance.

Continental lithosphere may develop by cooling and the *thermal accretion* of mantle material which has not been depleted of a basaltic first melting fraction; or it may develop by *diapiric accretion* of low-density, depleted mantle bodies rising from the upper parts of lithospheric slabs heated during their descent in subduction zones. The former process alone could not generate continental lithosphere with the observed characteristics. The latter process is likely to be important, possibly in combination with the former.

1. Introduction

The proper understanding of a wide range of tectonic and thermal phenomena within the continents depends upon a knowledge of the 'normal' continental temperature distribution and how it may have changed with time. Superimposed on this background thermal régime are a variety of thermal disturbances which may be regarded as controlled by extraneous factors, e.g. the magmatic activity associated with the rifting of an ancient continent such as the separation of Arabia from Africa by the Red Sea rift; or the fusion of crustal rocks at shallow depths in Skye to generate granites by heating through the extrusion of basalts from the mantle; or the tectonic burial of highly radioactive near-surface crustal rocks by overthrusting in regions of continental collision such as the Alps.

It is our purpose in this paper to review the factors which are likely to influence the continental temperature distributions upon which perturbations such as these are superimposed. For continental crust more than 500 Ma old, the overriding considerations are the thickness and mode of origin of the continental lithosphere, i.e. the thickness of the zone of mantle which is effectively coupled to the continental crust and moves with it with respect to the surrounding mantle material; the latter we call 'circulating mantle' because it must be able to flow around and under plates to conserve the overall figure of the Earth against the effects of plate motions.

2. Radioactivity, erosion and cooling

Thermal gradients within the continental crust change continuously and are never entirely steady. It is nevertheless convenient to begin by regarding them as if they were. Within any thickness of crust or lithosphere of interest, the steady state temperature distribution depends on three factors: the temperature at its base, the distribution of radioactive heat producing elements within it, and its thermal conductivity structure. In the absence of heat producing elements, and with uniform thermal conductivity, the surface heat flux, q, is given by $q = k\beta$, where k is the thermal conductivity and β is the thermal gradient; the temperature, T, at the depth of interest, z, is given by

$$T = q\,z/k + T_s,$$

where T_s is the surface temperature. If the depth interval contains radioactivity, an additional term must be added but its form will depend on the distribution of the radioactivity. In general, if the radioactivity is concentrated near the surface it will have least effect on T, whereas if it is concentrated near the base it will have its largest effect on T.

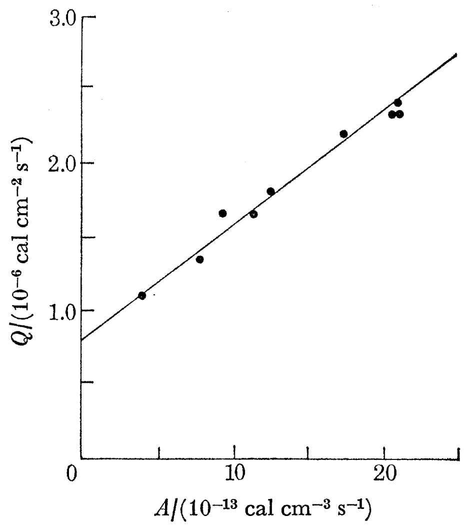

FIGURE 1. Heat production, A, in the near surface rocks plotted against surface heat flow, Q, for a number of New England plutons (Roy, Blackwell & Birch 1968). The slope of the line, h, is about 8.5 km suggesting that in this and similar continental areas most of the radioactivity occurs at depths less than 10 km.

Lambert & Heier (1968) showed on geochemical evidence that radioactivity was concentrated upwards in continental crust, and the work of Birch, Roy & Decker (1968) and Lachenbruch (1968) established the somewhat surprising generalization that in igneous and meta-igneous areas there is a correlation between the value of surface heat flow and the near-surface concentration of heat producing elements (figure 1). This led to the recognition of two distinct components to surface heat flow: one associated with near surface crustal radioactivity and the other a conductive component from the lower crust and mantle, q_m:

$$q = q_m + A_0 h,$$

where A_0 is the near surface concentration of radioactivity and h is a parameter with dimensions of length and is given by the slope of the line in figure 1. Lachenbruch pointed out that although numerous vertical distributions of radioactivity with depth in the crust would satisfy this relation, one which was both geochemically plausible and which would not be destroyed by erosion was a concentration of radioactivity which decreased exponentially with depth, i.e.

$$A_z = A_0 \exp(-z/h),$$

where z is depth below the surface; h is thus the exponential e-folding length.

Erosion at the Earth's surface has two thermal consequences: first, rocks which were previously some distance below the surface and at a higher temperature are brought to surface temperature, and the thermal gradient in all the underlying units must adjust to this new condition. The extent to which this re-equilibration occurs depends on the rate of downward movement of the erosion surface. If the latter is fast there will be a transient enhancement of the surface heat flow. Secondly, erosion also affects the lateral distribution of heat producing elements; the equilibrium value of the heat flow in the eroded area will fall and the relatively radiogenic sedimentary debris may be deposited in a sedimentary basin where the net local crustal heat production will in consequence be somewhat increased. Erosion, therefore, may affect the equilibrium temperature at the base of the crust in the areas both of erosion and of deposition.

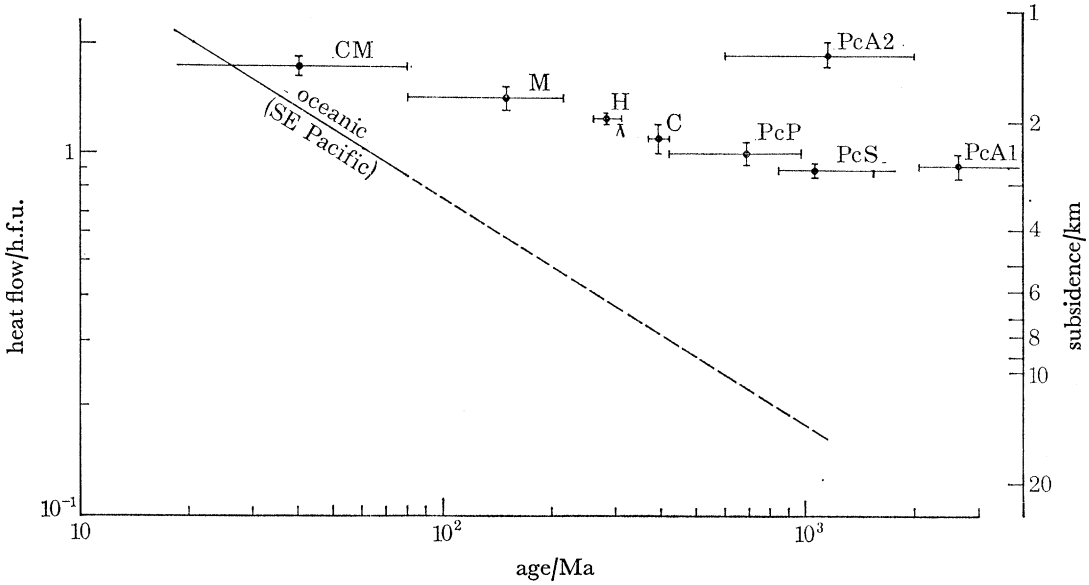

FIGURE 2. Heat flow and subsidence as a function of age. The points are for mean continental heat flow values in provinces of different ages (see below for letters). The solid line gives the observed dependence of heat flow (left scale) and subsidence (right scale) on age in the ocean basins. The dashed continuation gives some idea of the subsidence to be expected in old continental areas if continental lithosphere thickened by cooling over very long periods: CM, Cenozoic miogeosynclines; M, Mesozoic folding; H, Hercynian orogeny; C, Caledonian orogeny; PcP, Precambrian platforms; PcS, Precambrian shields; PcA2, Precambrian central Australia; PcA1, Precambrian western Australia. (From Sclater & Francheteau 1970; Sass *et al.* 1976.)

In 1968 Polyak & Smirnov presented data to show that in continental areas there was a broad correlation between the present surface heat flux and the age of the last major thermal event which affected the area (figure 2). The gross validity of this generalization was substantiated by Sclater & Francheteau (1970) with additional observations, but there is a very wide scatter of heat flow values about the mean. Although this may in part reflect errors in the heat flow observations or differences in the methods of reducing them, it is likely also to result from the poor definition of what should be regarded as the 'last major thermal event'. Future work will need to examine this concept more closely.

In so far as it had been established that radioactivity tended to be concentrated upwards in the crust, the most immediate interpretation of the age dependence of continental heat flow was that old regions showed low values because they had been most deeply eroded. This approach was thoroughly explored by Richardson (1975), who showed that an exponentially decreasing erosion rate (with a characteristic time of about 270 Ma) could account for much of the observed decrease in surface heat flow with age during the first 500 Ma of evolution of an area of continental crust.

Alternatively the age–heat flow relation on the continents may be interpreted in broadly the same way as that in the oceans, namely as a decrease in heat flow with time resulting from conductive cooling of the crust and upper mantle after the last major thermal event to have affected it (Crough & Thompson 1976; Pollack & Chapman 1977). This implies a progressive decrease in q_m with time. There are insufficient data for very old or very young areas to test this hypothesis adequately but there is some suggestion that q_m may be high in recently active areas (e.g. Basin & Range 1.4 h.f.u.) dropping to a value of 0.8 h.f.u. after 400 Ma. It is not clear whether q_m stays more or less constant at this value or slowly decreases with greater age, but Sass, Jaeger & Munroe (1976) quote two values close to 0.6 h.f.u. for the Australian Precambrian.

There are two main difficulties with the formation of continental lithosphere by conductive cooling of crust and upper mantle of initially uniform temperature. First, this should give the same $t^{-\frac{1}{2}}$ dependence of heat flow on time, t, as is observed in the oceans. Figure 2 shows both the continental values and for comparison the $t^{-\frac{1}{2}}$ curve for the Pacific. There clearly is a significant difference in slope (Sclater & Francheteau 1970).

A problem also arises from the very large amount of thermal contraction which would be expected in lithosphere which cooled continuously for periods of the order of 3 Ga. Extrapolation of the oceanic empirical age/subsidence curve gives 15 km of thermal subsidence for continental areas 1 Ga old (figure 2). This subsidence is unacceptably large, and takes no account of two other factors which tend to cancel each other: the effect of the radioactivity of mantle material incorporated in the thickening continental lithosphere, and which slows its cooling; and on the other hand the effects of erosion which, although partially offset by isostatic uplift, should add a further 4–5 km of subsidence.

In practice both erosional redistribution of near surface radioactivity and conductive cooling must influence continental heat flow, but additional factors are involved. In the discussion which follows we are primarily concerned with the factors which may influence q_m, and how it may have varied with time.

At this point it is necessary to view the thermal gradients of the continental crust within the wider context of the large scale motions implied by plate tectonics. Except in regions of active igneous intrusion or major tectonic movement, we expect temperature distributions within the lithosphere as defined earlier, to be governed largely by thermal conduction. Beneath the lithosphere we assume that the mantle behaves as a fluid on time-scales of interest, and that the tendency of the motions within it is to equalize temperatures within the zone they affect. We take thermal gradients within this zone of circulating mantle to be close to adiabatic. An adiabatic temperature distribution except in regions of localized ascending and descending flow is consistent with the results of thermal convection studies (e.g. Turcotte & Oxburgh 1967, 1969; McKenzie, Roberts & Weiss 1974), and with estimates of temperatures in the mantle based on the variation of electrical conductivity with depth (Tozer 1959).

In such a model the value of q_m is largely controlled by the thickness of the lithosphere and radiogenic heat production within it; very low values of q_m are possible only if the lower boundary of the lithosphere, which is maintained at the temperature of the circulating mantle, is well below the base of the crust so that the conductive thermal gradient through the lower lithosphere is very low and thus sustains a low conductive heat flux. Conversely, in the absence of other effects and if the radiogenic heat production were known, variation in q_m could be used to estimate variations in the thickness of the lithosphere (Pollack & Chapman 1977). Crough & Thompson (1976) consider the effect of heat supplied to the base of the lithosphere by the circulating mantle but do not include the heat production in the mantle part of the lithosphere. Their mathematical treatment of transient aspects of this problem involves assumptions which we consider unjustified.

It is therefore important to see what constraints may be placed on the thickness of the continental lithosphere. Unfortunately these are rather few. Sipkin & Jordan (1976), however, extending earlier work, propose that differences in upper mantle shear wave velocity between continental and oceanic areas extend down to 400 km or deeper. Systematic relations of this kind could obtain only if this were the thickness of the continental lithosphere. The authors also suggest that there are discernable differences between mantle underlying older and younger parts of continents, both showing higher velocities (i.e. shorter travel times) than sub-oceanic mantle at the same depth, but the older showing a larger difference than the younger. Jordan (1975) pointed out that these differences were too large to be explicable in terms of temperature differences and therefore implied differences in chemical composition.

3. Generation of continental lithosphere by diapiric accretion

It has been pointed out by several authors (O'Hara 1975; Boyd & McCallister 1976; Oxburgh & Turcotte 1976; Oxburgh & Parmentier 1977) that mantle material which has undergone partial melting and been depleted of its first melt fraction in the production of basalt, is less dense than before the depletion occurred. This is partly because the dense phase, garnet, makes a major contribution to the liquid, and partly because iron is preferentially fractionated into the liquid during the melting of the pyroxenes and olivine. Depleted mantle† is between 0.06 and 0.09 g/cm^3 less dense than undepleted mantle at the same temperature. This difference persists through the olivine–spinel phase transition, the pyroxenes in both depleted and undepleted mantle inverting to a garnet structure (Ringwood 1975).

It has been shown (Oxburgh 1965) that the partial melting which occurs under mid-ocean ridges and allows formation of the basaltic oceanic crust, should generate a thick zone of depleted mantle immediately below the M discontinuity (Gass, Smith & Vine 1975, fig. 1). The thickness of the depleted zone depends on the degree of depletion; the total amount of depletion, however, must be sufficient to provide the volume of the oceanic crust. Oxburgh & Parmentier (1977) take 20 km as the thickness of the depleted zone; in reality the base is likely to be gradational into underlying undepleted mantle. The mantle part of the oceanic lithosphere is thus compositionally stratified.

During subduction cold oceanic lithosphere is heated by friction along its contact with the overlying mantle and, neglecting the thin oceanic crust, the first part of the slab to be heated is its capping of depleted mantle. Because this is intrinsically less dense than the surrounding undepleted mantle, it becomes buoyant as it is heated. A possible consequence is that as soon as a sufficiently great thickness is heated and has become ductile, diapiric masses of depleted

† It is strictly inadequate to speak simply of 'depleted' and 'undepleted' mantle. The degree and character of depletion must necessarily vary. Nevertheless the term is retained here to signify mantle material which has undergone about 20% partial melting.

mantle separate in the solid state, and penetrate the mantle overlying the slab (Oxburgh & Parmentier 1977).

If subduction is occurring close to a continental margin, diapirs may be expected to accumulate beneath the continental crust and possibly to build up a subcontinental lithosphere which differs in composition from that of the 'average circulating mantle'. If continental crust is itself generated in such areas by calc-alkaline magmas derived from subducted oceanic crust, the process outlined above would permit the nearly simultaneous generation of a subscrustal lithosphere.

It is possible to imagine such a process of diapiric accretion of the lithosphere occurring in two different ways depending upon the depth at which the diapirs separated from the top of the slab; in the first they would separate at a relatively shallow depth, say 400 km or so, and would build up the lithosphere close to the continental margin. In the second, the separation might occur only when the slab was thoroughly heated and had perhaps travelled a thousand or more kilometres under the continent. In this case diapirs could build up the continental lithosphere at long distances from its margin. As we have pointed out elsewhere (Oxburgh & Parmentier 1977), the depleted zone within young oceanic lithosphere will tend to become buoyant quickly and that within older lithosphere more slowly.

The depth at which depleted material would separate from the slab is difficult to predict since it will depend on factors such as the flow geometry, the temperature dependence of viscosity and the variation of the degree of depletion with depth within the subducted slab. The fate of any depleted material carried with the slab into the deeper parts of the upper mantle is uncertain. If this material is entrained into the global mantle circulation, then it may be mixed with less depleted mantle. Mixing requires the smearing-out of compositional inhomogeneities by shearing deformation in the mantle flow. The shearing will tend to reduce the size of the inhomogeneities, but mixing on a very small scale must ultimately involve slow diffusion processes. Isotopic evidence from oceanic basalts can be interpreted as indicating that mixing at this scale is at best incomplete (Brooks, Hart, Hofman & James 1976). However, the smaller the inhomogeneities become, the less important would be their buoyancy compared to viscous forces. Therefore, depending on the form of mantle flow, mixing may be fast enough to prevent upward segregation of depleted material at significant rates.

As a first step in determining the depth at which depleted material is likely to separate from the slab, we examine the dependence of the buoyancy of the depleted zone on lithosphere age and depth of subduction. Taking the dependence of temperature on depth and age for lithosphere entering a subduction zone discussed previously (Oxburgh & Parmentier 1977) and assuming that the subducting lithosphere is heated frictionally along the slip zone with the overlying mantle according to the model of Turcotte & Schubert (1973), we find that the top of the depleted zone first becomes buoyant at about 100 km depth while the depleted zone as a whole, first becomes buoyant at between 200 and 250 km. These results are illustrated in figure 3; we assume a subduction rate of 8 cm/a; other parameter values are as taken by Turcotte & Schubert. Neutral buoyancy is achieved by the depleted zone at temperatures much below the ambient temperature for the mantle at that depth and buyoancy will increase as the slab as a whole continues to warm. Diapirs are expected to separate somewhere in this interval. We examine this aspect of the problem with figure 4; the curve for $B = 0$ indicates the depth at which the depleted zone as a whole achieves neutral buoyancy, while that for $B = 1$ corresponds to the buoyancy when all depleted material is at ambient mantle temperature. Considering the present day subduction along the western coast of South America, the oceanic lithosphere entering the Peru–N Chile trench is about 50 Ma old; the corresponding belt of high Andean topography is about 400 km wide (23° S), and at its centre the depth to the subduction zone is about 200 km. If this zone of high regional topography were a reflection of the width of the subcrustal zone affected by diapirism it would suggest that diapirs separated at a value of B close to 0.5. In this case, nearly all diapirism should be close to continental margins.

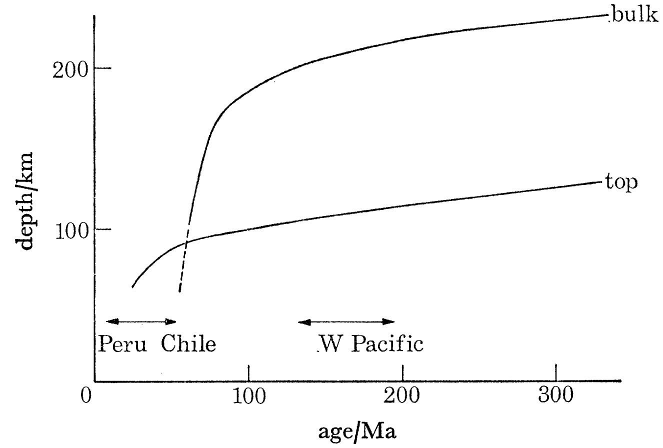

FIGURE 3. Depths below the surface at which the top of the depleted zone and the depleted zone as a whole (bulk) achieve neutral buoyancy, as a function of slab age. Age ranges are indicated for several present day subducting slabs.

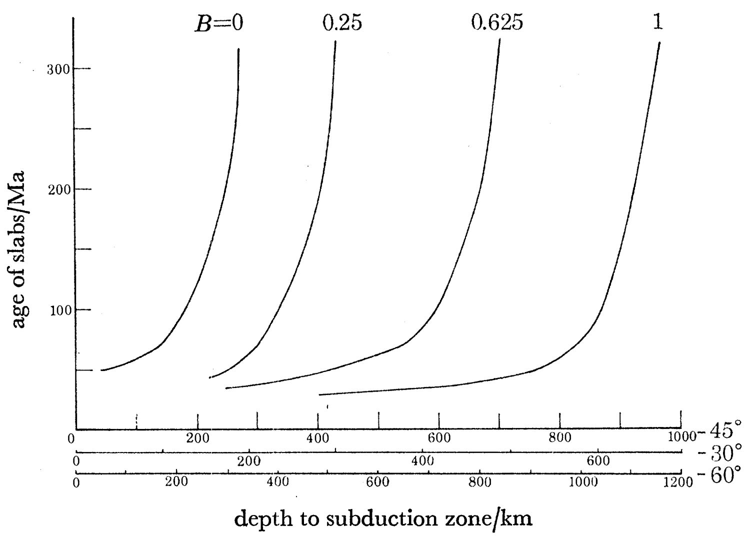

FIGURE 4. Bulk depleted zone buoyancy as a function of slab age and vertical depth to the slip zone for three different mean angles of subduction. See text for discussion.

4. THE CONSEQUENCES OF DIAPIRIC ACCRETION OF CONTINENTAL LITHOSPHERE

If important contributions to the continental lithosphere are made by depleted diapiric masses, there are consequences for a wide range of geophysical, geochemical and tectonic processes. We now draw attention to a number of them.

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(*a*) *Lithospheric temperatures*

The diapiric accretion of depleted mantle material to the base of newly formed continental crust could lead to a relatively rapid growth of continental lithosphere, and it is possible that the thickness of continental lithosphere is largely determined by this process at the time of its initial formation. This suggests the possibility of continental lithosphere which varies in thickness according to the duration and nature of the subduction processes at the time of its formation.

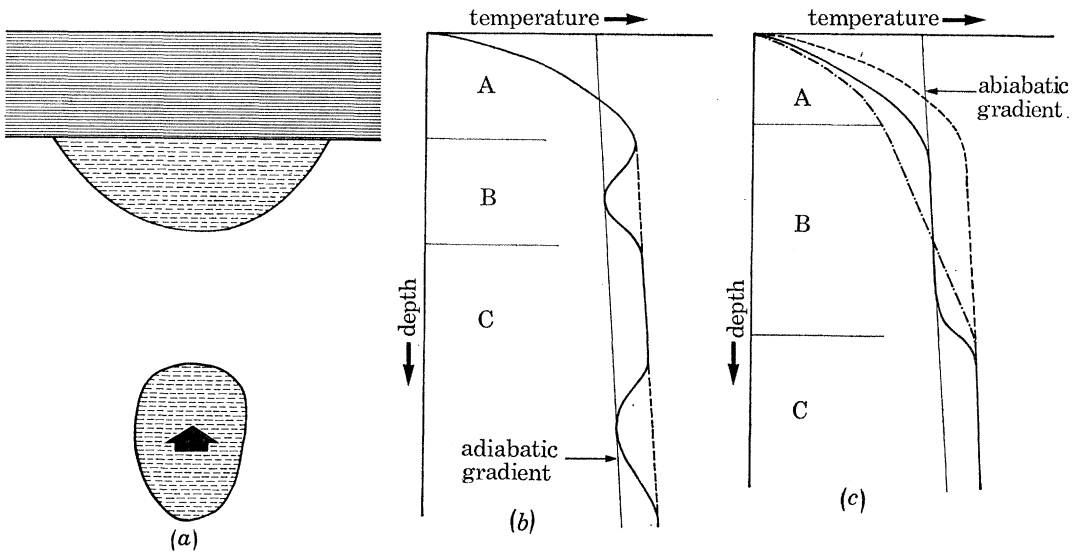

Figure 5. (*a*) A thin conductive lithosphere formed by cooling from above (ruled) has received one diapiric mass from below and another is *en route* (both dashed). (*b*) The diapirs have risen while still cooler than ambient mantle and have followed an adiabat during ascent; the solid line gives the temperature along the centre line of (*a*) shortly after the arrival of the upper diapiric mass; pre-arrival temperature dashed line. (*c*) Solid curve gives the temperature after the arrival of a number of diapirs (making up zone B); after diapirism ceases there is conductive relaxation to the dot-dashed profile. Dashed profile as in (*b*).

If the accreting diapirs rise before they have reached ambient mantle temperature, the lithosphere which they build will initially be cooler than the mantle material which circulates around it, and through which it moves during plate motion. This situation is shown schematically in figure 5*b*. The dashed curve shows the temperature before accretion began; a thin conductive upper zone (A) is underlain by a circulating zone within which the gradient is adiabatic. An episode of accretion adds the thickness of zone (B), and the resultant temperature distribution is given by the solid curve. The subsequent thermal history involves the addition of further diapirs and finally the heating of the depleted material from below and cooling from above (dot–dashed curve) to give conductive equilibrium across zones A and B (figure 5*c*).

This solid state diapiric accretion of lithosphere with distinctive chemical characteristics and its subsequent thermal equilibration, is in marked contrast to the view of continental lithosphere as a cold and therefore strong thermal boundary layer which thickens with time by the cooling at its upper surface, i.e. growth by thermal accretion. Such cooling must occur, but the question is whether the continental lithosphere is predominantly compositionally controlled, or thermally controlled (as the oceanic lithosphere) or some combination of both. We have not specified a thickness to the depleted material added beneath a continent; if on average it were relatively thin, say 180–200 km, it is possible that in very old areas cooling could extend through the full thickness of the depleted zone and thermal accretion of undepleted material could occur at its base. The force of the argument referred to earlier, that very deep cooling under continents leads to unacceptably large amounts of thermal contraction and therefore topographic subsidence, is in this case much reduced; the topographic subsidence is countered by the presence of low density depleted mantle and its special thermal evolution, i.e. warming and cooling, as described above (figure 5).

Regardless of whether undepleted mantle is added to the base of the lithosphere in this way, some undepleted mantle is very likely to be trapped between diapirs and thereby incorporated into the continental lithosphere. We view the continental lithosphere as probably inhomogeneous on a large scale and comprising zones of depleted and undepleted mantle.

(*b*) *Thickness of the continental lithosphere*

Neither the process of diapiric nor thermal accretion can be used to make any quantitative estimate of lithospheric thickness. If all of a 20 km thick depleted zone from the oceanic lithosphere were added to the base of the continents throughout geological time, the thickness of the depleted lithosphere beneath continents should be greater than 1000 km. This may be estimated in a variety of ways, but any set of reasonable assumptions gives this kind of result.† As discussed below, such a thickness is unreasonably large and indicates that if diapiric accretion occurs only a fraction of the volume of mantle material depleted during the production of oceanic crust can be incorporated in the subcontinental lithosphere, or that present subduction rates are substantially higher than the average during the last 3 Ga.

Estimates of lithosphere thickness derived from inferred values of q_{m} (Pollack & Chapman 1977) are subject to great uncertainty. Not only are they very sensitive to the thermal condition assumed at the base of the lithosphere, but to uncertainty in the value of thermal conductivity of ultramafic rocks at very high pressures. Recent experimental work by Cull (1975) up to 20 kbar (2 GPa) suggests that the pressure derivative of thermal conductivity in silicates may previously have been seriously underestimated. These uncertainties are directly expressed as large errors in estimates of rates and depths of cooling. Taking near surface values of thermal diffusivity, which are probably too low, significant conductive cooling could have occurred to a depth of 350 km or so during 4 Ga.

In so far as the continental lithosphere is defined as the layer of subcontinental upper mantle which moves as a coherent mechanical unit with the velocity of the surface plate, its thickness is determined by mechanical factors. The magnitude of the shear stress below the lithosphere, along with the rheology of mantle material, determine the temperature at which the mantle is sufficiently rigid to behave as part of the lithosphere. It is therefore reasonable to regard the temperature at the base of the lithosphere as being a function of pressure, shear stress, and degree of depletion. The rôle of depletion is clear if the viscosity is assumed to be a function of the homologous temperature (defined as the ratio of temperature to absolute melting temperature). Depletion increases the ratio Mg:Fe of the mantle olivine, thereby raising the

† For instance, taking the present length of world subduction zones as 50000 km, an average subduction rate of 7 cm/a over 3 Ga, and a depleted zone 20 km thick, derive a total volume of depleted mantle and divide by the area of the continents to give a 1200 km depleted layer under continents; *or* assume that continental crust is prograding laterally above subduction zones at 1 mm/a (see Oxburgh & Turcotte 1970) and that 7 cm × 20 km/a of depleted mantle is available for diapiric accretion beneath it to give a 1400 km thick depleted lithosphere.

melting temperature and increasing the viscosity for a given temperature. The dependence of the temperature at the base of the lithosphere on stress is important because it implies that the thickness of the lithosphere can change as plate motions and the pattern of mantle circulation change.

Froidevaux & Schubert (1975) and Froidevaux & Souriau (1978, this volume) also define the thickness of the lithosphere on the basis of stress and rheology. However, we take the temperature of the circulating subcontinental mantle as nearly adiabatic rather than as being determined by conductive cooling through the continental lithosphere. The adiabatic approximation is more consistent with the long times required for conductive cooling through a thick continental lithosphere and with ideas of mantle-wide circulation that changes on time scales of a few hundred million years or less (Parmentier & Oliver 1977).

(c) *Petrological constraints on the continental lithosphere*

As a number of other papers in this volume show, the uncertainties in estimating the temperatures and pressures at which particular mineral parageneses were in equilibrium are very great. Nevertheless, studies of xenoliths carried to the surface by kimberlites have suggested that they represent a random sampling of the rocks of the upper 200 km or so of the lithosphere (e.g. Dawson 1971; McGetchin 1972), and Boyd (1973) has proposed a thermal structure for the sub-continental upper mantle on the basis of a series of temperature/pressure values derived from their mineral assemblages. Although the absolute values may require revision (e.g. Mercier & Carter 1975), it is nevertheless likely that the depth order proposed for the nodules is substantially correct. This indicates that the shallower nodules are typically depleted peridotite; they have an essentially granular texture, and they appear to be underlain by undepleted garnet lherzolite with a sheared texture. Boyd & McCallister (1976) comment: 'This relationship involving light rocks overlying denser rocks now appears to be general for the uppermost part of the upper mantle under southern Africa.' Such a model is similar to that inferred by Dawson (1971, fig. 4). Although the correlation of nodule texture with inferred depth of origin seems to hold generally in southern Africa, there are exceptions (e.g. Dawson, Gurney & Lawless 1975; Harte, Cox & Gurney 1975).

The question remains as to how much of this inferred petrological structure is lithosphere, in the sense defined earlier. The granular, depleted zone (150 km; Boyd & Nixon (1975)) which shows no evidence of penetrative deformation can be assigned to the lithosphere with some confidence. The position of the sheared undepleted rocks beneath is less clear: they could be part of the circulating of the kimberlite from some depth below the base of the lithosphere. Alternatively they could represent former circulating mantle which had undergone thermal accretion to the base of the lithosphere and now formed part of it. Other interpretations of the deformational textures are discussed by Harte *et al.* (1975) and Green & Gueguen (1974).

Additional information is provided by the 'pseudoisochrons' now reported by a number of workers from suites of mantle extrusives (e.g. Brooks, James & Hart 1976). Pankhurst (1977) comments, 'The fact that Rb-Sr data for many sequences of basic continental volcanics plot on pseudo-isochrons giving ages greater than their extrusive ages is explicable in terms of inheritance of these characteristics from subcontinental mantle . . . heterogeneities in the continental lithosphere result from mantle differentiation at times up to 1500 Ma prior to volcanism. In many cases. . .it is possible that such fractionation events are chronologically related to major tectonothermal events which have a corresponding crustal expression, e.g. crust formation and cratonization, and that local crust/mantle systems have remained mechanically coupled since these events.' If this interpretation of the data is correct it has two main implications: that there is a long history of incorporation of some relatively undepleted mantle into the continental lithosphere, and that some continental basic volcanics are derived by partial fusion of such material *within* the continental lithosphere rather than from a source beneath it. It must be pointed out, however, that this interpretation of continental pseudoisochrons is not universally accepted and that continental basalts may undergo systematic contamination during their passage through the crust (S. Moorbath, personal communication).

In summary, the petrological evidence requires an unspecified (but probably not less than 150 km) thickness of mechanically coupled uppermantle beneath the continents. In some cases this lithosphere seems to have been acquired, at least in part, at the same time as major thermal events affected the crust. The upper parts of this lithosphere are generally the more depleted, although they probably contain eclogite pods (Dawson 1971) representing either basaltic magmas which failed to reach the surface, or remnants of old oceanic crust carried upwards by diapirs rising above a subduction zone.

(*d*) *Seismological evidence concerning the continental lithosphere*

One of the earliest papers to demonstrate that there were likely to be major differences in the uppermantle structure under the older parts of continents and under oceans was that of Anderson (1967) on the basis of surface wave dispersion. He showed that under old continental regions, a low velocity zone similar to that under the oceans was absent, and that shear velocities were generally higher under continents than oceans, possibly down to 500 km. This implied that although oceanic plates might be decoupled at a depth of 100 km, it was unlikely that continental plates were decoupled at that depth also. If systematic differences between oceanic and continental areas exist down to several hundred kilometres the plate structure in at least one of the two regions has to be at least that thick.

More recently the work of Jordan & Sipkin (e.g. Jordan 1975; Sipkin & Jordan 1975, 1976) on travel time residuals has suggested systematic differences in shear wave velocity between oceanic and continental areas down to 400 km or more, although some of their interpretations of surface wave and free oscillation data must be revised in the light of the results of Hart, Anderson & Kanamori (1976).

If these travel time residuals are substantiated, they are so large that compositional differences must be involved to explain them. As a rough guide, application of Birch's law based on empirical equations of state (Birch 1961; Soga 1971) shows that a change in the composition of olivine from Fo_{89} to Fo_{90} may be expected to give a shear wave velocity increase of about 0.5%. If the same method is applied to the two bulk compositions analysed by Boyd & McCallister (1976; for analyses see Boyd & Finger 1975), the depleted granular nodule gives a shear wave velocity slightly more than 1% higher than that of the undepleted nodule.

5. Discussion

We now summarize a possible model for the generation and evolution of continental lithosphere (figure 6).

At convergent plate margins solid diapirs of depleted mantle rise from the top of the descending slab and are added to the base of the overlying continental crust, building a lithosphere which is slightly different in composition (i.e. less aluminous and more magnesian) and therefore has a higher melting temperature than mantle at equivalent depths under the oceans. We have no means of estimating the thickness of depleted material which may be built up in this way, but because the nature of the diapirism will depend upon the slab age and subduction rate, it is likely to vary from place to place. We note, however, that only a small fraction of the depleted zone on the top of the slab may be added to the continental lithosphere; otherwise the continental lithosphere would be unacceptably thick. We should also expect a certain amount of undepleted mantle to be trapped between and above the depleted diapirs producing a lithosphere which was compositionally inhomogeneous.

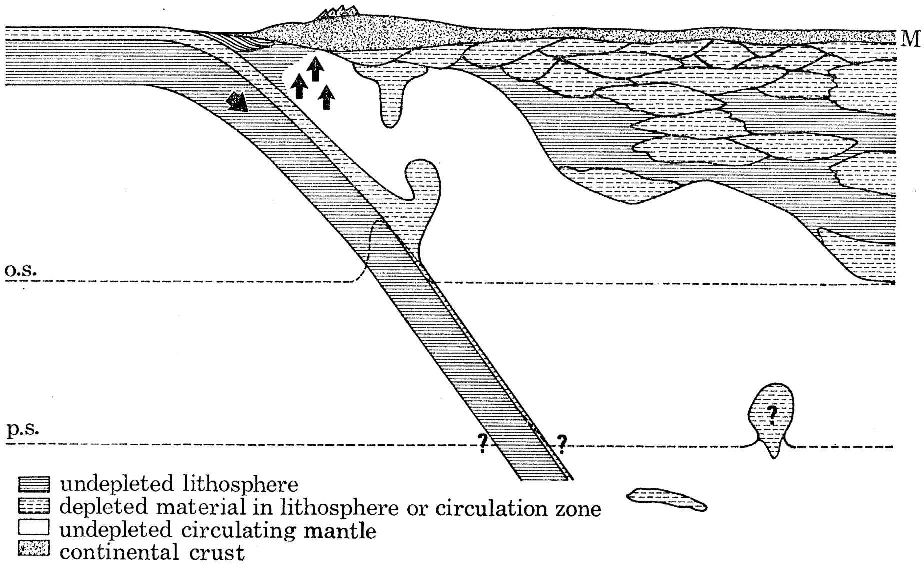

Figure 6. Subduction of a lithospheric slab with a depleted top and undepleted base (oceanic crust omitted), at a continental margin; vertical arrows indicate rise of magmas. A diapir is shown about to separate from the top of the slab: o.s., olivine–spinel phase change; p.s., post-spinel phase change.

Conductive loss of heat from the surface of the continents will allow cooling effects to extend progressively deeper into the lithosphere with time; this effect becomes particularly marked for times greater than 500 Ma when a significant fraction of the near surface radioactivity may have been eroded and redeposited elsewhere in sedimentary basins. Initially cooling would affect only the diapirically accreted part of the lithosphere, but unless that part was more than 400 km thick, cooling would eventually extend to its base and the possibility would exist of continued downward lithospheric growth by the thermal accretion of undepleted material from the circulating mantle beneath. However, if superplasticity (Sammis & Dein 1974) of mantle material undergoing phase changes is important, a zone of weakness associated with the olivine–spinel phase change might provide a lower limit to lithospheric growth.

The thermal and mechanical processes of lithospheric growth have a profound effect on the variation of the mantle contribution to surface heatflow with time. As previously discussed, diapiric accretion may add to the base of the continent material at a temperature a few hundred degrees lower than that of the ambient circulating mantle; the subsequent conductive equilibration of this depleted mantle involves cooling from above and warming from below; thermal contraction and expansion will both occur and there should be relatively little volume, and thus topographic, change. The mantle contribution to surface heat flow will initially fall more rapidly with time than in the absence of diapiric accretion but at longer times there will be little difference. The value to which it falls will depend on the radioactivity content of the lithosphere, the ultimate thickness of the lithosphere and the temperature of the circulating mantle below. Surface heat flow measurements indicate that q_m does not fall below about 0.5 h.f.u. even in the very oldest terrains and this constraint is sufficient to exclude various lithospheric models, e.g. a 400 km thick lithosphere composed entirely of depleted material would provide too low a heat flow; however, a lithosphere of that thickness which comprised only 50 % depleted material and for the rest comprised undepleted material with a radioactive heat production of about 0.2 h.g.u., would meet this requirement. If such a lithosphere were on average about 200 K cooler than mantle under the oceans the combined compositional and cooling effect would provide a velocity contrast of about 3 % with oceanic areas. A velocity contrast of this magnitude distributed over a 400 km thick layer would result in about a 2 s one-way vertical shearwave travel time difference between oceans and continents. Such a lithosphere could probably be made compatible with the evidence of suites of kimberlite nodules.

The recent proposal by Avraham & Nur (1976) is of interest in the light of the processes discussed above. They argue that the elevation of the topography in zones of continental collision is proportional to the total amount of oceanic lithosphere subducted prior to collision. The effect is too large to be attributable to any reasonable amount of thermal expansion and if the phenomenon is real, the most reasonable explanation is that low density mantle is in some way continuously contributed by the subducting slab. This could be the diapirism of depleted mantle from the top of the slab.

6. Conclusions

Mantle compositional inhomogeneity generated at oceanic ridges is associated with density differences which are of the same order as those normally attributed to temperature and thought to drive mantle convection. They should be sufficient to cause the diapiric ascent of depleted mantle above descending lithospheric slabs.

Diapiric ascent of depleted mantle and its accretion beneath continents is able to build a continental lithosphere with thermal and chemical characteristics significantly different from those of lithosphere formed solely by the thermal accretion of undepleted mantle.

If required by other evidence, a lithosphere 400 km or so thick could be generated by diapiric and thermal accretion which would both satisfy surface heat flow constraints and have a mean seismic velocity greater by a small percentage than oceanic upper mantle.

The authors wish to thank K. G. Cox, S. Moorbath, M. J. O'Hara, S. W. Richardson and P. N. Taylor, all of whom read early drafts of the manuscript and made valuable criticisms.

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Discussion

C. Froidevaux (*Laboratoire de Physique des Solides, Université Paris Sud, 91405 Orsay, France*). The authors' idea of underplating the continental lithosphere with buoyant depleted material coming from the downdipping oceanic plate behind island arcs raises two problems. First, the nodules in kimberlites indicate that the depleted material forms the top part, not the bottom, of the continental lithosphere. Secondly, I may remark that seismologists do not unanimously require a 400 km deep specific structure under continents.

E. R. Oxburgh. The authors' proposal for the diapiric rise of depleted lithosphere in no way depends upon the great thicknesses proposed, by some, of the lithosphere under the continents.

They simply point out that if such thicknesses do exist, diapiric accretion would be one, if not the only, way of generating them.

The authors agree that under the continents granular, depleted lherzolites often overlie sheared, undepleted lherzolites although this is not universally so. This relation is to be expected in lithospheres which formed initially by diapiric accretions of depleted mantle and subsequently cooled and thickened by the addition of undepleted material to its base.

B. Harte (*Grant Institute of Geology, Edinburgh University, West Mains Road, Edinburgh EH*9 3*JW*, *U.K.*). From the model presented by Dr Oxburgh, it would appear likely that underplating of oceanic crust with dunite and harzburgite would occur in intra-oceanic subduction zones. Such underplating would generate variations in the nature of oceanic lithosphere for which there appears to be little evidence.

E. R. Oxburgh. The authors should expect diapiric accretion to add depleted mantle material to the base of the lithosphere; this should occur behind intra-oceanic subduction zones as elsewhere. What little evidence is available suggests that the lithosphere and its crust is indeed anomalous in such areas (see, for example, Louden 1977 *Geophys. J.* **49**, 285). While not wishing to suggest that accretionary processes are necessarily or entirely responsible for such anomalies, it would be misleading to suggest either that the anomalies do not occur, or that they are adequately explained.

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