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Continents as a chemical boundary layer

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Tectospheric structure can be described in terms of three basic types of surficial boundary layers: chemical (c.b.l.), mechanical (m.b.l.) and thermal (t.b.l.). Beneath old ocean basins the thickness of the c.b.l. (*ca.* 40 km) is less than that of either the m.b.l. (*ca.* 100 km) or the t.b.l. (*ca.* 150 km), but the hypothesis that a similar structure underlies the old continental cratons is difficult to reconcile with seismic observations. We therefore examine an alternate model which postulates a much thicker c.b.l. beneath the cratons whose mantle component consists of a low-density peridotite depleted in its basaltic constituents. On the basis of seismological and petrological data it is inferred that this augmented c.b.l. extends below the m.b.l. to depths exceeding 150 km and acts to stabilize a thick (> 200 km) t.b.l. against convective disruption. Because of its refractory nature the sub-m.b.l. portion of the c.b.l. constitutes a stable geochemical reservoir which has evidently been impregnated by large-ion lithophile elements fluxing from the deep mantle or from descending slabs. Consequently, its heat production is high (*ca.* 0.1 $\mu W/m^3$) and it contributes significantly to the surface heat flux. The evolutionary history and dynamics of the continental c.b.l. are not well understood, especially the role of double-diffusive instabilities, but the fusion of the continental masses into 'supercontinents' and the orogenic compression that this entails are thought to be important processes in c.b.l. formation.

1. Introduction

The Earth's mass density ρ increases with depth to a value of *ca.* 13 Mg/m^3 at its centre. Some of the increase is due to gravitational self-compression, but well over half occurs across two major compositional transitions, one at the planetary surface ($\Delta\rho \approx 3.3\ Mg/m^3$) and one at the core–mantle boundary ($\Delta\rho \approx 4.4\ Mg/m^3$). The mass-transport mechanisms presumed to be operating within the mantle and core should tend to deposit chemically differentiated material at these transitions and thereby to form compositionally distinct boundary layers of intermediate density. As yet, the evidence for a chemical boundary layer (c.b.l.) at the core–mantle interface is largely circumstantial (Jordan 1979*a*; Ringwood 1979), but at least the upper part of the c.b.l. at the Earth's surface, the crust, can be observed directly.

The most striking feature of the crust is the dichotomy of continents and oceans. Whereas the thin simatic crust of the oceans is continually being resorbed into the mantle via the subduction process, the thicker sialic crust of the continents remains buoyantly trapped at the surface. The disposition of continental crust thus exerts considerable control on plate kinematics (McKenzie 1969). Furthermore, crustal thickness is a significant parameter in determining the rheology of the lithosphere and consequently the strain field within the plates caused by forces exerted at plate boundaries (McKenzie 1969).

It has been conjectured that these recognized crustal effects are sufficient to explain the various peculiarities of continental tectonics visible at present or evident in the geological record: e.g. the diffuse nature of continental seismicity, the complexity of continental plate boundaries and continent–continent interactions, the tendency for continental masses to collect into 'supercontinents', the age distribution of continental rocks. In particular, a significant group of investigators believes that below a depth of 50 km or so the thermal, chemical and mechanical properties of the lithospheric plates are largely independent of crustal type and can be couched in a unified theory of thermal-boundary-layer evolution without endowing special properties to the sub-continental mantle or special dynamics to continental drift (see, for example: Crough & Thompson 1976; Kono & Amano 1978; Sclater *et al.* 1980; Crough 1979).

These models are elegant, simple and consistent with much of the data on plate structure and crustal evolution; as straightforward extensions of plate tectonic concepts (McKenzie 1967, 1969) they represent the paradigm against which all alternative hypotheses must be compared.

The alternative model examined in this paper is founded on the notion that the crust forms only the upper part of the surficial c.b.l. Below the crust is presumed to be a layer of refractory peridotite depleted in major-element basaltic constituents whose influence on plate structure and dynamics is not inconsiderable, especially beneath the continental shields, where it is inferred to be thickest (Jordan 1975, 1978). According to the model it is this augmented c.b.l. that in cratonic areas governs the thickness of the thermal boundary layer (t.b.l.) and, therefore, the depth to the base of the plate (tectosphere). The situation postulated for the continental cratons thus differs fundamentally from that envisaged for the oceans, where a mechanical boundary layer (m.b.l.), or lithosphere, is thought to control the tectospheric thickness (Elsasser 1969; Parsons & McKenzie 1978).

Admittedly the model introduces complications avoided by simple t.b.l.-cooling models; continental drift cannot then be the merely passive rafting of a superficial crustal edifice on otherwise undistinguished lithosphere. In return, however, the model accounts for certain perplexing seismological and petrological observations and offers new insights into the problem of continental evolution.

2. Densities of lherzolites

Magma generated at subcrustal depths is primarily basaltic in composition and, according to most petrological models (Ringwood 1975; Yoder 1976), is produced by the partial fusion of lherzolite. It is a curious yet fundamental fact that the peridotite left as a residuum of such melting generally has a density below that of the parental rock (O'Hara 1975; Green & Liebermann 1976; Boyd & McCallister 1976; Oxburgh & Parmentier 1977).

The details of the density relationships among garnet lherzolites have been investigated in a previous paper (Jordan 1979*b*), where a simple garnet lherzolite norm was devised to account for the variations of mineralogy. The algorithm employs various apparent distribution and partition coefficients to compute the compositions of four phases (garnet (Gt), clinopyroxene (Cpx), orthopyroxene (Opx) and olivine (Ol)) from the seven major oxide components of the whole rock [SiO_2, Al_2O_3, Cr_2O_3, FeO, MgO, CaO, Na_2O]. The densities of the mineral phases the whole rock are calculated by adding the end-member molar volumes estimated at room temperature and pressure. The whole-rock density obtained by the normative scheme is thus the density for a garnet lherzolite assemblage equilibrated at mantle conditions (*ca.* 5×10^9 Pa, 1200 °C) but observed at standard conditions (10^5 Pa, 25 °C). This normative density, denoted $\hat{\rho}$, provides a standard for the comparison of both model compositions, such as pyrolite, and real rocks, such as kimberlite xenoliths. Examples are given in table 1. In addition to normative mineralogies and densities, seismic velocities calculated by the Voight–Reuss–Hill averaging method are also listed.

The effect of basalt depletion is to reduce the normative density and increase the seismic velocities. The subtraction of 20 mol % olivine basalt from pyrolite decreases $\hat{\rho}$ by about 1.7 % and increases $\hat{V}_s$ by 0.5 %. The relative differences between the 'fertile' kimberlite xenolith PHN1611 and the 'barren' xenolith PHN1569 are even larger, -3.1 % and $+2.2$ %, respectively. The latter example provides a check on the algorithm: the calculated densities of PHN1569 and PHN1611 agree with those measured on the actual rock samples by Boyd & McCallister (1976).

Table 1. Normative parameters of five compositions†

	(1)	(2)	(3)	(4)	(5)
Gt (% by mass)	12.3	1.7	10.4	3.1	6.0
Cpx (% by mass)	16.3	5.6	16.7	2.0	4.5
Opx (% by mass)	13.5	17.7	9.5	39.7	22.5
Ol (% by mass)	57.9	75.0	63.4	55.2	67.0
$\hat{\rho}$ (Mg/m^3)	3.397	3.341	3.415	3.309	3.353
$\hat{v}_p$ (km/s)	8.265	8.277	8.217	8.300	8.290
$\hat{v}_s$ (km/s)	4.802	4.826	4.763	4.868	4.833

(1) Pyrolite (Ringwood 1966). (2) Pyrolite minus 20 mol % olivine basalt: basalt composition from Ringwood (1975, tbl. 4–2). (3) Kimberlite xenolith PHN1611 (Boyd & McCallister 1976). (4) Kimberlite xenolith PHN1569 (Boyd & McCallister 1976). (5) Average continental garnet lherzolite (Jordan 1979*b*).

† Parameters calculated with use of the garnet lherzolite norm of Jordan (1979*b*).

For our purposes the compositional control on the density of peridotites in the garnet stability field is adequately parametrized by two variables: $X_{\mathrm{Al}}^{\mathrm{WR}}$, the mole fraction of Al_2O_3 in the whole rock, and the molar ratio $R = X_{\mathrm{Fe}}^{\mathrm{WR}}/(X_{\mathrm{Fe}}^{\mathrm{WR}}+X_{\mathrm{Mg}}^{\mathrm{WR}})$. Within the range of compositions observed for garnet lherzolites, the former is approximately proportional to the molar amount of the dense ($\hat{\rho} = 3.7$ Mg/m^3) garnet phase, whereas the latter is approximately proportional to the mean atomic mass. Let ρ_0 be the density of some reference assemblage at a particular (not necessarily standard) pressure and temperature; then, to first order in small quantities, the density perturbation corresponding to specified changes in the compositional parameters is

$$\Delta\rho = \rho_0 \left(\frac{\partial \ln \hat{\rho}}{\partial X_{\mathrm{Al}}^{\mathrm{WR}}}\,\Delta X_{\mathrm{Al}}^{\mathrm{WR}} + \frac{\partial \ln \hat{\rho}}{\partial R}\,\Delta R\right). \tag{1}$$

The particular reference assemblage adopted here is that defined by Ringwood's (1966) pyrolite, whose normative parameters are listed in table 1 and for which

$$\frac{\partial \ln \hat{\rho}}{\partial X_{\mathrm{Al}}^{\mathrm{WR}}} = 0.70, \tag{2}$$

$$\frac{\partial \ln \hat{\rho}}{\partial R} = 0.32. \tag{3}$$

The fractionation of 20 mol % of olivine basalt from pyrolite gives $\Delta X_{\mathrm{Al}}^{\mathrm{WR}} = 0.017$ and $\Delta R = -0.024$. Substituting in equation (1) we find $\Delta\rho/\rho_0 = -1.9$ %, which compares reasonably well with the value of -1.7 % computed by the more elaborate algorithm used in table 1. The contributions of the two terms in equation (1) will generally be comparable for basalt subtraction from garnet lherzolites; in the example, the density changes are -1.1 % due to the decrease in alumina and −0.8% due to the decrease in the iron/magnesium ratio. For garnet-free lherzolites, the logarithmic derivative in the first term is much smaller and the second term dominates.

As first pointed out by O'Hara (1975) and Boyd & McCallister (1976), the dynamical effects of compositionally induced density variations can be quite large: the density decrease of 1.7% calculated for the 20% depletion of pyrolite is sufficient to offset a temperature decrease of about 500 °C. Peridotites depleted in basaltic constituents tend to float on undepleted mantle and can therefore be incorporated into the surficial c.b.l.

3. Plate structure

The lithosphere is usually defined as a mechanical boundary layer (m.b.l.) of high enough strength or high enough viscosity that its non-elastic behaviour can be ignored in dynamical equations (see, for example, Walcott 1970). Because the temperature rises rapidly with depth and the viscosity depends exponentially on inverse temperature, an isotherm can be used as a crude but convenient delineator of the lithosphere–asthenosphere boundary. Parsons & McKenzie (1978) adopted the 975 °C isotherm to specify the base of the m.b.l. in their discussion of oceanic plate structure; the rounder value of 1000 °C is used here.

In contrast to either the c.b.l. or the m.b.l. is the plate itself, or the tectosphere, most conveniently taken to be the surficial thermal boundary layer (t.b.l.), within which the strains are small and the dominant mode of vertical heat transport is conduction.

(a) *Oceanic tectosphere*

The thicknesses of the three boundary layers, chemical, mechanical and thermal, will generally differ at any particular point. In the oceans, for example, a c.b.l. comprising a basaltic crust and a subjacent depleted layer of dunitic and harzburgitic residua and cumulates is emplaced at the spreading centres (Ringwood 1969; Kay *et al.* 1970; Oxburgh & Parmentier 1977). This ophiolitic sequence, which has an estimated thickness of 30–35 km, translates away from the rise crests with little modification, whereas both the m.b.l. and t.b.l. thicken with age.

Out to an age of approximately 90 Ma the temperature profile of an oceanic plate appears to decay in the manner predicted by simple half-space cooling models; in particular, the water depth (corrected for variations in crustal loading) increases as the square-root of time (Davis & Lister 1974; Parsons & Sclater 1977). Beyond 90 Ma or so the water depth increases less rapidly with age, levelling towards an asymptotic value about 3500 m below the rise crests (Sclater *et al.* 1975; Parsons & Sclater 1977).

Various explanations have been put forward to explain this asymptotic behaviour; all postulate mechanisms that input heat from within or below the t.b.l. and thus retard the rate of t.b.l. growth. The mechanisms include radioactivity in the upper mantle (Forsyth 1977), shear-strain heating (Schubert *et al.* 1976), and the advection of heat from depth by either large-scale (Jarvis & Peltier 1980) or small-scale flow (Richter & Parsons 1975; Crough 1975; McKenzie & Weiss 1975). A particularly plausible model has recently been advanced by Parsons & McKenzie (1978) in which the plate thickness in old oceanic regions is limited by small-scale convective instabilities that develop at the base of the m.b.l.; these authors derive an approximate relation between the thicknesses of the m.b.l. and the (time-averaged) t.b.l. for the asymptotic steady-state régime.

The plate structure of thermally mature ocean basins is schematically represented in figure 1. An oceanic adiabat with a potential temperature of 1300 °C and a slope of 0.5 °C/km is assumed; the other parameters of the thermal profile are similar to those of Parsons & McKenzie (1978). The c.b.l. is defined by the deviation of the normative density profile below that of a mantle reference assemblage, here taken to be Ringwood's (1966) pyrolite ($\hat{\rho}$ = 3.40 Mg/m^3). Iron and alumina depletions of the mantle part of the c.b.l. are assumed to reduce the normative density by 0.06 Mg/m^3, equivalent to the removal of about 20 mol % of olivine basalt from pyrolite (table 1). The thicknesses of the c.b.l., m.b.l. and t.b.l. in an old ocean basin are estimated to be about 40 km, 100 km and 150 km, respectively.

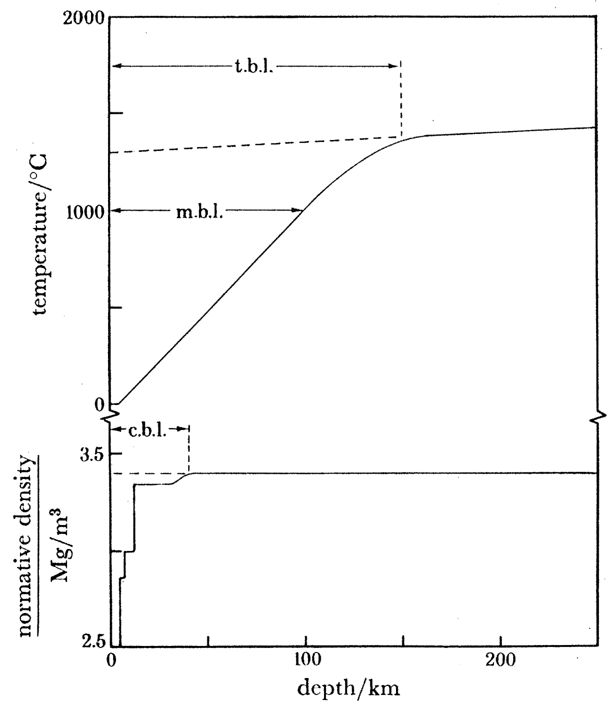

FIGURE 1. Tectospheric structure of an old ocean basin.

(b) *Continental tectosphere*

The simplicity of the oceanic t.b.l. models and their ability to satisfy the relevant geophysical observations have motivated attempts to extend them to the continents. Like the oceans, the continents show a systematic decrease in surface heat flow and elevation and an increase in t.b.l. thickness with 'tectonic age', provided that the age is defined not as the time since the original formation of the crust but as the time since the last major thermal or orogenic event (Polyak & Smirnov 1968; Sclater & Francheteau 1970). The thermal reactivation and subsequent conductive relaxation of a continental t.b.l. have been used to explain a wide variety of epeirogenic phenomena, including continental margin subsidence (Sleep 1971; Steckler & Watts 1978), the formation of continental basins (Sleep 1971; Sleep & Snell 1976; Haxby *et al.* 1976; McKenzie 1978), and plateau uplift (Crough 1979).

Within the late Proterozoic and Phanerozoic orogenic belts peripheral to the cratons, where most large vertical motions have occurred, the parameters of continental t.b.l. growth are compatible with those observed for the oceans; in particular, the asymptotic thickness of the epicontinental t.b.l. is similar to the oceanic value (Steckler & Watts 1978).

Less well established, however, is the character of the t.b.l. beneath the ancient cratons where Phanerozoic vertical motions have been limited. Three models of the cratonic tectosphere will be evaluated.

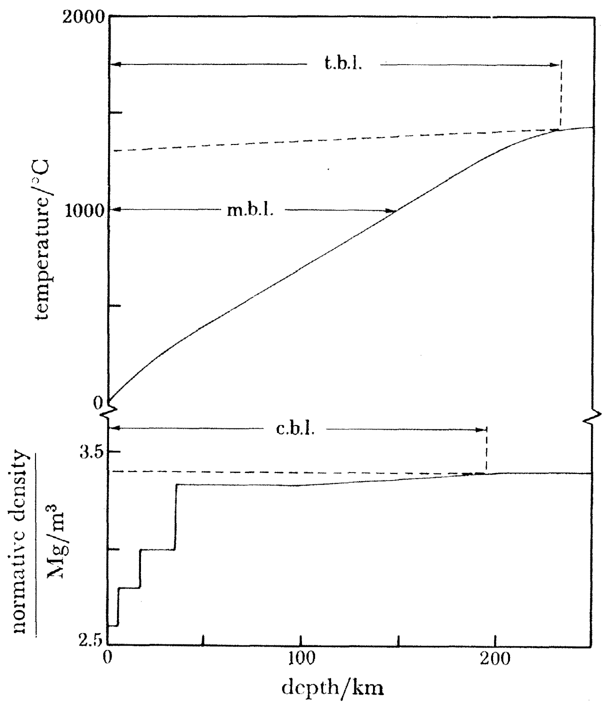

FIGURE 2. Tectospheric structure of an old continental craton according to model C. Thicknesses of the c.b.l. and t.b.l. are lower bounds to the average values.

Model A (Sclater *et al.* 1980)

Model A is a straightforward extension of the Parson–McKenzie oceanic model to the continents. Continents are viewed as superficial crustal edifices that modify the strength of the m.b.l. and, owing to their buoyancy, play an important role in plate tectonics, but they do not correlate with profound differences in tectospheric structure. The t.b.l. configuration of old continents is postulated to be similar to that of old ocean basins; following rejuvenation, both oceanic and continental t.b.ls grow to an asymptotic thickness of about 150 km, which is regulated in the steady-state by small-scale convective instabilities at the base of the m.b.l. The ancient cratons are the portions of the continental masses that for long periods have accidentally escaped any major thermal reactivation and subsequent collisional orogenesis.

Model B (Chapman & Pollack 1977)

Model B is based on the postulate that t.b.l. growth is not limited by the small-scale convective instability of the sub-m.b.l. tectosphere; unless perturbed by large-scale convective activity or collisional orogenesis continental thermal profiles continue their conductive decay indefinitely. Model B, like A, ascribes only a superficial character to the c.b.l. However, because the continental c.b.l. is buoyant, the thermal ages of the continents are typically much greater than

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those of the oceans and, hence, there exists a correlation between tectospheric thickness and crustal type. The cratons are simply those areas that have been tectonically stabilized by the growth of a very thick t.b.l.

Model C (Jordan 1975, 1978)

Model C assigns a major role to an augmented c.b.l. beneath the continental cratons. The subcrustal portions of the c.b.l. are presumed to consist primarily of peridotite depleted in major-element basaltic constituents (i.e. Al_2O_3, FeO, CaO) relative to the average upper-mantle composition, although magmatic cumulates (e.g. eclogites, wehrlites) may be ubiquitously distributed in small proportions. This refractory layer has had a complex history of development and, consequently, is extremely variable in thickness and minor-element composition. Beneath most orogenic zones and magmatic belts the c.b.l. is thin, and the t.b.l. thus behaves according to model A or model B. Beneath most cratons, however, the c.b.l. extends below the m.b.l. to depths of 150 km or more and acts to stabilize a thick (> 200 km) t.b.l. against small-scale convective disruption. Model C thus postulates a fundamental coherence between long-term surface tectonic behaviour and upper-mantle compositional structure. The tectospheric profile of a typical craton is schematically illustrated in figure 2.

(*c*) *Testing the models*

Although variants have been proposed (see, for example, Crough & Thompson 1976; Kono & Amano 1978; Oxburgh & Parmentier 1978; Crough 1979; Vitorello & Pollack 1980), the canonical models just described appear to encompass current geophysical speculation on continental tectospheric structure. Furthermore, each of the three models makes predictions testable with existing data.

Model A, for example, implies that the temperature and composition and thus the elastic parameters of the t.b.l. underlying old continents should be, on the average, the same as those of an old ocean basin; significant local variations may occur due to the thermal fluctuations associated with small-scale convective action, but these should not be a strong function of crustal type.

Seismologists have known for some time that the average shear velocity in the upper mantle beneath the cratons is considerably greater than the averages for oceanic and orogenic provinces (Brune & Dorman 1963; Toksöz & Anderson 1966). Originally based on fundamental-mode dispersion data, this inference has now been corroborated by observations of vertical delay times (Sipkin & Jordan 1975, 1976; Jordan 1979*a*) and higher-mode phase and group velocities (Cara 1979; Cara *et al.* 1980). Models for fundamental-mode and higher-mode propagation across northern Eurasia (Cara *et al.* 1980) show substantially greater (> 0.2 km/s) shear velocities in the depth range 100–200 km than do similarly derived models of the Pacific basin (Cara 1979), and differences as large as 0.1 km/s persist to depths exceeding 250 km. Taken at face value these results are not consistent with model A.

It may be argued, however, that such a comparison is an inadequate test of model A since the Pacific-crossing paths used by Cara sample young as well as old ocean basin. Indeed, the crustal ages along the New Hebrides – North America paths analysed by Cara (1979) range from 20 to 110 Ma, with an average age of about 80 Ma.

More diagnostic of the structure of a thermally mature oceanic plate are observations from the western Pacific, comprising some of the oldest oceanic crust now in existence.

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Fundamental-mode Rayleigh-wave data reveal a well developed low-velocity zone in the depth range 100–200 km (Yoshii 1975; Leeds 1975) not found beneath most cratonic regions (Knopoff 1972; Cara *et al.* 1980). The higher-mode data are limited, but the shear-wave travel times and attenuation along nearly vertical propagation paths have been studied extensively by means of multiple ScS waves with surface-reflexion points in the western Pacific (Sipkin & Jordan 1980*a*, *b*). The following results have been established.

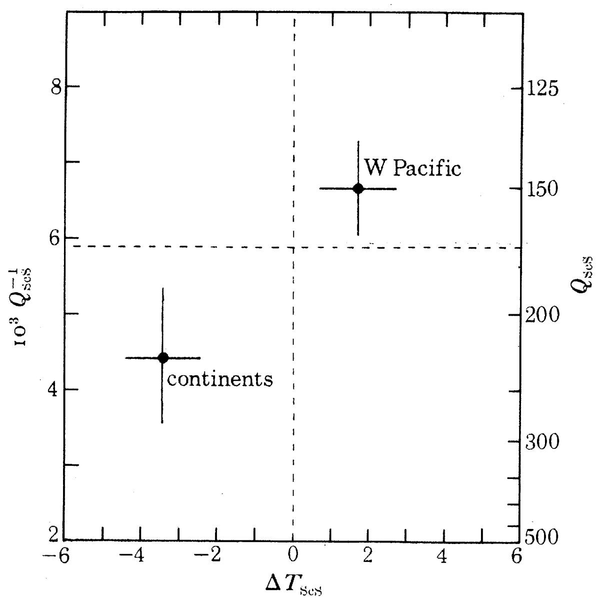

FIGURE 3. Attenuation parameter against travel-time residual for multiple ScS waves. Data from Sipkin & Jordan (1980*a*, *b*); dashed lines show values inferred for the spherically averaged Earth, whose ScS time (938.1 s) is used to calculate residuals; all times are corrected to an outer radius of 6371 km.

(i) Significant local variations in both ScS travel times (T_{ScS}) and attenuation factors (Q^{-1}_{ScS}) are observed for crustal ages greater than 100 Ma. Variations in T_{ScS} and Q^{-1}_{ScS} in the western Pacific correlate with one another, suggesting a thermal control (small-scale convection?), but not with crustal age; the average ScS parameters for paths with surface-reflexion points in Jurassic (*ca.* 160 Ma) crust are not significantly different from those observed for Cretaceous (*ca.* 100 Ma) paths.

(ii) When averaged over all crustal ages greater than 100 Ma T_{ScS} for the western Pacific is greater than the median continental value by 5.1 ± 1.4 s (the times are referenced to a surface radius of 6371 km). Corrections for known crustal structures increase this difference in two-way vertical travel time by at least 2 s. Therefore, contrasts in averaged upper-mantle structure must account for a *one-way* travel-time difference of approximately 3.5 s.

(iii) ScS waves propagating in the western Pacific are significantly more attenuated than those propagating in typical continental regions. Sipkin & Jordan (1980*b*) obtained best estimates of $Q_{ScS} = 155 \pm 11$ for the average of western Pacific paths and $Q_{ScS} = 225 \pm 45$ for average continent.

Results (ii) and (iii) are summarized in figure 3. The positive correlation between ScS travel times and attenuation factors, observed globally (Sipkin & Jordan 1980*b*), is important because it indicates that the travel-time variations are most likely due to thermal and/or compositional variations in the upper mantle and not to other structural effects (e.g. anisotropy or core–mantle interface topography). The contrasts in upper-mantle shear velocities implied by result (ii) are very large indeed; Cara's models for the Pacific and N Eurasia, cited previously, predict a one-way time difference of only +2.7 s, even though they are characterized by shear-velocity variations extending below a depth of 250 km. The ScS data suggest that the structural contrasts at these great depths may be more pronounced than Cara's models would indicate. Thus, despite the statements of Sclater *et al.* (1980) to the contrary, model A does not provide an adequate explanation of the seismic data pertaining to old continents and old ocean basins.

Model B, on the other hand, implies the existence of a thickened tectosphere beneath the ancient cratons, consistent with the inferences drawn from the seismology. Continued cooling of the cratons along an $(\mathrm{age})^{\frac{1}{2}}$ curve for, say, 1000 Ma is sufficient to grow a t.b.l. with a thickness greater than 300 km.

At least two objections can be raised against model B. Local Rayleigh numbers calculated for the sub-m.b.l. portions of such thickened t.b.ls are large ($> 10^4$), indicating that they are likely to be convectively unstable (Jordan 1975; Parsons & McKenzie 1978). This first objection may not be fatal, because it does not properly account for the strong dependence of viscosity on temperature, which acts to stabilize the t.b.l. against convective disruption (Yuen *et al.* 1981); the second undoubtedly is, however. To keep a cooling t.b.l. in isostatic balance, the crust must also thicken with time; this is, of course, the mechanism responsible for the increase in water depth with oceanic crustal age. The growth of a t.b.l. to a depth of 300 km should increase the thickness of the continental crust by 10–15 km, which is not observed (Jordan 1978; Oxburgh & Parmentier 1978). In fact, within the large tracts of Archaean and early Proterozoic rocks now exposed on the cratons the crustal thicknesses have not varied by more than a few kilometres during the last 2000 Ma or more (Watson 1976; Condie 1976).

The considerations that lead us to reject models A and B are precisely those that motivate model C. The compositional variations postulated by the basalt-depletion hypothesis provide a stabilizing mechanism for the thick cratonic tectosphere indicated by the seismic data (Jordan 1978).

The most direct tests of model C come from petrological data. The average composition of the oceanic upper mantle has been reasonably well established by geochemical and petrological studies on the origin of basaltic magmas (Ringwood 1975; Yoder 1976; Green *et al.* 1980) and spinel lherzolite inclusions in basalts (Kuno & Aoki 1970); Ringwood's (1966) pyrolite gives adequate estimates of the major oxides. For comparison, more-or-less random samples of the continental upper mantle are furnished by the xenoliths recovered from kimberlite pipes. Detailed analyses now exist for nodules from African, North American and Siberian pipes (see, for example: Nixon 1973; Sobolev 1977; Boyd & Meyer 1979). The normative parameters have been computed for a representative suite of 78 garnet lherzolite xenoliths by Jordan (1979*b*). Most of the xenoliths have equilibration pressures ranging from 3×10^9 to 6×10^9 Pa, indicating depths of origin between 100 and 200 km and a mean depth of sampling around 150 km (see Jordan (1979*b*) for references). With very few exceptions these rocks are depleted in basaltic constituents with respect to pyrolite; they have lower values of Fe/(Fe+Mg) and lower normative clinopyroxene and garnet. Only one sample (the unusual nodule PHN1611) has a normative density exceeding that of pyrolite and 60 of the 78 have normative densities lower by more than 1%. Listed in table 1 are the parameters for an average continental garnet lherzolite (a.c.g.l.) constructed by taking the mean of all 78 oxide compositions. The a.c.g.l., whose normative mineralogy closely approximates the estimates of Mathias *et al.* (1970) for the South African xenolith population as a whole, has a normative density 1.3 % below pyrolite.

These data suggest rather strongly that an augmented c.b.l. underlies the cratons to depths of at least 150 km, in accord with the basic premise of model C. In fact, the existence of such a refractory layer has been recognized by petrologists for many years (Ringwood 1966; Nixon *et al.* 1973). It is the confluence of geophysical and petrological thinking on the subject of continental deep structure that most convinces the author that model C is viable.

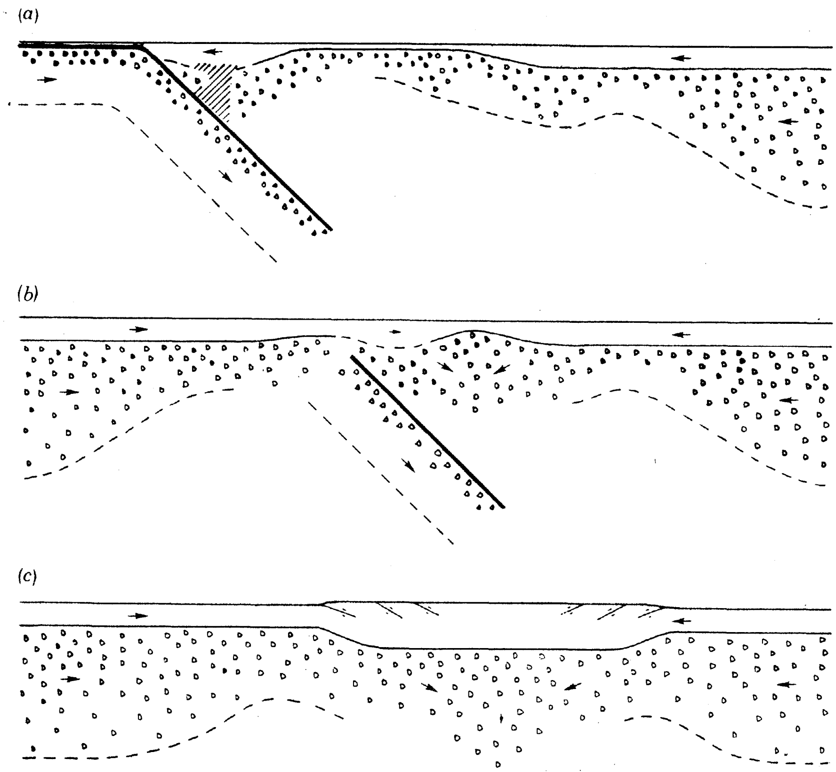

FIGURE 4. Schematic illustration of how a thin, inhomogeneous c.b.l. along an active continental margin (*a*) could be consolidated and thickened by continent–continent collision (*b*) and subsequent compressive orogenesis (*c*). Dotted pattern is low-density mantle depleted in basaltic constituents, cross-hatching shows region of partial melting, and arrows indicate relative motions.

4. Evolution of the c.b.l.

Given the variety of tectonic and magmatic processes that can modify tectospheric structure and the few constraints available on subcrustal tectospheric evolution, there is considerable latitude for speculation on just how a thickened continental c.b.l. might be built up. A tentative model has been presented by Jordan (1978). Vulcanism along the active continental margins, and to a lesser extent in intracontinental rifts, presumably generates buoyant, refractory mantle resistant to convective recycling. Augmenting this material is any depleted peridotite transferred from the oceanic c.b.l. by lateral accretion across subduction zones or, as suggested by Oxburgh & Parmentier (1977, 1978), by diapirism from descending slabs. Dispersed regions of buoyant mantle and their superjacent (but not necessarily chemically complementary) crustal columns are amalgamated during the course of plate motions and eventually swept up and accreted to the major continental masses. Major compressive events at convergent plate

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boundaries, particularly the violent episodes of continent–continent collision, further consolidate and thicken the c.b.l. by the downwards advection of basalt-depleted mantle (figure 4). Competing against these constructive processes are the destructive instabilities associated with tectospheric heating; during episodes of continental extension, for example, heat is advected from depth and the c.b.l. is thinned by lateral dispersal, occasionally initiating Atlantic-type drift. Ultimately, however, perhaps only after repeated orogenic cycles, a thick, cohesive and relatively stable c.b.l., a craton, is formed.

Such a scenario, though hypothetical, conforms to actualistic principles and illustrates that a thickened c.b.l. is a plausible consequence of the Wilson cycle. Moreover, as an evolutionary model it possesses a number of predictive and potentially testable ramifications.

The model implies that primary episodes of basalt extraction precede cratonization; hence, the c.b.l. structure of an ancient craton as regards major elements (though not necessarily minor ones) should date from its stabilization. Pertinent data are available from South Africa. Basalt depletion of the South African upper mantle evidently pre-dates the emplacement of the Premier kimberlite pipe (minimum age ≈ 1100 Ma) whose xenolith suite is remarkably similar to those from Cretaceous pipes (Danchin & Boyd 1976; Danchin 1979). A more stringent bound is given by the isotopic dating of mantle materials. Clinopyroxenes from various types of kimberlite xenoliths yield model lead ages in excess of 2000 Ma, as do sulphide inclusions in diamonds (Kramers 1979). According to Kramers, 'Pb isotopic heterogeneities have persisted in local closed systems in the upper mantle underneath Southern Africa for approximately 2.5 b.y. [2.5 Ga]', i.e. since the stabilization of the Kaapvaal craton in late Archaean – early Proterozoic time (Hunter 1974).

Because its easily fusible components have been largely extracted by previous magmatic episodes, the mantle part of the continental c.b.l. is not likely to be the site of further large-scale melting; therefore, it constitutes a particularly stable, long-term reservoir for large-ion lithophile (l.i.l.) chemical species (Jordan 1978). As this reservoir translates across the Earth's surface by plate motion, it becomes impregnated by volatiles and highly differentiated liquids fluxing from the deep interior or from descending slabs. Over the course of aeons this flux substantially modifies the l.i.l. chemistry of the c.b.l., introducing isotopic heterogeneities and occasionally triggering intracratonic vulcanism. A growing body of petrological and geochemical evidence is consistent with extensive metasomatic alteration of the c.b.l. (see, for example: Lloyd & Bailey 1975; Harte *et al.* 1975; Boettcher *et al.* 1979), including recent strontium and neodymium studies that point to large, long-standing isotopic heterogeneities (Basu & Tatsumoto 1979, 1980; Menzies & Rama Murthy 1980; Stosch *et al.* 1980).

Metasomatism of the c.b.l. introduces heat producing elements to the continental tectosphere that can make significant contributions to surface heat flux. Granular garnet peridotites from intracratonic environments, although depleted in their major-element chemistry, have high (and variable) potassium contents correlative with high rubidium and other l.i.l. contents (Shimizu 1975; Rhodes & Dawson 1975). The a.c.g.l. of table 1, for example, contains 0.11% (by mass) K_2O (K *ca.* 0.9 mg/g), a factor of 37 greater than the estimate of Green *et al.* (1980) for the source of mid-ocean-ridge basalts. Isotopic studies of kimberlite xenoliths demonstrate that such high l.i.l. concentrations have existed in these rocks for some time and do not involve appreciable contamination by the kimberlitic host magma (Barrett 1975; Kramers 1977; Basu & Tatsumoto 1979, 1980). Assuming that K/U = 3000 and K/Th = 1000, ratios consistent with the empirically determined values of Wakita *et al.* (1967), we find a heat generation rate for the a.c.g.l. of 0.19 μW/m³. This value compares favourably with a heat generation rate of 0.16 μW/m³ calculated for a single garnet harzburgite xenolith from Kimberley whose U, Th and K abundances were measured by Wakita *et al.* (1967, sample no. 19).

The high radioactivity inferred for the metasomatized c.b.l. has important implications for the thermal structure of the continental tectosphere. The heat flux from the mantle beneath the stable cratons is thought to be less than 30 mW/m² (Sclater *et al.* 1980; Vitorello & Pollack 1980).† A metasomatized layer between the depths of 100 and 200 km with an average heat production of, say, 0.15 μW/m³ would contribute 15 mW/m² to this value, leaving less than 15 mW/m² as the contribution from the subtectospheric mantle. A similar conclusion has been derived independently by Jordan (1975, 1978), who showed that a low background heat flux is required to satisfy the equilibration temperatures and pressures inferred for kimberlite xenoliths. It implies that tectospheric temperatures are probably somewhat lower and the t.b.l. somewhat thicker than indicated by the cratonic profile of figure 2, putting the model in better agreement with the seismic data discussed earlier but in more serious conflict with the conclusions of Sclater *et al.* (1980).

5. Discussion

Variations in the chemical characteristics and depth of the continental c.b.l. can perhaps account for the tectonic behaviour of cratons and mobile belts and explain why the present-day boundaries between these features have, in many cases, persisted since mid-Proterozoic time (Watson 1976): the repeatedly active mobile belts and orogenic zones mark regions where a deep c.b.l. has never stabilized, whereas the Archaean cratonic nuclei are the areas where a deep c.b.l. was stabilized at an early date. Thus, the formation of an augmented c.b.l. provides a plausible mechanism for cratonization or, to borrow Sutton's (1963) term, 'chelogenesis'.

The dynamics of a thickened c.b.l. are far from being understood, however, and it is with respect to the model's dynamical feasibility that certain caveats should be stated. If the depth to the base of the c.b.l. varies laterally by 150 km or more, as advocated here, then convective instabilities of the double-diffusive type (Turner 1973) must occur. Stevenson (1979) has done the linear stability analysis for the problem of a constant-viscosity fluid initially configured to have both horizontal and vertical compositional gradients whose effects on density are exactly compensated by temperature gradients. Assuming a c.b.l. thickness of 200 km, a gradient of 10^{-8} m^{-1} (i.e. a 1% density change in 1000 km) and a kinematic viscosity of 3×10^{17} m²/s, he finds that the fastest growing instability has a growth time of *ca.* 200 Ma and a horizontal length scale of *ca.* 500 km. Small-scale double-diffusive instabilities of this sort should act to disperse and thin the continental c.b.l.; in fact, they may be responsible for, or at least contribute to, episodes of continental rifting and break-up. However, if these instabilities evolve rapidly as Stevenson's preliminary analysis suggests, then a thickened c.b.l. cannot be easily maintained for the thousands of millions of years required by the model.

The substantiation of this argument would constitute grounds for rejecting the model, but further work on the dynamical problem is needed. Stevenson's (1979) growth-time calculation does not include several effects that may tend to dampen double-diffusive instabilities. First, a m.b.l. resistant to deformation occupies the upper part of the c.b.l.; the c.b.l. probably extends

† The mantle heat flux may, in fact, be significantly less than this value if there is any appreciable heat generation in the intermediate and lower crust; see, for example, Smithson & Decker (1974).

no more than 50–100 km below its base (figure 2). Secondly, beneath the m.b.l. the viscosity structure is probably not uniform; because it depends on both temperature and the degree of basalt depletion, the viscosity should be greatest in regions where the c.b.l. is thickest (Jordan 1978). Thirdly, the vertical compositional gradients are generally greater than horizontal gradients by as much as an order of magnitude; the sub-m.b.l. portions of the c.b.l. could actually be stably, rather than neutrally, stratified.

Complementing these problems of small-scale instability are the questions of large-scale dynamics. One of the most curious aspects of continental tectonics is the tendency for continents to congregate close to one another to form 'supercontinents'. The evidence for this behaviour comes from both geological and palaeomagnetic data; the reader is referred to Irving's (1980) maps for post-Devonian reconstructions and to Piper (1976), McElhinny & McWilliams (1977) and Windley (1977) for summaries pertinent to earlier periods. Continental dispersal of the present magnitude appears to be an exception, not the rule. The continents are gregarious. As illustrated in figure 4, the fusion of continental masses into supercontinents and the orogenic compression that this entails may be important events in the formation of a thick continental c.b.l. Although the gregarious behaviour of the continents before the recent episodes of drift does not mean that plate tectonics was inoperative in previous epochs, as some authors have supposed, it may require some sort of special large-scale dynamics to keep the continents grouped on the same or nearby plates. If so, perhaps the forces involved are related to the existence of the continental c.b.l.

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