On the emergence of plate tectonics

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ABSTRACT

At present, young oceanic lithosphere is positively buoyant, and it does not become negatively buoyant until it is older than about 20 m.y. If, in the past, the mantle was hotter, the oceanic crust would have been thicker and the lithosphere's age of neutral buoyancy greater. On the other hand, the hotter mantle would have had a lower viscosity and convected faster, so the average age of oceanic plates at subduction would have been less than the present 100 m.y. At some time in the past, average oceanic lithosphere would have only just become negatively buoyant as it reached a subduction zone. It is estimated here that this condition occurred when the mantle was only about 50 °C hotter than at present and as recently as 0.9–1.4 b.y. ago. Transformation of basaltic crust to eclogite might have enhanced the vigor of the plates, but possibly only modestly and intermittently. In earlier times, plate tectonics could have operated, but more slowly, so that it could not by itself have accomplished the necessary rate of heat removal from the earth. A different tectonic mode must also have operated, perhaps subcrustal delamination or dripping. Plate tectonics would have gradually taken over from the earlier mode. Plumes are a complementary mode of mantle convection driven by a lower thermal boundary layer, and so they could have operated in conjunction with both regimes of the upper boundary layer.

INTRODUCTION

Although the signature of plate tectonics is recognized with some confidence in the Phanerozoic geologic record of the continents, its action becomes less certain further back in time. There was considerable debate until about a decade ago about whether plate tectonics might have operated in the Archean (e.g., Windley, 1977). More recently, plate-tectonics interpretations of the Proterozoic and Archean have become more common (e.g., Hoffman, 1980; Cassidy et al., 1991). However, Taylor (1987) has argued for a major transition between the Archean and Proterozoic regimes, and Campbell and Hill (1988) have argued explicitly that plumes may have played a major role in Archean tectonics. The possibilities that plate tectonics operated during the Archean and that it was the dominant tectonic regime would seem to be still quite uncertain.

If Earth's interior has been cooling steadily, as is widely believed, then there is a good reason why plate tectonics might not have worked in early times. A hotter mantle would have produced more melt at spreading centers and hence thicker oceanic crust. Since oceanic crust is less dense than the mantle, it makes the oceanic lithosphere more buoyant and harder to subduct. A hotter mantle would also convect faster, so that the oceanic lithosphere would have less time to thicken and become gravitationally unstable. Both factors would tend to inhibit the operation of plate tectonics.

The idea that early plates would have subducted less readily because they were more buoyant has been discussed before (e.g., Baer, 1977; Arndt, 1983; Nisbet and Fowler, 1983), but it does not seem to have been examined quantitatively in the context of current understanding of how plates, plumes, and mantle convection work (e.g., Davies and Richards, 1992; Campbell and Griffiths, 1992). Earlier discussions were based on assumptions that now seem implausible. For example, Baer (1977) and Nisbet and Fowler (1983) assumed an implausibly hot Archean mantle and relied on the basalt-eclogite transition to initiate and maintain subduction, and Arndt (1983) assumed that the thickness and composition of the oceanic crust are independent parameters.

One important point concerns the inferred temperature of the Archean mantle. Campbell and Hill (1988) proposed that komatiites are produced from mantle plumes, and Campbell and Griffiths (1992) have stressed that the high source temperatures of komatiites would not then be representative of normal mantle. They have inferred from the occurrence and abundance of Archean basalts that the normal Archean mantle was only 50–100 °C hotter than the present mantle. Finally, it does not seem to have been considered that plates might have operated in tandem with earlier regimes and that they have become dominant only over an extended period. These possibilities are explored quantitatively here.

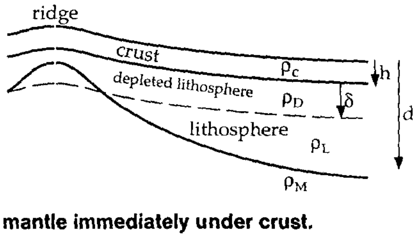

Figure 1. Sketch of oceanic crust and lithosphere showing definitions of densities and thicknesses used in this paper; ρ_L is mean density of normal-mantle part of lithosphere, and ρ_D is density of layer of depleted mantle immediately under crust.

BUOYANCY OF OCEANIC LITHOSPHERE

The oceanic lithosphere comprises a cool thermal boundary layer, which thickens with time by conduction of heat to the sea floor, and a low-density crust about 7 km thick. Because the thermal boundary layer has an error-function temperature profile whose depth scale increases with time, its mean density, averaged over this depth scale, is constant. Thus, the effect of thermal contraction can be represented by a layer with a constant mean density and a thickness that increases in proportion to the square root of time. This situation is illustrated schematically in Figure 1.

With the densities and thicknesses denoted as shown in Figure 1, the mean density, ρ_P, of the plate is

$$\rho_P = [\rho_c h + \rho_L(d - h) + (\rho_D - \rho_L)\delta]/d. \tag{1}$$

The plate will be neutrally buoyant when its density is equal to the density of the underlying mantle, ρ_M, from which the following condition can be derived:

$$d = h[(\rho_L - \rho_c) + (\rho_D - \rho_L)\delta]/(\rho_L - \rho_M). \tag{2}$$

Ringwood and Irifune (1988) said that the density deficiency of the depleted mantle layer from which the crust is extracted is as much as 50–80 kg/m^3 and that thickness of the layer is about 20 km. Here a mean density deficit of 30 kg/m^3 is assumed. This residual layer is quite significant, contributing about one-third as much buoyancy as the oceanic crust.

The thickness of the thermal boundary layer as a function of age, τ, is

$$d = \gamma\tau^{1/2}, \tag{3}$$

where γ is a constant (value given in Table 1). Eliminating d from equations 2 and 3, the age at which the plate reaches neutral buoyancy is

$$\tau_n = \beta h^2, \tag{4}$$

where β is another constant, given by

TABLE 1. VALUES USED IN CALCULATIONS

Symbol	Description	Value
γ	Plate-thickening constant	10 km/m.y.$^{1/2}$
ρ_c	Crust density	$\rho_{c0} + c(T - 1280)$
ρ_{c0}	Present density of oceanic crust	2990 kg/m^3
c	Crust density coefficient	0.4 kg/m^3 K
ρ_L	Lithosphere density	$\rho_M(1 + \alpha\Delta T)$
ρ_D	Depleted-mantle density	ρ_L – 30 kg/m^3
ρ_M	Mantle density	3300 kg/m^3
α	Thermal expansion coefficient	3×10^{-5} K^{-1}
ΔT	Mean temperature deficiency of lithosphere	700 °C
h_0	Present thickness of oceanic crust	7 km
b	Crust-thickening coefficient	0.085 km/K
d	Thickness of depleted-mantle layer	$20 + 0.2\Delta T$ km
A	Area of sea floor	3×10^{14} m^2
S	Areal sea-floor–spreading rate	2.5 km^2/yr
μ	Oceanic heat flux at 1 Ma	0.55 W/m^2
m	Heat-flow temperature exponent	10–13
λ	Heat-flow decay constant	(2–3 b.y.)$^{-1}$

$$\beta = \left[\frac{(\rho_L - \rho_c) + (\rho_D - \rho_L)\delta}{(\rho_L - \rho_M)\gamma} \right]^2 . \tag{5}$$

With the values given in Table 1, the present value of τ_n is 22 m.y.

THICKNESS OF THE OCEANIC CRUST

At present the oceanic crust is about 7 km thick. The thickness as a function of mantle temperature has been estimated by White and McKenzie (1989). Their results can be approximated as

$$h = h_0 + b\,(T - 1280), \tag{6}$$

where T is the potential temperature of the mantle (temperature of adiabatically decompressed mantle). In equation 6, h_0 and b are constants (values given in Table 1). White and McKenzie also estimated the density of the oceanic crust as a function of mantle temperature, and this result is represented by

$$\rho_c = \rho_{c0} + c\,(T - 1280), \tag{7}$$

where values for the constants are also given in Table 1.

AGE OF PLATES AT SUBDUCTION

At present, the mean age of plates at the time they subduct is about 100 m.y. (Parsons, 1982). If the mantle were hotter, its viscosity would be lower and it would convect faster, other things being equal (Davies, 1980). Because at present the plates are an integral part of mantle convection and the cycle of plate formation, cooling, subduction, and reheating is the principal means by which heat is removed from the mantle (Davies and Richards, 1992), the rate of plate creation and destruction would be greater. From these factors a relation between mean age at subduction and mantle temperature can be derived.

The present areal rate of plate production, S, is about 2.5 km^2/yr (Parsons, 1982), and the total area of sea floor, A, is about 3×10^{14} m^2. Assuming steady state, the mean age of the sea floor at the time it subducts is thus about 120 m.y.:

$$\tau_s = A/S. \tag{8}$$

This value is somewhat larger than the estimate by Parsons, suggesting that the present value may not represent the long-term steady state. The heat flux through the sea floor is

$$q = \mu\tau^{-1/2}, \tag{9}$$

where μ is a constant. The total heat flow, Q, is obtained by summing over all ages and multiplying by S:

$$Q = 2S\mu\tau_s^{1/2}. \tag{10}$$

Substituting for S from equation 8 and rearranging,

$$\tau_s = (2A\mu/Q)^2. \tag{11}$$

This expresses the first main factor: that the higher the heat flow, the faster the plates move and hence the sooner the plates subduct.

I have discussed the relation between mantle temperature and heat flow (Davies, 1980) and have shown that, for conditions not too different from present conditions, it can be approximated as

$$Q = Q_0(T/T_0)^m, \tag{12}$$

where subscript 0 denotes present values and m is a constant with a value of about 10 to 13. The constant m depends on the activation energy of the mantle material and on the exponent in the Nusselt number–Rayleigh number relation.

PLATE SUBDUCTIBILITY VS. MANTLE TEMPERATURE AND TIME

Equations 11 and 12 express the mean age at subduction as a function of mantle temperature, whereas equations 4–7 express the age of neutral buoyancy as a function of mantle temperature. These quantities are plotted in Figure 2 for different values of m and crustal density ρ_c, chosen so that the range of results is representative of the total uncertainty. It can be seen that for mantle temperatures only about 50 °C above the present temperature, the time of neutral buoyancy is greater than the mean age of subduction: plates, on average, would not be ready to subduct when they arrived at a trench. This is a contradiction that implies that something else must have been happening instead.

This result has been expressed initially in terms of mantle temperature because the relation between mantle temperature and time involves some additional uncertainty. Calculations of thermal evolution yield mantle cooling rates of 50–100 °C/b.y. (Jackson and Pollack, 1984). Except for early in Earth's history, the heat flow tracks the decline in radioactive heat production in the mantle with an approximately exponential behavior:

$$Q = Q_0e^{\lambda t}, \tag{13}$$

where t is age and $1/\lambda$ is 2–3 b.y. From equation 12, the temperature will decline as

$$T = T_0e^{\lambda t/m}. \tag{14}$$

Equation 14 is used to convert from temperature to time. Calculations like those of Jackson and Pollack (1984) implicitly assume that the convection rate is controlled by the mantle viscosity. Christensen (1985) questioned this, but Gurnis (1989) showed that his results were due to the immobility of the upper boundary layer in his calculations, which contrasts with the mobility of Earth's lithosphere.

The resulting plate age vs. time curves are shown in Figure 3 for two values of λ. It can be seen that the crossover occurred between about 0.9 and 1.4 Ga. This range is representative of the variation arising from the uncertainty in other parameters, such as the crustal density, the thermal

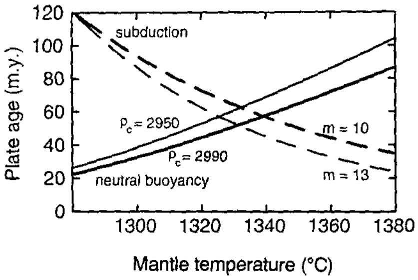

Figure 2. Average age of plates at time of subduction and age at which plates become neutrally buoyant as functions of mantle temperature, taking present mantle temperature to be 1280 °C. Heavy curves are for preferred values, and light curves are for plausible alternative values, chosen to illustrate likely total uncertainty; *m* is exponent in heat flux–temperature relation, equation 12, and ρ_c is present oceanic crustal density, in kg/m^3.

expansion coefficient, and the thickness of the crust as a function of mantle temperature (equation 6).

IMPLICATIONS FOR PAST PLATE ACTIVITY

The conditions under which the plates would not have routinely sunk at subduction zones are surprisingly close to present conditions in terms of both mantle temperature and time. What would have happened prior to this? It would still have been possible for oceanic lithosphere to subduct, but it would have required longer for the lithosphere to mature to the point where it would be negatively buoyant. Thus, plates could still have formed and functioned in the modern way, but they would not have been going fast enough to remove heat at the rate "expected" of them.

Two possibilities are (1) heat was removed in some other way or (2) the mantle was not being cooled as efficiently as at present. If the present understanding of Earth's thermal regime is correct, the rate of radioactive heat production is not less than about 70% of the present rate of heat loss (Jackson and Pollack, 1984), and this ratio would have applied through the past 1 or 2 b.y. Thus, the efficiency of heat loss cannot fall by more than about 30% without implying that the mantle was heating up. If we go far enough back, to the Archean, for example, when radioactive heat production was perhaps three times its present level, the need for an alternative cooling mechanism seems to be inescapable.

Before discussing what the alternative might be, we can look at the amount of heat that plates could have removed before the crossover time depicted in Figure 3. The subduction age would have been constrained to be the same as the time to reach neutral buoyancy, which would have been longer in earlier times when the oceanic crust was thicker. As a result, the heat removed by plate tectonics would have been given by

$$Q = 2A\mu\tau_n^{-1/2}. \tag{15}$$

The variation of the heat removed by plates, both before and after the crossover, is depicted in Figure 4. Plate tectonics might have been even less important than implied by Figure 4 if the maturation of the plate was interrupted by other processes, as seems likely in the light of the discussion to follow.

Included in Figure 4 is an estimate of the variation of the radioactive heat-generation rate, assuming that at present it is 75% of the heat-loss rate (Jackson and Pollack, 1984). According to this estimate, during the Archean the plates could have removed only a fraction of the heat generated.

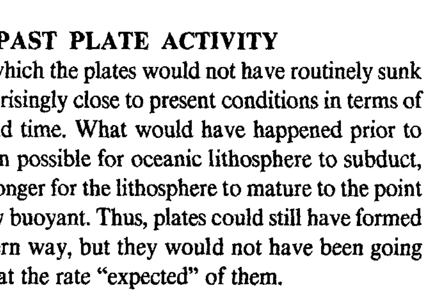

Figure 3. Subduction age and neutral buoyancy age vs. time before present, in same format as Figure 2. Conversion from mantle temperature to time was made using equation 13. Curves from two values of λ (lambda) are shown, illustrating likely uncertainty.

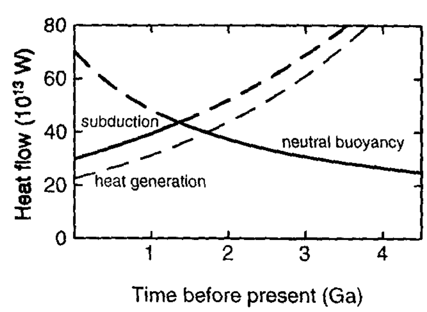

Figure 4. Heat flow carried by plates (solid curve segments) as function of time before present, according to ideas developed in this paper. Before about 1.8 Ga, heat transported by plates is limited by buoyancy of oceanic lithosphere. After that time, it is limited only by efficiency of plate-mantle convection system. Heat production is also shown (lighter dashed line), assuming decay constant of ⅓ b.y.$^{-1}$.

PREPLATE TECTONICS

Deducing what form of tectonics might have existed before plate tectonics is not straightforward, because the interactions among decompression melting, conductive cooling, and buoyancy are potentially complex. Regardless of such difficulties, I emphasize that "plume tectonics" is *not* an alternative to plate tectonics. Mantle plumes most plausibly arise from a lower, hot thermal boundary layer, whereas plates arise in the upper, cold thermal boundary layer. Plumes and plates are thus complementary, rather than being alternatives. What we are looking for is an alternative mode of behavior of the *upper* boundary layer.

Two possible tectonic regimes are noted here. First, if the crust were not too thick, so that conductive cooling could penetrate through it to a substantial degree, the mantle part of the thermal boundary layer might still be quite strong and platelike. In that case, the behavior might not be very different from plate tectonics (see Fig. 5A). In effect, it would be as if plates operated, but the thicker oceanic crust was scraped off before subduction. This is essentially what is proposed to happen at present when overthickened oceanic crust reaches a subduction zone. Something like this was proposed by Bird (1979) for the modern Basin and Range province of North America and by Park (1981; his "mantle decoupling") to account for a predominance of horizontal fabrics in Archean high-grade gneiss terranes.

On the other hand, if the mantle were hotter and the crust thicker, then the mantle part of the boundary layer might be too thin and soft to be platelike, and something closer to conventional constant-viscosity, nonplate convection might ensue. In this regime (Fig. 5B), the sinking boundary-layer behavior is more likely to be "drip" style than the asymmetric subduction characteristic of plate tectonics. Something like this process has been proposed for some modern continental areas (Houseman et al., 1981) and for the Early Archean by Campbell and Griffiths (1992), who have quantified some aspects of it and have proposed some important implications for the chemical evolution of the mantle. Earlier, loosely related ideas are the "viscous drag" proposal of Hargraves (1978) and a sketch of symmetric downwelling by Smith (1981). A deficiency of some discussions is a tendency to regard mantle convection as an agent independent of the lithosphere, rather than as involving the lithosphere (or part of it) intrinsically as the source of negative buoyancy driving the convection.

Whether the regimes of Figure 5 ever existed and when they might have existed are unclear at this stage. The timing is uncertain because the uncertainty of the tectonic regime translates into uncertainty in the rate of cooling of the mantle. Some possible regimes that might have existed with an even hotter mantle have been discussed by Davies (1990).

Observational tests will be necessary to narrow the range of possibilities. There is a large literature considering the possible tectonic and magmatic evolution of crustal piles such as those sketched in Figure 5, detailed

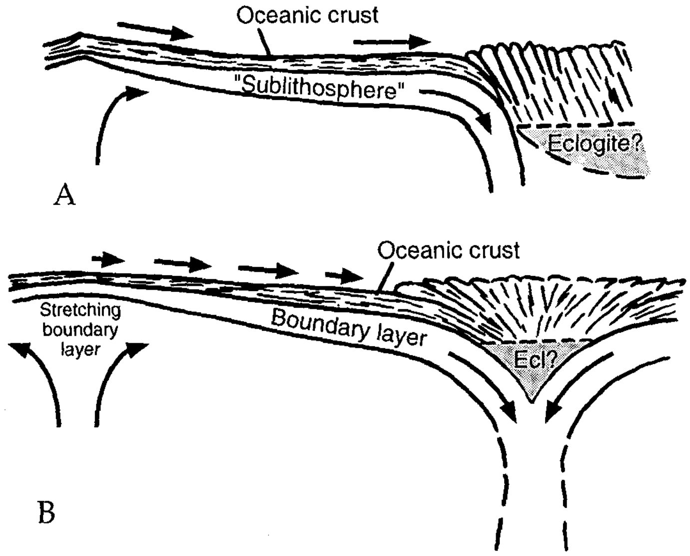

Figure 5. A: Sketch of conjectured regime preceding plate tectonics proper, in which mantle part of oceanic lithosphere is thick enough to behave rigidly, but whole mantle-plus-crust package is buoyant; lower, mantle part is negatively buoyant and subducts by shearing off from crustal part, which accumulates at surface. Detachment of crust would be aided by higher temperatures that would exist at base of crustal part. This regime resembles one proposed by Park (1981). Crustal pile might differentiate internally and come to resemble an island arc, especially if its deeper parts transformed to dense eclogite that would sink. B: Sketch of conjectured preplate regime in which thin, soft, thermal boundary layer forms under thick basaltic crust and "drips" back into deep mantle. Crust would accumulate over downwellings and might also come to resemble island arcs and to have eclogite (Ecl) sinkers, as suggested for regime of A. This kind of regime has also been proposed by Campbell and Griffiths (1992), who quantified some aspects of it and noted some possible implications for chemical evolution of mantle.

examinations of preserved Archean assemblages, their possible resemblances to modern island arcs, and their possible relation to plate tectonics. Recent examples are provided by de Wit et al. (1987) and Nutman and Collerson (1991).

One important distinction that might have observable structural consequences is that between subduction and "dripping." Subduction is intrinsically asymmetric. It results from the essentially brittle behavior of the lithosphere, which yields by faulting. Horizontal compression is accommodated by reverse faults, and a necessary consequence is that with continued convergence the footwall can descend indefinitely, while the hanging wall is uplifted by a finite amount (determined by the gravitational restoring force). On the other hand, dripping may be symmetric. If the material is in a ductile mechanical regime where faulting does not occur, horizontal compression could result in both sides of a convergence zone being forced downward.

One possible escape from the conclusion that plates could not have worked in the past would be if the transformation of the basaltic component to eclogite at depth (Green and Ringwood, 1967) could outweigh its low density at shallower depths. However, the feasibility of this possibility is not obvious. The density excess of eclogite is about 80 kg/m^3 (Ringwood and Irifune, 1988), whereas the density deficit of basalt is about 300 kg/m^3, so a substantial body of eclogite would be needed. Then there is the question of the kinematics of the transformation, which may require the basalt to be heated significantly or the pressure to be substantially overdriven. Large bodies like those depicted in Figure 5 will heat only slowly from around their edges. It is quite possible that the effect of the basalt-eclogite transformation could have increased the rate of plate tectonics, but perhaps not dramatically and only intermittently.

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