9/1

Composition and development of the continental tectosphere

Thomas H. Jordan - Seismologist

Geological Research Division, Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California 92093

Beneath the old continental nuclei are thick root zones which translate coherently during plate motions. These zones are apparently stabilised against convective disruption by the depletion of the continental upper mantle in a basalt-like component. Construction of this delicately balanced tectosphere is accomplished by the dynamic and magmatic processes of the Wilson cycle.

THE model of lithospheric plate tectonics has provided a framework for rationalising a vast array of geological, geophysical and geochemical observations. Its essential elements are easily described in qualitative terms. Barrell and others in the early part of this century recognised that the mechanical behaviour of the outer layers to surficial loads can be portrayed adequately by a lithosphere overlying an asthenosphere. By its classical definition, retained throughout this article, the lithosphere is the region of the crust and upper mantle capable of statically supporting significant deviatoric stresses over geologically long periods, whereas the asthenosphere is the region below the lithosphere where the mass flows associated with isostatic adjustments occur. More recently, it has been shown that the Earth's surface is partitioned into a mosaic of plates which deform very little as they engage in large-scale horizontal motions. The region of the Earth occupied by these coherent plates is here termed the tectosphere. Plate motions are presumably the surface manifestations of gravitational instability induced by heat production within the interior. As new sea floor spreads laterally from its site of creation on the oceanic rift system, some of this heat is lost by conduction to the surface, and the temperatures within a thermal boundary layer decrease with time.

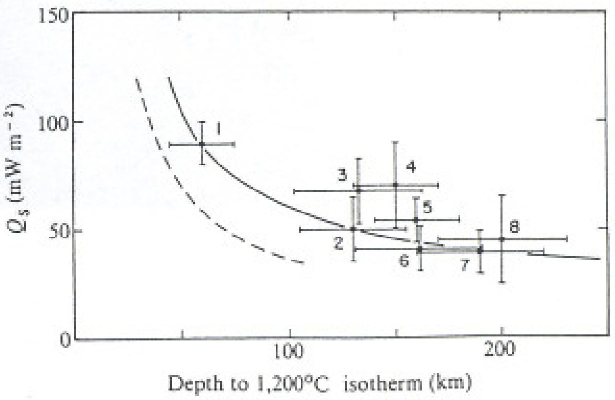
The essence of the lithospheric plate model is the assertion that the lithosphere, the tectosphere and the thermal boundary layer are identical in spatial extent, at least to the precision with which these terms are usefully defined. The need for the special term tectosphere, first used in this modern context by Elsasser[1] and Morgan[2] and later by Jordan[3], is thus obviated. The model in its current stage of development is less specific about the nature of the sub-lithospheric mantle, but it is generally assumed that the asthenospheric material is well mixed by convective action, hence, chemically homogeneous, the dominant heat transport mechanism is advection, and the vertical temperature gradients are approximately adiabatic[4]. This lithospheric plate model has been quite successful in explaining oceanic tectonics, but can it be applied to the continents?

Continental thermal evolution

The first measurements of oceanic heat flow in the early 1950s suggested that the temperatures beneath the ocean basins exceed those beneath the ancient continental nuclei to considerable depth[5,6]. Additional evidence supporting this view has since accumulated. Figure 1 shows the depth to the 1,200 °C isotherm inferred from geothermometric data on mantle xenoliths plotted against the estimated surface heat fluxes for localities in Africa, Asia, Australia and North America. The depth to the 1,200 °C isotherm is about 200 km beneath the stable continental terrains (such as Yakutia, Kimberley) whereas the average oceanic temperatures at this level are significantly greater[3]. It is inferred from these and other data that, to 200 km depth and more, the vertical temperature gradients beneath the shields and stable platforms are significantly super-adiabatic and the horizontal temperature gradients in the transitions to oceanic structures are large[3,8]. These inferences accord with the recent seismic data requiring substantial ocean–continent shear velocity and attenuation variations below a depth of 200 km[3,9,20,21,51].

Figure 1 also shows the well-known observation that the continental mantle is hottest in regions of recent magmatic activity (such as New South Wales, western USA) and apparently cools with time after the last major thermal event[4,7,10,11]. Crough and Thompson[11] have used the thermal boundary layer hypothesis to formulate an approximate model describing this behaviour. According to their model, the continental temperature profiles decay by vertical conductive heat transport, as do their oceanic analogues, evolving to steady state with a time constant $\tau \sim Q^{-2}$, where Q is the heat flux from below the boundary layer. They advocate a value for Q in the range 23–33 mW m^{-2}, which yields 150 Myr $< \tau <$ 300 Myr. The locus of the Crough–Thompson model for $Q = 33$ mW m^{-2} is shown in Fig. 1. This high value of the background heat flux predicts geotherms which are too steep and are difficult to reconcile with the available geothermometric data. To satisfy these data requires $Q < 15$ mW m^{-2}, implying a

Fig. 1 Continental surface heat flow plotted against depth to 1,200 °C isotherm. Data points are regionally averaged heat fluxes[42,43] versus 1,200 °C depths for palaeo-geotherms constrained by xenolith petrology. These palaeo-geotherms are assumed to be representative of present-day thermal conditions. Localities, age of xenolith emplacement and references are: 1, New South Wales, Australia (Quarternary)[49]; 2, Montana, USA (late middle Eocene)[44]; 3, Eastern Mojave, USA (post-Miocene) (A. L. Boettcher, unpublished); 4, Navajo Volcanics, USA (late Oligocene)[47]; 5, Northern Lesotho (late Cretaceous)[35,46]; 6, Frank Smith, S. Africa (late Cretaceous)[39]; 7, Kimberley, S. Africa (late Cretaceous)[40]; 8, Yakutia, USSR (late Palaeozoic)[48]. Error bars are subjective estimates. Solid line is locus of Pollack and Chapman's[8] conductive steady-state models. Dashed line is locus of Crough and Thompson's[11] thermal boundary layer model for $Q = 33$ mW m^{-2} (their equation (4)).



0028-0836/78/0274–0544$01.00

larger decay constant ($\tau > 700$ Myr) and a thicker steady-state boundary layer. The Pollack–Chapman[8] geotherms form a locus consistent with these constraints, and they are therefore used in subsequent calculations.

Assuming that the thermal boundary layer hypothesis was correct; then, the mass excess from thermal contraction integrated over the boundary layer would increase smoothly with time along a curve similar to Fig. 2. This mass excess is large, and, since it is not evident in the global free-air anomaly field[12], it must be isostatically compensated for all times, presumably at shallow depths as in the oceans. However, the behaviour implied for the crust is inconsistent with the known history of the continents. For example, cooling from a geotherm with a surface heat flow of 150 mW m^{-2} to one with 40 mW m^{-2} generates a mass excess $>9 \times 10^6$ kg m^{-2} below 40 km depth (Fig. 2). The compensation of this mass by the crustal column requires a crustal thickening of nearly 20 km. Although some thickening associated with thermal contraction has occurred locally in continental interiors (such as cratonic basins), in orogenic zones and along the margins[11,13], this model is vitiated by the existence of the exposed shields.

Furthermore, the base of the thermal boundary layer extends well below what can properly be called lithosphere. Studies of the topographic response to temporal changes in surface loads[14,15] and the correlation between existing surface loads and the gravity field[16,17] indicate that the effective thickness of the continental lithosphere for time intervals >10 Myr is <100 km. Hence, significant horizontal and vertical temperature gradients must exist within the continental asthenosphere. If the mantle is chemically homogeneous, these thermal gradients are accompanied by density gradients, and the subcontinental asthenosphere cannot be in hydrostatic equilibrium. For the configuration to be in static equilibrium requires deviatoric stresses to be maintained with magnitudes of the order of the vertical load differences[18,19]. These load differences exceed 10^7 Pa (100 bar) for the thermal models considered here (Fig. 2) and cannot be maintained without invoking a rheology for the mantle inconsistent with the surface loading data. Alternately, mass flow must occur, but such flow will generally tend to establish hydrostatic equilibrium, quickly destroying any large super-adiabatic thermal gradients below the continental lithosphere.

A stabilising mechanism

From these considerations, it seems that the lithospheric plate model in its simplest form does not account for the existence or evolution of a thick super-adiabatic (conductive) layer beneath the old continental nuclei. To establish a more satisfactory model, we retain the assumption that the tectosphere and the conductive layer are identical but abandon the identification of tectosphere with lithosphere[3]. From the seismic data[9,20] and the heat flow data[8], the thickness of the continental tectosphere must be highly variable; it is thin (<100 km) in unstable regions such as the western USA but probably reaches depths of the order of 400 km beneath the stable shields and platforms.

The previous discussion has emphasised the need for a mechanism to stabilise the large sub-lithospheric thermal gradients within the continental tectosphere against convective disruption. I have postulated the existence of chemical variations[3,21]. Compositional gradients are envisaged to be adjusted by dynamic processes so that the tectosphere extending below the lithosphere is in approximate hydrostatic equilibrium with its surrounding mantle; the *in situ* densities at a given level beneath continent and ocean are equal and any thermally produced mass excesses within the sub-lithospheric continental tectosphere are locally cancelled by chemically produced mass deficiencies. By this hypothesis, the mineralogical assemblages within the thick continental root zones are less dense than those beneath the oceans if referenced to the same pressure and temperature.

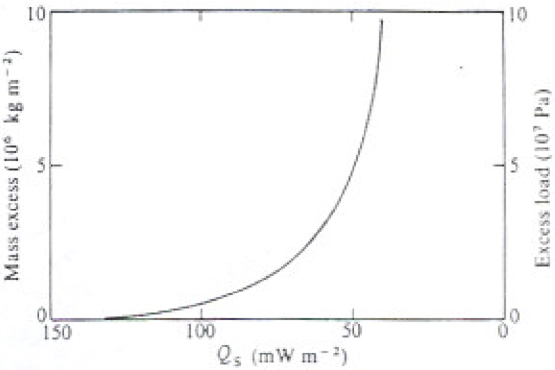


Fig. 2 Mass excess and corresponding vertical load difference due to thermal contraction in sub-continental mantle as functions of surface heat flow. Curve derived from Pollack and Chapman's[8] geotherms assuming geotherms decay from an adiabat with potential temperature of 1,300 °C and slope of 0.5 °C km^{-1}. Mass difference $-\alpha\rho\Delta T$ is integrated between 40 km depth and intersection of geotherm with adiabat using $\rho = 3{,}400$ kg m^{-3}, $\alpha = 3 \times 10^{-5}$ °C^{-1}. (10^7 Pa = 100 bar).

Basalt depletion hypothesis

A simple and attractive mechanism for generating these density differences is available: the chemical gradients stabilising the thermal gradients can be induced by the removal of a basaltic magma from the mantle[22,23]. Mantle mineralogy in the depth range 70–400 km approximates that of a garnet lherzolite[26,27], and the subtraction of basalt from garnet lherzolite leaves a residue less dense than the parental material[23–25]. If the degree of basaltic depletion has a negative correlation with temperature, the continental tectosphere can be in hydrostatic equilibrium with its surrounding mantle. The requisite relative density decrease is $-\alpha\Delta T$, where α is the coefficient of thermal expansion and ΔT is the ocean–continent temperature contrast. In the depth range 150–200 km, the average temperature contrast between the ocean basins and shields probably lies in between 300 and 500 °C (refs 3, 8). Thus for $\alpha = 3 \times 10^{-5}$ per °C the hypothesis requires a compositionally induced density difference between −0.9 and −1.5% at these depths.

To test this hypothesis, a simple algorithm for estimating the mineral compositions and densities of garnet lherzolites from their whole rock oxide composition has been devised[25]. The mineral composition of four phases (garnet, clinopyroxene, orthopyroxene, olivine) are calculated from the seven major oxide components (SiO_2, Al_2O_3, Cr_2O_3, FeO, MgO, CaO, Na_2O; total Fe computed as FeO) by specifying various apparent distribution and partition coefficients. The densities of the mineral phases and the whole rock are computed assuming the additivity of end-member molar volumes. The whole rock density obtained by this normative scheme, denoted $\hat{\rho}$, is the density estimate for a garnet lherzolite assemblage equilibrated at mantle conditions (~5×10^9 Pa, 1,200 °C) but measured at standard conditions (10^5 Pa, 25 °C). The present application requires only the density differences of various compositions at fixed pressure and temperature; hence, the corrections for elevated pressures and temperatures are second-order in small quantities and can be ignored.

The normative density algorithm has been applied to the available whole rock analyses of garnet lherzolite xenoliths from kimberlite pipes and to various model compositions of the oceanic upper mantle[25]. The mean oxide composition of 78 xenoliths is listed with its normative parameters in Table 1; this average continental garnet lherzolite (ACGL) is similar to the estimates of other authors[28–30], although more samples with a greater geographical distribution are used here. Most of the xenoliths are from South African localities, but pipes in the USA and USSR are also represented. The ACGL is an approximation to the average mantle composition beneath the

Table 1 Parameters of an average continental garnet lherzolite (ACGL) compared with model compositions of the oceanic upper mantle

Composition (wt %)	ACGL*	Pyrolite[31]	St Paul's Rocks (wt avg)[32]
SiO_2	45.55 ± 1.45	45.30	44.63
TiO_2	0.11 ± 0.06	0.71	0.29
Al_2O_3	1.43 ± 0.60	3.55	3.78
Cr_2O_3	0.34 ± 0.14	0.43	0.52
FeO	7.61 ± 0.94	8.48	8.15
MnO	0.11 ± 0.02	0.14	0.12
MgO	43.55 ± 1.98	37.60	39.40
CaO	1.05 ± 0.77	3.09	2.67
Na_2O	0.14 ± 0.10	0.57	0.34
K_2O	0.11 ± 0.10	0.14	0.10
Norm (wt %)			
Gt	6.0	12.3	15.3
Cpx	4.5	16.3	11.3
Opx	22.5	13.5	11.9
Ol	67.0	57.9	61.5
Normative density ($kg\ m^{-3}$)	3,353	3,397	3,401

* Average of 78 analyses of kimberlite xenoliths[25]; oxide composition expressed as sample mean ± sample standard deviation.

shields and stable platforms in the depth range 150–200 km (ref. 25). Two model compositions for the oceanic mantle are compared with the ACGL in Table 1, Ringwood's pyrolite[31] and a weighted composition of St Paul's Rocks[32]. Relative to these compositions, the ACGL is depleted in normative garnet and clinopyroxene, and its whole rock Fe/Fe+Mg ratio is lower, suggesting basaltic depletion of the subcontinental upper mantle (Table 1, Fig. 3). The depleted major-element chemistry of the mantle sampled by the xenoliths has been noted by many petrologists, most explicitly by Ringwood[31,26] and Nixon and Boyd[33–35].

Because it is depleted in basaltic components, the ACGL is less dense than the model oceanic garnet lherzolites by about 1.3% ± 0.2%. This density difference is sufficient to offset the volume contraction resulting from a cooling of about 400 °C, which falls in the range of the ocean-shield temperature contrasts estimated from thermal models. Thus, one implication of the basalt depletion hypothesis is substantiated.

Another implication provides a more stringent test. Beneath the continents, in regions of large super-adiabatic gradients, the degree of basaltic depletion should decrease with depth coherently with the decrease in the ocean–continent temperature contrast. This is, in fact, observed[35], as shown in Fig. 4. The normative densities for various xenoliths in Fig. 4 correlate negatively with the Ca/Ca+Mg ratios in the clinipyroxenes. Since these ratios decrease with temperature[37,38] and temperature increases with depth, the data in Fig. 4 indicate a rise in normative density with depth caused by increases in Fe/Fe+Mg and normative garnet. The magnitude of this rise agrees well with the theoretical curves based on the basalt depletion hypothesis (Fig. 4).

The model of a thick basalt-depleted continental tectosphere is also consistent with the seismic observations of substantial ocean-shield shear velocity and attenuation contrasts in the upper mantle[3,9,20,21,51]. Temperature is probably the dominant factor in controlling the lateral variations in shear velocity, but chemical depletion may also contribute to establishing high mantle shear velocities in the cratons. The mean atomic weight is lowered by basalt removal, which increases the shear velocities[50,25], but, more importantly, basalt depletion raises the mantle solidus; for temperatures near the solidus, this can substantially elevate the velocities and reduce attenuation[57].

Isotope and trace element problem

The hypothesis of a basalt-depleted continental tectosphere agrees well with several independent lines of geophysical and petrological evidence, but it apparently conflicts with trace element and isotopic data. Continental basalts, for example, consistently display $^{87}Sr/^{86}Sr$ ratios which are higher and $^{143}Nd/^{144}Nd$ ratios which are lower than those of mid-ocean ridge basalts[52,53], and geochemists have inferred that the subcontinental mantle is considerably more enriched in large-ion lithophile (LIL) elements than its oceanic counterpart. Brooks, James and Hart[52] extend this reasoning to argue for the existence of a continental tectosphere (or, in their terminology, lithosphere) "that is far from depleted, and which, under favourable thermal conditions, would be a ready source of basaltic magma." Clearly, from the previous discussion, this conclusion is favoured by neither the geophysical nor the petrological data.

The model advocated by Brooks and others is one interpretation of the available isotopic data, but it is by no means unique; there are tenable explanations which preserve the basic hypotheses stated here. For example, the ultimate source regions for the continental flood basalts, which provide much of the isotopic data, probably lie below the tectosphere in a more fertile layer of the mantle. This (hot) fertile material may intrude the tectosphere during episodes of continental rifting, when most flood basalts are produced.

Nevertheless, we must admit the possibility that the continental tectosphere is actually depleted in basaltic components (low Al_2O_3, FeO, CaO relative to MgO) but is enriched, at least locally, in LIL minor elements (for example, Rb, Sr, Cs, Ba, U, Th). This configuration is suggested by data from the basalt-depleted garnet lherzolite xenoliths themselves; those analysed show high $^{87}Sr/^{86}Sr$ ratios and low K/Rb, K/Sr, K/Cs and K/Ba ratios[54,58–61]. LIL enrichment of previously depleted mantle could result from metasomatic processes involving the flux of volatiles or highly differentiated liquids[54–56]. In fact, given the great age and complex evolutionary history of most tectospheric columns, extensive metasomation is quite plausible; once depleted in basaltic components, the tectosphere is less likely to be the site of further large-scale melting, and it provides a ready trap and long-term reservoir for the very mobile mantle constitutents.

Aspects of continental evolution

Previous investigators have attributed the substantial contrasts between continental and oceanic tectonics to the differences in crustal structure[62]. These differences are certainly important and probably explain why continental lithosphere is generally more easily deformed than oceanic lithosphere and why continental seismicity is more diffuse. But crustal structure is only part of the story.

It is proposed here that irreversible basaltic depletion of the upper mantle is crucial in regulating continental development and cratonic stability; that is, a fundamental coherence between surface tectonic behaviour and upper mantle compositional structure is postulated. Once depleted, low-density residual peridotite cannot be easily remixed with undepleted mantle. A depleted tectosphere is stabilised at lower temperatures, its solidus is elevated and its effective viscosity is thereby greater than undepleted mantle. These properties should drive the horizontal shear straining associated with plate motions to greater depths and isolate the crust from the thermal perturbations and mass motions associated with small scale convective disturbances. The existence of a 'decoupling zone' of concentrated horizontal shear strain beneath the continental tectosphere remains problematic, as is the distribution of driving forces, but the coherent motions of thick continental roots are more plausible if convection involves the lower mantle. This concept is in accordance with other data[21,63–65]. Clearly, in the context of this model, continental drift involves more than just the passive rafting of sialic crust on an otherwise uniform lithosphere. If the model is

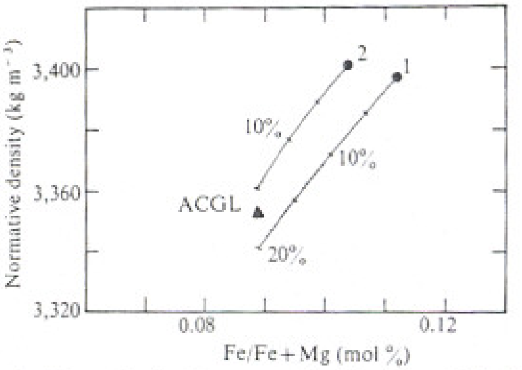


Fig. 3 Normative density compared with whole rock Fe/Fe+Mg ratio for average continental garnet lherzolite (ACGL) and two oceanic upper mantle model composition: 1, pyrolite[31]; and 2, weighted average of St Paul's Rocks[32]. Shaded field contains 67% of the samples used to derive the ACGL. Curves show displacements of oceanic model compositions by subtraction of indicated molar percentages of basalt[25]; pyrolite–olivine basalt (ref. 26, Table 4-2) and St Paul's Rocks–average oceanic tholeiite[36].

creditable, a full understanding of mantle dynamics will require the inclusion of forces due to compositional gradients into the governing equations. (Such forces are often encountered in physical oceanography; indeed, there is a notable analogy between the salt of the oceans and the basalt of the Earth.)

A thick, cool, depleted tectosphere capped by a mature continental crust is envisaged to be the eventual consequence of the magmatic and dynamic processes intrinsic to the Wilson cycle of rifting, drifting, closure and collision. This basic plate tectonic cycle has evidently operated throughout the Phanerozoic and the Proterozoic[66,67], and probably well back into the Archean[68,69]. For a tentative sketch of tectospheric evolution, there is no need to deviate from actualistic statements of the fundamental processes. Obvious temporal differences, such as the contrasts between Archean and modern tectonics, seem to arise from progressive tectospheric growth and associated crustal stabilisation[68].

In the current continental environment, basaltic depletion of the uppermost mantle occurs in two primary settings: during the magmatic episodes associated with continental rifting, especially by the extrusion of massive continental flood basalts[74], and in the mantle wedge above subduction zones[70–73]. (Averaged over Phanerozoic time, the latter site seems to have produced a much greater volume of basaltic rock.) Once significant depletion has occurred, the mantle is resistant to subduction and can support lower temperatures without convective recycling. Volatile flux or highly differentiated magmas from either subducted slabs or the deep mantle can impregnate the tectosphere with mobile LIL constituents and modify the details of its trace-element chemistry without inducing large-scale melting.

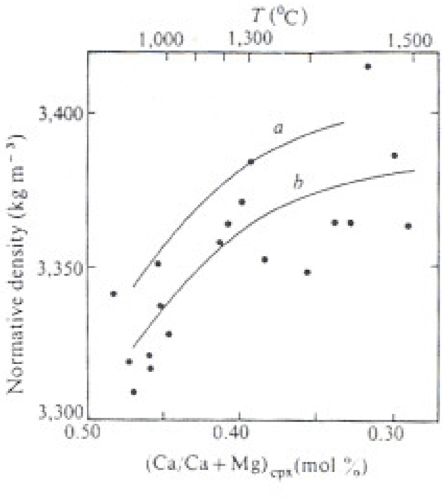
But geometrical[72] and petrological[26,73] constraints indicate that magmatic events efficiently deplete the mantle in basaltic constituents only to depths of 100–200 km. Some residual peridotite could diapirically rise from the descending slabs to augment the depleted wedges above subduction zones[75], but the protocontinental tectosphere produced in island-arc environments or along the active margins is probably thin, chemically heterogeneous and poorly consolidated. In the eventual course of the Wilson cycle, however, the dispersed regions of depleted mantle and their superjacent crustal columns are swept up and accreted to the primary continental masses.

Further consolidation and thickening of the tectosphere result from the major compressive events at convergent boundaries, particularly episodes of continent–continent collision. The history of the crust following a continental collision has been analysed by Dewey and Burke[76]. During very violent collisions of Atlantic-type and Andean-type margins, such as the one occurring today between India and Asia, the crust of the Andean-type margin is compressively thickened by a factor of two or more, leading to the uplift of a Tibetan-like plateau. Dewey and Burke envisage this process of compressive crustal thickening to involve deformation essentially continuously distributed over a broad zone. If this occurs, it should be concomitant with a proportional thickening of the subcrustal tectosphere by the lateral and downwards advection of depleted peridotite. In this way, the thick root zones of the stable shields and platforms could be formed.

Thus, depletion, consolidation and thickening are the important constructive processes in continental tectospheric development. However, their iterative application over many Wilson cycles may be necessary to stabilise a thick tectosphere in any one region, because these constructive processes must compete against destructive processes. Whereas large-scale basaltic depletion is nearly irreversible, consolidation and thickening of the tectosphere are not. Any significant heating of a cool, thick, cratonic tectosphere by conduction from its periphery, by diapiric intrusion or by internal heat sources results in its thinning and dispersal by the upwards and lateral flow of low-density depleted material. Such motions may have an important role in episodes of continental rifting and breakup. Extension within the Basin and Range Province and the uplift of the Colorado Plateau may exemplify the consequences of an instability induced by tectospheric heating.

It is hypothesised that, throughout geological history, continental growth and stabilisation have been regulated by tectospheric development; that is, tectospheric depletion, consolidation and thickening are the fundamental chelogenic processes of the Wilson cycle. Presumably, magmatic differentiation before and during the Katarchean (>3,000 Myr BP) was very active, leading to the localised formation of primitive sialic crust[69,78] and generating a large volume of depleted mantle. However, the existing continental tectosphere was probably thin, perhaps because mantle convection was more vigorously driven by an increased rate of radiogenic heat production[77]. The diachronous stabilisation of the nuclear Archean cratons in

Fig. 4 Normative density plotted against Ca/Ca+Mg in clinopyroxene. Temperatures correspond to the diopside–enstatite solvus of Nehru and Wyllie[38]. Points represent garnet lherzolite xenoliths for which both whole rock and clinopyroxene analyses are available[25]. Theoretical curves are based on the hypothesis that basaltic depletion exactly cancels the density increase due to temperatures lower than the oceanic adiabat. To construct curves *a* and *b*, temperatures from Pollack and Chapman's[8] 40 mW m^{-2} geotherm were subtracted from oceanic adiabats with potential temperatures of 1,300 °C and 1,500 °C, respectively, and slopes of 0.5 °C km^{-1}. Normative densities were constructed assuming a pyrolite composition for the oceanic upper mantle ($\hat{\rho} = 3{,}397$ kg m^{-3}) and a thermal expansion coefficient of 3×10^{-5} °C^{-1}; these densities are plotted against temperatures on the continental geotherm using the upper scale. The increase in normative density with equilibration temperature observed for the xenoliths supports the basaltic depletion hypothesis.



the interval 2,800–2,000 Myr BP (ref. 78) is inferred to mark the earliest formation of thick tectospheric root zones with long-term stability. This development may be responsible for the fundamental transition from Katarchean permobility to Proterozoic linear (geosynclinal and mobile belt) tectonics.

I thank A. L. Boettcher, F. R. Boyd, S. Ehrenberg and I. D. MacGregor for their unpublished data. This research was sponsored by the NSF.

Received 21 March; accepted 7 June 1978.

1. Elsasser, W. M. in *The Application of Modern Physics to Earth and Planetary Interiors* (ed. Runcorn, S. K.) 223 (Interscience, New York, 1969).
2. Morgan, W. J. *J. geophys. Res.* **73,** 1959 (1968).
3. Jordan, T. H. *Rev. geophys. Space Phys.* **13,** 1 (1975).
4. Sclater, J. G. & Francheteau, J. *Geophys. J. R. astr. Soc.* **20,** 509 (1970).
5. Bullard, E. C. *Nature* **170,** 200 (1952).
6. MacDonald, G. J. F. *Rev. Geophys.* **1,** 58 (1963).
7. Pollack, H. N. & Chapman, D. S. *Earth planet. Sci. Lett.* **34,** 174 (1977).
8. Pollack, H. N. & Chapman, D. S. *Tectonophysics* **38,** 279 (1977).
9. Sipkin, S. A. & Jordan, T. H. *J. geophys. Res.* **81,** 6307 (1976).
10. Polyak, B. G. & Smirnov, Ya. B. *Geotectonics* **4,** 205 (1968).
11. Crough, S. T. & Thompson, G. A. *J. geophys. Res.* **81,** 4857 (1976).
12. Kaula, W. M. in *The Nature of the Solid Earth* (ed. Robertson, E. C.) 385 (McGraw-Hill, New York, 1972).
13. Sleep, N. H. & Snell, N. S. *geophys. J. R. astr. Soc.* **45,** 125 (1976).
14. Walcott, R. I. *J. geophys. Res.* **75,** 3941 (1970).
15. Cathles, L. M. *The Viscosity of the Earth's Mantle* (Princeton, University Press, New Jersey, 1975).
16. Banks, R. J., Parker, R. L. & Huestis, S. P. *Geophys. J. R. astr. Soc.* **51,** 431 (1977).
17. McNutt, M. & Parker, R. L. *Science* **199,** 773 (1978).
18. Artyushkov, E. V. *J. geophys. Res.* **78,** 7675 (1973).
19. Sclater, J. G., Lawver, L. A. & Parsons, B. *J. geophys. Res.* **80,** 1031 (1975).
20. Knopoff, L. *Tectonophysics* **13,** 497 (1972).
21. Jordan, T. H. *Nature* **257,** 745 (1975).
22. Jordan, T. H. *Geol. Soc. Am. Abstr. Progr.* **8,** 944 (1976).
23. O'Hara, M. J. *Nature* **253,** 708 (1975).
24. Boyd, F. R. & McCallister, R. H. *Geophys. Res. Lett.* **3,** 509 (1976).
25. Jordan, T. H. *Proc. 2nd Int. Kimberlite Conf.* (eds Boyd, F. R. & Meyer, H. O. A.) (American Geophysical Union, Washington, D.C., in the press).
26. Ringwood, A. E. *Composition and Petrology of the Earth's Mantle* (McGraw-Hill, New York, 1975).
27. Yoder, H. S., Jr, *Generation of Basaltic Magma* (NAS, Washington, D.C., 1976).
28. Ito, K. & Kennedy, G. C. *Am. J. Sci.* **265,** 519 (1967).
29. Carswell, D. A. & Dawson, J. B. *Contr. Mineral. Petrol.* **25,** 163 (1970).
30. Sobolev, N. V. *Deep-Seated Inclusions in Kimberlites and the Problem of the Composition of the Upper Mantle* (trans. by Brown, D. A., ed. Boyd, F. R.) (American Geophysical Union, Washington, D.C., 1977).
31. Ringwood, A. E. in *Advances in Earth Sciences* (ed. Hurley, P.), 287 (MIT Press, Cambridge, 1966).
32. Melson, W. G., Jarosevich, E., Bowen, V. T. & Thompson, G. *Science* **155,** 1532 (1967).
33. Nixon, P. H. & Boyd, F. R. in *Lesotho Kimberlites* (ed. Nixon, P. H.) 48 (Lesotho National Development Corporation, Maseru, 1973).
34. Nixon, P. H., Boyd, F. R. & Boullier, A.-M. in *Lesotho Kimberlites* (ed. Nixon, P. H.) 312 (Lesotho National Development Corporation, Maseru, 1973).
35. Boyd, F. R. & Nixon, P. H. *Phys. Chem. Earth* **9,** 431 (1975).
36. Engel, A. E. J., Engel, C. G. & Havens, R. G. *Geol. Soc. Am. Bull.*, **76,** 719 (1965).
37. Davis, B. T. C. & Boyd, F. R. *J. geophys. Res.* **71,** 3567 (1966).
38. Nehru, C. E. & Wyllie, P. J. *Contr. mineral. Petrol.* **48,** 221 (1974).
39. Boyd, F. R. *Yb. Carnegie Inst. Wash.* **73,** 285 (1974).
40. Boyd, F. R. & Nixon, P. H. *Yb. Carnegie Inst. Wash.* **75,** 554 (1976).
41. Mercier, J.-C. & Carter, N. L. *J. geophys. Res.* **80,** 3349 (1975).
42. Jessop, A. M., Hobart, M. A. & Sclater, J. G. *Geothermal Ser. No. 5* (Earth Physics Branch, Ottawa, 1975).
43. Melnikov, P. I., Balobaev, V. T., Kutasov, I. M. & Devyatkin, V. N. in *Geoelectric and Geothermal Studies* (ed. Adám, Á.) 473 (Akadémiai Kiadó, Budapest, 1976).
44. Hearn, B. C., Jr & Boyd, F. R. *Phys. Chem. Earth* **9,** 247 (1975).
45. MacGregor, I. D. *Phys. Chem. Earth* **9,** 455 (1975).
46. MacGregor, I. D. *Proc. 2nd Int. Kimberlite Conf.* (eds Boyd, F. R. & Meyer, H. O. A.) (American Geophysical Union, Washington, D.C. in the press).
47. Ehrenberg, S. *Proc. 2nd Int. Kimberlite Conf.* (eds Boyd, F. R. & Meyer, H. O. A.) (American Geophysical Union, Washington, D.C., in the press).
48. Boyd, F. R., Fujii, T. & Danchin, R. V. *Yb. Carnegie Inst. Wash.* **75,** 523 (1976).
49. Ferguson, J., Ellis, D. J. & England, R. N. *Geology* **5,** 278 (1977).
50. Birch, F. in *The Earth's Crust and Upper Mantle* (ed. Hart, P. J.) 18 (American Geophysical Union, Washington, D.C., 1969).
51. Sacks, I. S. & Okada, H. *Phys. Earth planet. Inter.* **9,** 211 (1974).
52. Brooks, C., James, D. E. & Hart, S. R. *Science* **193,** 1086 (1976).
53. DePaolo, D. J. & Wasserberg, G. J. *Geophys. Res. Lett.* **3,** 743 (1976).
54. Shimizu, N. *Phys. Chem. Earth* **9,** 655 (1975).
55. Erlank, A. J. & Rickard, R. S. in *Proc. 2nd Int. Kimberlite Conf.* (eds Boyd, F. R. & Meyer, H. O. A.) (American Geophysical Union, Washington, D.C., in the press).
56. Boettcher, A. L., O'Neil, J. R., Windom, K. E., Stewart, D. C. & Wilshire, H. G. *Proc. 2nd Int. Kimberlite Conf.* (eds Boyd, F. R. & Meyer, H. O. A.) (American Geophysical Union, Washington, D.C. in the press).
57. Goetze, C. in *High Pressure Research, Applications in Geophysics* (eds Manghnani, M. H. & Akimoto, S.-I.) 3 (Academic, New York, 1977).
58. Barrett, D. R. *Phys. Chem. Earth* **9,** 637 (1975).
59. Rhodes, J. M. & Dawson, J. B. *Phys. Chem. Earth* **9,** 545 (1975).
60. Ridley, W. I. & Dawson, J. B. *Phys. Chem. Earth* **9,** 559 (1975).
61. Kramers, J. D. *Earth planet. Sci. Lett.* **34,** 419 (1977).
62. McKenzie, D. P. *Geophys. J. R. astr. Soc.* **18,** 1 (1969).
63. O'Connell, R. J. *Tectonophysics* **38,** 119 (1977).
64. Jordan, T. H. *J. Geophys.* **43,** 473 (1977).
65. Davies, G. F. *Geophys. J. R. astr. Soc.* **49,** 459 (1977).
66. Jacobs, J., Russell, R. D. & Wilson, J. T. *Physics and Geology* 2nd edn (McGraw-Hill, New York, 1972).
67. Burke, K., Dewey, J. F. & Kidd, W. S. F. *Tectonophysics* **40,** 69 (1977).
68. Burke, K., Dewey, J. F. & Kidd, W. S. F. in *The Early History of the Earth* (ed. Windley, B. F.) 113 (Wiley, London, 1976).
69. Condie, K. C. *Plate Tectonics and Crustal Evolution* (*Pergamon, New York*, 1976).
70. Wyllie, P. J. *J. geophys. Res.* **76,** 1328 (1971).
71. McBriney, A. R. (ed.) *Proc. Andesite Conf.* Bull. 65 (Department of Geology & Mineral Industries, State of Oregon, 1969).
72. Dickinson, W. R. *Geology* **3,** 53 (1975).
73. Boettcher, A. L. *Tectonophysics* **17,** 223 (1973).
74. Cox, K. G. *J. geol. Soc. Lond.* **128,** 311 (1972).
75. Oxburgh, E. R. & Parmentier, E. M. *J. geol. Soc. Lond.* **133,** 343 (1977).
76. Dewey, J. F. & Burke, K. C. A. *J. Geol.* **81,** 683 (1973).
77. Dickinson, W. R. & Luth, W. C. *Science* **174,** 400 (1971).
78. Windley, B. F. *Phil. Trans. R. Soc.* **A273,** 321 (1973).

Palynology, palaeomagnetism and radiometric dating of Flandrian marine and freshwater sediments of Loch Lomond

J. H. Dickson & D. A. Stewart
Department of Botany, University of Glasgow,

M. S. Baxter & N. D. Drndarsky
Department of Chemistry, University of Glasgow

R. Thompson & G. Turner
Department of Geophysics, University of Edinburgh

J. Rose
Department of Geography, Birkbeck College, University of London, UK

Loch Lomond in Scotland was part of the sea not only in late Devensian times but also in the middle Flandrian. Deep water cores from the southern basin show sediment with marine plankton and low remanent magnetism (RMN) between freshwater sediments. Raised beaches, deltas and estuarine flats were formed during the Flandrian transgression which ^{14}C dating shows to have lasted some 1,450 yr from 6,900 to 5,450 yr BP. Palaeomagnetic matching enables an intercomparison of Windermere and Lomond ^{14}C timescales and implies that both may be in error.

LOCH LOMOND takes the form of a north–south trending glacial trough[1] which is deep and narrow at the north where bedrock is predominantly mica schist and schistose grit, and shallow and wide in the south where the bedrock is Old Red Sandstone and Carboniferous sandstones, marls and basalts. In the highland zone the main rivers such as the Falloch, Douglas and Fruin drain into the northern and western sides of the loch, whereas in the lowland zone the main river is the Endrick Water which drains into the south-east corner of the loch from a large catchment that includes much of the Kilpatrick Hills and the Campsie Fells (Fig. 1, inset). The River Leven forms the outlet to Loch Lomond and drains southwards through the Vale of Leven to the Clyde estuary at Dumbarton. The water level in Loch Lomond is about 8 m o.d.

0028-0836/78/0274—0548$01.00