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PRECAMBRIAN PLATE TECTONICS: A GEODYNAMIC APPROACH

by

ELSIE JOY WILLIAMS, B.Sc.

A thesis submitted to
the Faculty of Graduate Studies and Research.

in partial fulfilment of
the requirements for the degree of
Master of Science

Department of Geology

Carleton University

Ottawa, Ontario

April 1986

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in partial fulfillment of the requirements for
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ABSTRACT

The Archean geologic record is discontinuous, poorly exposed and largely uncorrelatable, which has resulted in controversy regarding the style of Archean tectonics. Plate tectonics assumes that heat is dissipated through conductive continents and spreading oceans. Justification of plate tectonic models in the Archean is sought by examining the dissipation of terrestrial heat as a function of time.

The mean age of the oceanic crust at subduction was 30 Ma (potassium-depleted earth composition) or 10 Ma (chondritic earth composition) in the early Archean, corresponding to rates of creation of oceanic area (spreading activity) 5 and 17 times the present, respectively. These results are independent of the chosen continental geothermal gradient, continental accretion model, and mantle rheology.

Archean plate tectonics is viable for a potassium-depleted bulk earth model. The chondritic bulk earth model requires a denser oceanic crust in the Archean to permit negative-buoyancy subduction; however, the considerably greater spreading activity reduces the likelihood of plate tectonics.

These results do not exclude other models of Archean tectonics.

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INTRODUCTION

1

The principle of uniformitarianism states that the present is the key to the past. This uniformity of physical law is reasonably extrapolated through geologic history. However, strict actualism, the idea that the present is the same as the past, may be dismissed (Nisbet, 1985) if for no other reason than the significantly higher Archean heat flux. The problem becomes increasingly important the further into earth history the geologist looks, culminating in the problem of Archean geology. The theory of plate tectonics necessarily raises the question of tectonic actualism. Much debate has centred on this topic. The applicability of plate tectonics to Proterozoic rocks is generally accepted by the geologic community, but the question of Archean tectonics remains open. In this thesis the uniformity of plate tectonic processes will be examined from a geodynamic viewpoint.

Two possibilities exist, either modern plate tectonics, hereafter referred to as mobilist tectonics (after Blackburn, 1980), was operative throughout earth history, or some mechanism of fixist tectonics was predominant, before some critical time. In order to test the validity of the extrapolation of mobilist tectonics to the Proterozoic and Archean, it is herein assumed that mobilist tectonics was operative, viz., the earth's surface was characterized by spreading oceans and subducting oceanic crust.

The assumption of mobilist tectonics in earth history permits the calculation of the relevant tectonic parameters: velocity of plate motion (spreading), length of the spreading

ridge, mean age of subduction of oceanic lithosphere and the mean thickness of oceanic lithosphere at subduction. Since the geologic and geodynamic constraints on the problem are not very restrictive, the extrema are calculated for the tectonic parameters, as boundary conditions of Precambrian tectonics. The viability of Archean mobilist tectonics may then be examined. A method of testing hypotheses of Archean tectonics is hereby obtained. Models within the calculated bounds are acceptable, given internal consistency and agreement with the geologic record.

The Problem of Archean Geology

Direct information regarding relatively recent tectonic processes is easily obtained for the Phanerozoic eon, as the geologist can observe the tectonic morphology of the earth: the time-dependent relative positions of the plates as recorded in the magnetic signature of the oceanic and continental crust; and observation of the geologic effects of tectonics. The youth of the lithologic record results in a remarkable tectono-stratigraphic correlation, which is due, in part, to the abundance of paleontological stratigraphic indicators. This leads to a relatively unambiguous picture of the tectonic history of Phanerozoic rocks. On the other hand, the Precambrian lithologic record is generally discontinuous, poorly exposed and difficult to correlate. The apparent low preservation potential of Precambrian rocks (Allègre, 1982), resulting in a relative paucity of exposure, is a function of the dynamic balance between continental growth and continental recycling.

TABLE 1. Symbols

	Symbol	Units
Terrestrial heat loss	Q	W
Radiogenic heat production	Q_R	W
Radionuclide decay constant	λ	Ga^{-1}
Time before, present	t	Ga
Heat loss by secular cooling	Q_K	W
Thermal conductivity	k	$\text{W m}^{-1} \text{K}^{-1}$
Continental geothermal gradient	$\frac{dT_{cn}}{dz}$	K km^{-1}
Area	A	m^2
Tectonothermal age: particularly the mean age of oceanic lithosphere at subduction	t_*	Ma
Mean rate of opening	v	m a^{-1}
Total length of ridge	l	m
Temperature	T	K
Thermal diffusivity	α	$\text{m}^2 \text{s}^{-1}$
Age of oceanic crust	t_{cr}	Ma
Mantle heat loss	Q_m	W
Adiabatically extrapolated mantle temperature at the mantle-core boundary	T_s	K
Rate of change of the adiabatic mantle temperature	$\dot{T}_s$	K a^{-1}
Mantle solidus temperature at the mantle-core boundary	T_m	K
Heat capacity of the mantle	ϕ_m	J K^{-1}
Constant of proportionality	β	
Mantle geothermal gradient	$\frac{dT}{dz}$	K km^{-1}
Coefficient of volume expansion	α	K^{-1}
Specific heat	c	$\text{J kg}^{-1} \text{K}^{-1}$
Gravitational acceleration	g	m s^{-2}
Density	ρ	kg m^{-3}
Sub-lithospheric upper mantle	SLUM	
Sub-crustal upper mantle	SCUM	
Thickness of the oceanic lithosphere at subduction	d_{ll}	km
Heat flux	q	W m^{-2}

4

The inference of Archean mobilist tectonics is not unfounded. The basic characteristic of mobilist tectonics is plate motion. Since spreading ridges are geodegradable, the best evidence for plate motion is found in subduction-related phenomena. Brown (1979), for instance, supports the subduction-related nature of Archean calc-alkaline suites. The observed characteristic bimodalism of Archean volcanism invites comparison with Phanerozoic tectonostratigraphic regimes (Chadwick and Coe, 1984). Further evidence of plate motions is demonstrated by the existence in the Archean of regional (crustal (continental) extension (Nisbet, 1982). Finally, the existence of noncoincident intercratonic polar wander paths (dating from 3.4 Ga) (Kröner, 1982) documents continental drift, which implies large-scale horizontal rigid-plate motion.

The absence of blueschists, relative absence of andesites and apparent absence of rift zones, collisional sutures, ophiolites and other clear structural manifestations of mobilist tectonics tends to argue against Archean mobilist tectonics. However, the petrologic differences have been related to young subduction of Archean oceanic lithosphere, as indicated by Hynes (1982) and Abbott and Hoffman (1984). As an alternate approach Veizer and Jansen (1985) demonstrate that the poor preservation potential (maximum residence times < 300 Ma) of petrogenetic regimes of the oceanic tectonic domain (active margins, passive margins, immature orogenic belts, and oceanic crust) can account for such petrologic and structural differences without invoking any fundamental change in tectonic style.

As the geologist has no unambiguous direct information

TABLE 2. Basic Quantities

Symbol	Value	Reference
Q_{Earth}	$43. \cdot 10^{12} \text{ W}$	Davies (1980)
$Q_R(0)$	$29.2 \cdot 10^{12} \text{ W}$	McKenzie and Weiss (1975)
λ (Chondritic)	0.451 Ga^{-1}	McKenzie and Weiss (1975)
λ (K-depleted)	0.308 Ga^{-1}	McKenzie and Weiss (1975)
Q_K	$13.8 \cdot 10^{12} \text{ W}$	Vitarello and Pollack (1980)
A_{en}	$2.015 \cdot 10^{14} \text{ m}^2$	Sclater et al. (1980)
k_{en}	$2.5 \text{ W m}^{-1} \text{ K}^{-1}$	Reymer and Schubert (1984)
$\frac{dT}{dz}$	22.8 K km^{-1}	Bott (1982)
$v(0)$	$7.25 \cdot 10^{-2} \text{ m a}^{-1}$	Sclater et al. (1982)
$l(0)$	$5.6 \cdot 10^7 \text{ m}$	
$t^*(0)$	80.0 Ma	
T_{surface}	300 K	
α	$1.2 \cdot 10^{-6} \text{ m}^2 \text{ s}^{-1}$	Stacey (1977)
α	$2.68 \cdot 10^{-5} \text{ K}^{-1}$	
$A_{\text{SLUM}}(0)$	3.43 kg m^{-3}	Stacey (1977)
k_m (Chondritic)	$3.95 - 3.77 \text{ W m}^{-1} \text{ K}^{-1}$	
(K-depleted)	$3.83 - 3.89 \text{ W m}^{-1} \text{ K}^{-1}$	
c	$1.26 \cdot 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$	Bott (1982)
g	9.78 m s^{-2}	

concerning the Archean tectonic style, indirect evidence must be sought.

Heat Loss of the Earth

Tectonic activity is generally accepted to be the surface expression of thermal processes within the earth. Kanamori (1977) estimated the annual energy loss of the earth through work (straining of the lithosphere) to be $1.1 * 10^{17}$ J, equivalent to $3 * 10^{10}$ W. Forsyth and Uyeda (1975) estimate the same as $6.0 * 10^{17}$ J, equivalent to $16.4 * 10^{10}$ W. Recent surface heat flow measurements indicate that terrestrial heat, Q , is being lost at a rate of $4.3 * 10^{13}$ W (Davies, 1980), indicating that the earth is less than one percent efficient in dissipating its energy as work. Surface heat flux measures the conductive energy loss through the crust. The measured surface heat loss exceeds estimates of internal radiogenic heat production (see below), indicating that the earth is operating at a net heat loss, viz., it is cooling. Thus the total heat loss is the sum of the radiogenic heat production, Q_R , and the heat loss resulting from the decrease in the internal temperature of the earth, Q_K .

$$Q(t) = Q_R(t) + Q_K(t)$$

Radiogenic heat is produced by the decay of long-lived radioactive isotopes. The long-lived radioactive isotopes which significantly contribute to the earth's heat budget are ^{235}U , ^{238}U , ^{232}Th and ^{40}K . Radiogenic heat as a function of time may be expressed as

$$Q_R(t) = Q_R(0) \exp(-\lambda_R t) \quad (1)$$

where $Q_R(0)$ is the present value of radiogenic heat production, λ_R is the terrestrial weighted decay constant of the long-lived radioactive nuclides and t is time (Ga: $t = 0$ now, negative for times past). Although the radioactive signature of crustal rocks is fairly well defined, the inaccessibility of the earth's interior prevents the definition of the entire terrestrial radioactive signature. Earth compositional models generally fall into two well-accepted categories (Table 3): chondritic and K-depleted chondritic earth models. The chondritic earth model utilizes radionuclide abundances derived from carbonaceous chondrites (Mason, 1971; McKenzie and Weiss, 1975). Wasserburg et al. (1964) model K depletion from chondritic to crustal K/Th and K/U ratios, which is currently the preferred model (Bott, 1982). Using these two models we have

Chondritic	$Q_R(0) = 3.07 * 10^{13} \text{ W}$
K-depleted	$Q_R(0) = 2.74 * 10^{13} \text{ W}$

The decay of short-lived radioactive isotopes undoubtedly contributed vast quantities of heat to the proto-earth during the first 0.5 Ga following initial accretion (Lee et al., 1977; Singer, 1978; Bott, 1982). Combined with terrestrial accretionary energy and gravitational energy released by core separation, this produces a large amount of energy. Lack of quantification of this pulse of primordial heat generated by the proto-Earth precludes modelling of terrestrial heat flux prior to -4.0 Ga.

TABLE 3. Radiogenic Heat Production

Element	Elemental Abundance (ppm)		Elemental Heat Production (W / kg)	Total Heat Production (10^{-13} W / kg)	
	K	Ch		K	Ch
^{40}K	.0238	/.101	$27.9 * 10^{-6}$	6.64	/ 28.2
^{232}Th	.074	/.043	$26.6 * 10^{-6}$	19.7	/ 11.4
^{238}U	.020	/.012	$94. * 10^{-6}$	18.8	/ 11.3
^{235}U	.0012	/.00009	$570 * 10^{-6}$	0.684	/ 0.513

Terrestrial Heat Production (10^{-13} W/kg) 45.8 / 51.4

Mass of the Earth (Bott, 1982) = $5.974 * 10^{24}$ kg

Terrestrial Heat Production (10^{13} W) 2.74 / 3.07

Ch: Chondritic bulk earth composition.

Elemental abundances from Bott (1982), Mason (1971)

K: K-depleted bulk earth composition.

Elemental abundances from Wasserburg (1969), O'Nions et al. (1979), Richter (1985).

By numerical analysis of the thermal history of the earth, Stacey (1981) shows that the secular cooling rate, Q_K , has been constant throughout geologic time. Vitorello and Pollack (1980) have estimated $Q_K = 13.8 \times 10^{12} \text{ W}$, equivalent to 27 mW m^{-2} .

The combination of radiogenic heat production and secular cooling (Figure-1) indicates terrestrial heat flux three to five times higher in the early Archean. The problem of Precambrian tectonics is typically approached as follows: how could this excess heat have been dissipated? If we assume mobilist tectonics, what tectonic parameters may be altered within the coexisting frameworks of sea-floor spreading and subduction, considering the Precambrian lithologic record, which will account for dissipation of the heat produced? Or otherwise, was tectonics fundamentally different? Numerous tectonic models for the Archean have been proposed, based on both approaches (For example: Bickle, 1978; Baer, 1981; Hargraves, 1981; Sleep and Windley, 1982; Arndt, 1983; Nisbet and Fowler, 1983; Abbott and Hoffman, 1984; England and Bickle, 1984). The variation in these models bears testimony to the weak constraints on the geologic evidence.

Considering the outer shell of the earth, where the internal heat is dissipated, there are two crustal provinces through which this heat may pass, given mobilist tectonics: the spreading oceanic lithosphere (oc) and the continental lithosphere (cn). The total heat loss of the earth may be expressed as follows:

$$Q = Q_{oc} + Q_{cn} \quad (2)$$

The decreasing rate of heat loss implies that continental

Terrestrial Heat Loss

Figure 1.

The total earth heat loss, the sum of radiogenic heat production and secular cooling, assuming no time lag between radiogenic heat generation and dissipation. Rate of earth cooling, 13.8×10^{12} W taken from Vitorello and Pollack (1980).

Chondritic bulk earth radiogenic heat production is based on elemental concentration (^{40}K , ^{238}U , ^{235}U , ^{232}Th) as originally given by Mason (1971).

K-depleted bulk earth radiogenic heat production is based on elemental concentration originally given by Wasserburg (1964).

geothermal gradients were higher or the rate of oceanic heat dissipation was higher in the past (Burke et al., 1976a; Bickle, 1978; Burke and Kidd, 1978; Davies, 1979; Boak and Dymek, 1982). Increased oceanic heat dissipation could be attained by increased ridge length, increased spreading velocity, or a combination of both, hereafter referred to as the spreading activity. Quantification of ancient continental and oceanic thermal signatures will therefore permit quantification of the parameters of tectonic processes. The mantle temperature is required to define numerically the oceanic thermal regime, and therefore must be considered.

MANTLE TEMPERATURE

The immediate effect of the past increased heat production of the earth is an increased terrestrial thermal gradient. This is necessarily reflected in the mantle thermal gradient, which may then be expressed as the terrestrial cooling function, a balance between internal heat production, residual heat, heat input and heat output.

Stacey and Loper (1984) derive this function based on adiabatic decompression and rheologically controlled mantle-wide thermal convection. Their formula is as follows

$$Q_m(o) \exp \left(\frac{\beta T_m}{n+2} \left[\frac{1}{T_s(o)} - \frac{1}{T_s} \right] \right) = Q_{Rm} - \phi_m T_s$$

Stacey and Loper (1984: manipulation of their equation 11)

This expresses the thermal state of the mantle, viz., the rate of decay of mantle heat, $\phi_m T_s$, is equal to the heat produced

by radioactive decay (mantle), Q_{Rm} , minus the rate of secular cooling. The rheology of the mantle is described by the variable n , which is the stress exponent of the creep law. It is unity for Newtonian viscosity, and 3 for power-law creep (non-Newtonian rheology).

The derivation of the expression given above assumes that all the core's heat bypasses the mantle via hot rising plumes. Therefore the present heat loss of the mantle, Q_m , is equal to the terrestrial heat loss minus the sum of the lithospheric heat loss (7.0×10^{13} W) and the core heat flux ($1.6 / 1.7 \times 10^{13}$ W for $n = 1, 3$ respectively). The heat capacity of the mantle, C_m , is 4.65×10^{27} J K⁻¹. The initial rate of change of the mantle temperature, T_S , is calculated by substitution of present values of all other parameters. All values are as given by Stacey and Loper (1984). A compilation of the parameters is given in Table 4. Their calculation is here corrected to the adiabatic core-mantle boundary temperature, $T_S = 2800$ K, given by Ranalli and Fischer (1984: Figure 2).

The radiogenic heat production of the mantle, Q_{Rm} , is calculated using equation (1). The present radiogenic heat production of the mantle, $Q_{Rm}(0)$, acts as an adjustable parameter, fitting the temperature curve to the condition that the adiabatically-extrapolated temperature at the mantle-core boundary, T_S , must equal the mantle solidus temperature at the mantle-core boundary, T_M , during the first half billion years of the earth's existence (Stacey and Loper, 1984). This precludes the existence of a thermal boundary layer at the base of the mantle within the geologic history considered here (< 4.0 Ga).

TABLE 4. Parameters Used in Calculation of Mantle Temperature

ϕ_m	$4.65 * 10^{27} \text{ J K}^{-1}$	
β	35	
$T_s(0)$	2800 K	
T_m	3860 K	
	$n = 1$	$n = 3$
$Q_m(0)$	$33.4 * 10^{12} \text{ W}$	$33.3 * 10^{12} \text{ W}$
$Q_{Rm}(0)$	$29.95 * 10^{12} \text{ W}$	$26.40 * 10^{12} \text{ W}$
T_s	$-7.4 * 10^{-9} \text{ K a}^{-1}$	$-13.8 * 10^{-9} \text{ K a}^{-1}$

The best fit to $Q_R(0)$ is achieved for the chondritic earth model by the Newtonian mantle model, and for the K-depleted earth model by the non-Newtonian mantle model. These will be adopted in the calculations as the preferred mantle models for the respective earth models.

Numerical iteration of this function from $t=0$ to $t=-4.5$ Ga. for each rheological model of the mantle defines T_s as a function of time (Figure 2).

CONTINENTAL HEAT LOSS

The continental conductive heat loss is given by

$$Q_{cn} = k \frac{dT_{cn}}{dz} A_{cn} \quad (3)$$

This relation may be calculated as a function of time if the variation of continental area, A_{cn} , average geothermal gradient, $\frac{dT_{cn}}{dz}$, and mean thermal conductivity, k , are known as functions

of time. These three parameters are only poorly constrained by the geologic evidence, which as noted earlier has led to the large variation of Archean tectonic models. Using bounding rather than explicit values for these parameters removes this model-dependant uncertainty. The resultant extrema of the tectonic parameters (velocity, etc.) provide a strong basis for validation of Archean models: those that fit the calculated range are valid, although not necessarily correct.

Thermal conductivities of different rock types at crustal conditions fall within a limited range of values (Table 5). Therefore the thermal conductivity of the mean continental crust

Mantle-Core Boundary Temperature

Figure 2.

The temperature of the mantle-core boundary. Modelled to fit the present temperature distribution, and the condition that this temperature was equal to the melting temperature in the early history of the earth. Each bulk earth composition is modelled with both Newtonian and non-Newtonian mantle rheology.

Table 5: Thermal Conductivity Based on Rock Type

Rock Type	Thermal Conductivity ($\text{W m}^{-2} \text{K}^{-1}$)
Shale	1.2 - 3.0
Sandstone	1.5 - 4.2
Limestone	2.0 - 3.4
Dolomite	3.2 - 5.0
Marble	2.5 - 3.0
Pyroxenite	4.1 - 5.0
Basalt	1.3 - 2.9
Diorite	2.8 - 3.6
Gabbro	1.9 - 2.3
Diabase	1.7 - 2.5
Granodiorite	2.6 - 3.5
Granite	2.4 - 3.8
Anorthosite	1.7 - 2.1
Gneiss	2.1 - 4.2
Amphibolite	2.5 - 3.8
Mean Continental Crust	2.5
Mean Oceanic Crust	2.5
Peridotite	2.3 - 3.0
Dunite	3.7 - 4.6
Mean Sub-Oceanic Upper Mantle	3.3 (10-120 km) 3.7-6.9 (120-370 km)
Mantle Transition Zone	6.9 - 7.3
Lower Mantle	7.3 - 9.6

After Turcotte and Schubert (1982: Appendix 2E) and Bott (1982: Table 7.3).

composition may be deduced to have remained approximately constant ($2.5 \text{ W m}^{-1} \text{ K}^{-1}$: Stacey, 1977) through time.

Continental Area

Numerous continental accretion models have been presented. (Figure 3A). However, each model considers different aspects of the problem. The combined constraints of the models may be placed on each individual accretion model. Only three of the proposed models meet all of these constraints, of which the maximum and minimum may be considered to be extreme models of continental accretion. The following is a discussion of the constraints which must be met by continental accretion models.

Continental accretion is known to be accomplished through at least four mechanisms (Dewey and Windley, 1981):

1. Arc magmatism along destructive plate margins,
2. Intraplate volcanism,
3. Overthrusting of ophiolitic suites, melange packages and epeirogenic prisms, and
4. Underplating.

The postulated increased spreading activity in the past would undoubtedly have led to increased melange accretion, overthrusting and arc magmatism. This implies that the rate of generation of continental material has been decreasing from the time of continental nucleation.

An absolute minimum of cumulative continental area at any given age of the earth is provided by measuring the present-day areas of continental crust of each age interval (Hurley and Rand, 1969). However, this grossly underestimates the minimum, due to

Continental Accretion Models

Figure 3A.

A selection of models of continental accretion.
Modified after Reymer and Schubert (1984: figure 6).

A + J	Allègre and Jaupart (1985)
Am	Armstrong (1981)
B	Brown (1979)
F	Fyfe (1978: curve f)
H + R	Hurley and Rand (1969)
M + T	McLennan and Taylor (1985)
O'N	O'Nions et al. (1979)
R + S	Reymer and Schubert (1984)

the decreasing preservation potential of continental crust with increasing age (Allègre, 1982), a result of the competing processes of continental growth and continental recycling.

Models which assume continental nucleation at 3.8 Ga assume that all crust formed prior to this time was reincorporated into the mantle. The presence of sediments containing zircons with cooling ages of 4.1-4.2 Ga (Froude et al., 1983) strongly suggests that protocontinents nucleated prior to 3.8 Ga. Furthermore, large portions of the mantle were geochemically and isotopically depleted by 4.0 Ga (Arndt et al., in press). The age dependence of preservation potential noted above could also be invoked to explain the apparent absence of crust older than 3.8 Ga. Continental accretion curves with nucleation of continents post-dating 3.8 Ga are therefore inferred to be inaccurate. The occurrence of Archean cratons beneath Phanerozoic and Proterozoic orogenic zones, implies the existence of large volumes of Archean crust, from which Archean continental accretion is inferred to be rapid (Villeneuve, 1984).

Present-day through Phanerozoic continental accretion rates may be estimated from island arc accretion, which is likely to be the main mechanism of continental accretion (Reymer and Schubert, 1984). Continental freeboard calculations may also be used to indicate progressive continental accretion (Schubert and Reymer 1985). Curves which give anomalously high or low to negative Phanerozoic addition rates relative to calculated accretion rates may be considered inaccurate. Allègre and Jaupart (1985) and Veizer and Jansen (1985) demonstrate that continental growth is expected to achieve a steady state through the dynamic balance

between addition and subtraction of continental material. Allègre and Jaupart (1985) and Veizer and Jansen (1979) further show that the premise of a present-day, near steady state condition produces geologically-acceptable estimates of accretion and recycling rates, and mean continental age.

Three models of continental accretion seem to meet the above constraints (Figure 3B): those of Brown (1979: Model I), O'Nions et al. (1979: obtained by integration of rate of accretion diagram), and the models of Allègre and Jaupart (1985). Allègre and Jaupart (1985) numerically model the time-dependent tectonic extraction-destruction continental growth curve. O'Nions et al. (1979) numerically model the time-dependent geochemical extraction-destruction continental growth curve. Brown (1979) models continental growth rates directly proportional to the exponentially decaying terrestrial heat flux.

The Brown and O'Nions models may be adopted as the maximum and minimum models, respectively, of volumetric continental growth. Present-day accretion rates, as computed by Reymer and Schubert (1984), show good agreement with the O'Nions model, but are overestimated by the Brown model. Both demonstrate steadily decreasing accretion rates from at least -4.0 Ga from a non-catastrophic nucleation event. The Brown model incorporates high nucleation accretion rates, while the O'Nions model incorporates low rates. The O'Nions model represents a valid minimum since it indicates greater than fifty percent post-Archean accretion.

Based on the above discussion, the continental accretion models of Brown and O'Nions will be adopted in this paper as the

Continental Accretion Models

Figure 3B.

Models of continental accretion having non-catastrophic continental nucleation prior to 3.8 Ga, and acceptable accretion histories.

A + J	Allègre and Jaupart (1985)
B	Brown (1979)
O'N	O'Nions <u>et al.</u> (1979)

absolute extremes of continental accretion models.

Strictly speaking, these curves give continental volume, not continental area. Obviously, if the continents were thinner in the past, an equal volume of crust would represent a larger area, relative to today. However, pressure estimates based on metamorphic assemblages indicate that the crustal portion of the protocontinents were as thick in the Archean and Proterozoic as they are now (Davies, 1979; Grambling, 1981; England and Bickle, 1984; Nisbet, 1984). Boak and Dymek (1982) infer that this has been the case from at least -3.8 Ga. Windley and Davis (1976) have provided an independent estimate of Archean crustal and lithospheric thickness, of comparable magnitude, from volcanic centre spacings (restoration by simple shear) using the Mohr and Wood (1976) model of lithospheric thickness vs. volcanic spacing.

The local effects of crustal attenuation and orogenic shortening (thickening) only prevail at time scales less than a Wilson cycle. Within a time scale longer than a Wilson cycle, by the evidence above continental thickness has remained virtually constant. Based on the mean time between continental collisions (Allègre and Jaupart, 1985), the mean length of a Wilson cycle is taken to be 500 Ma, which is geologically reasonable. Although the time-density of continental collisions is less in the past (Allègre and Jaupart, 1985) the Wilson cycle time scale is reduced for the oceanic domains, which are considered here. Thus 500 Ma is a reasonable time scale throughout geologic time.

Since thickness has remained constant, percent volume is equivalent to percent area. The volumetric accretion curves are therefore equivalent to areal accretion curves. This breaks down

In the earliest history of earth during the initial stages of cratonization before appropriate continental thicknesses were achieved. Based on Boak and Dymek's (1982) data, this equivalence of area and volume may be assumed from at least -3.8 Ga.

Geothermal Gradients

Since the terrestrial heat loss was higher in the past, it is reasonable to assume that the present value of the mean continental geothermal flux represents a minimum value as a function of the age of the earth. Estimates of this mean are given by Clark and Ringwood (1964: shield values), Lee (1970: equal area grid average), Sclater et al. (1980: areal average) and Turcotte and Schubert (1982: unweighted average). The areal averaging method utilized by Sclater et al. probably gives the most accurate figure, 57 mW m^{-2} , equivalent to a geothermal gradient in the upper lithosphere of 22.8 K km^{-1} . Present continental heat loss of $11.5 * 10^{12} \text{ W}$ is calculated based on continental thermal conductivity $k = 2.5 \text{ W m}^{-1} \text{ K}^{-1}$ and present continental area $A_{\text{cn}} = 2.105 * 10^{14} \text{ m}^2$.

An independent estimate of past geothermal gradients may be obtained from geothermobarometric analysis of prograde metamorphic mineral assemblages. Care must be exercised in applying this method to the lithologic record due to possible tectonic or thermal overprint (Burke and Kidd, 1978; Lambert, 1983), where underestimation of pressure of formation implies a lesser depth of burial, thereby overestimating the associated geothermal gradient. The geothermobarometric analyses cited

previously as evidence of constant continental thickness further indicate constancy of geothermal gradients in time. Additional evidence is derived from garnetiferous and silicious inclusions in xenolithic diamonds (Boyd et al., 1985). These indicate continental lithosphere at 150 - 200 km and 1200 - 1500 K circa 3.0 Ga which are in good agreement with present geothermal gradients.

Grambling (1981) has compiled P-T estimates for published descriptions of nearly 100 Precambrian rocks. His compilation produces the following mean geothermal gradients:

<u>Time</u>	<u>Geothermal Gradient</u>	<u>Standard Deviation</u>
1.5 - 1.0 Ga	35 K / km	9.0 K / km
2.5 - 1.5 Ga	47 K / km	13.7 K / km
>2.5 Ga	54 K / km	16.5 K / km

However, these data have no areal weighting factor. The increased spreading activity in the past tends to imply a higher time density of orogenic events. This is not necessarily true, but for the purpose of assigning a maximum geothermal gradient it may be assumed. The higher ratio of length of continental margin to continental surface area (implied by lesser continental area) lends credence to this hypothesis. The decrease of heat flux with increasing age of the last tectonothermal event (Vitorello and Pollack, 1980) also implies higher geothermal gradients for the tectonically younger Archean crust. Hence, high P-T values should be more common in the past. It is assumed that the continental mean geothermal gradient, if decreasing at all, should do so exponentially (by analogy with the heat loss). By polynomial interpolation the following exponential representation

of Grambling's data is derived.

$$\frac{dT(t)}{dz} = 23.84 / t^{0.29}$$

This curve fits Grambling's data within a 95 % confidence level. Lambert (1983) cites maximum Proterozoic surface heat flux values as $80-120 \text{ mW m}^{-3}$ ($32-48 \text{ K km}^{-1}$), assuming a tectonothermal overprint. Since the geothermal gradients computed above overestimate, albeit slightly, Lambert's values, they will be adopted as absolute maxima.

In order to minimize Q_{cn} as a function of time, the minimum values of A_{cn} and $\frac{dT_{cn}}{dz}$ are substituted into equation (3).

Maximization of Q_{cn} is similarly accomplished by using maximum values of A_{cn} and $\frac{dT_{cn}}{dz}$ (Figure 4).

OCEANIC HEAT FLUX

Using the heat budget equation (2), we may now solve for the total heat loss which must be dissipated through the oceanic lithosphere, Q_{oc} , as a function of time. As noted earlier, the purpose is to compute the extrema, rather than model-dependent explicit values of the tectonic parameters. Maximization of oceanic heat dissipation is accomplished by subtracting the minimum Q_{cn} from the maximum Q . Similarly, subtracting the maximum Q_{cn} from the minimum Q minimizes Q_{oc} . The maximum and minimum will necessarily coincide at -4.5 Ga as there was no continental contribution to heat dissipation. The oceanic heat flux is dissipated through the mechanism of plate creation at oceanic spreading ridges, and subsequent cooling prior to

Continental Heat Loss

Figure 4.
Models of continental heat loss, computed using

$$Q_{cn} = k \frac{dT}{dz}_{cn} A_{cn}$$

Maximum model adopts the Brown model of continental accretion, and maximum estimated geothermal gradients. Minimum model adopts the O'Nions model of continental accretion, and minimum estimated geothermal gradients (present-day value).

reincorporation of the oceanic lithosphere into the mantle. The total oceanic heat dissipation depends on the total oceanic area as defined by

$$A = A_{cn} + A_{oc}$$

Paleoradius calculation based on paleomagnetic data indicates that the earth has maintained approximately constant volume (McElhinny et al., 1978). Since the earth may be considered to have maintained constant volume, it therefore has maintained constant surface area. Two results follow. First, the oceanic area is defined within the error of limits whenever the continental area is known. Second, within the framework of plate tectonics, area loss by subduction must be balanced by area gain through lateral continental accretion and oceanic plate creation at spreading centres. Schubert and Reymer (1985) estimate the present rate of continental accretion as $1 \text{ km}^3 \text{ a}^{-1}$. Turcotte and Schubert (1982) estimate the present rate of creation of oceanic crust as $25 \text{ km}^3 \text{ a}^{-1}$. Thus the tectonic creation and destruction of surface area are largely oceanic functions. Since terrestrial area is conserved, oceanic area must be conserved, and creation must balance destruction. Therefore the area of the oceans may be approximated by

$$A_{oc} = \text{width} * \text{length}$$

$$= v t_* l$$

$$= v l t_*$$

(4)

where t_* is the mean age at subduction, l is the total ridge length and v is the mean opening rate. The function is a trivial statement for symmetric oceans (ridge in the middle, bounded by

subduction). Nevertheless the present asymmetric distribution of subduction zones with respect to oceanic spreading centers is necessarily controlled by the terrestrial conservation of area. Thus, younger subducting lithosphere (relative to the mean) must be areally balanced somewhere in the earth's total oceanic area by non-subducting and older subducting oceanic lithosphere or by higher spreading activity. Ridge-trench geometry is therefore intrinsically controlled by the area function given by equation (4). This is the same function describing oceanic area as used by Bickle (1978) and also may be inferred from Allègre and Jaupart (1985).

The present value for t_* , 80 Ma, is attained by length-averaging the actual age of the subducting lithosphere for all currently active subduction zones. Comparison of this value with the slab buoyancy curve of Sacks (1983) indicates that on average subduction is presently negative-buoyant. Wortel (1980) compiles the ages of the subducting lithosphere for the terrestrial subduction zones (see Appendix 4).

The present value for the mean velocity of spreading, $v = 7.25 \times 10^{-2} \text{ m a}^{-1}$ is estimated based on the data given by Minster and Jordan (1978), using only the angular velocity data corresponding to spreading zones. This mean is weighted by length, as measured on the tectonic globe. Comparison of the total measured length of ridge with that of Reymer and Schubert (1984) shows agreement within 6 %. The ridge data are compiled in Appendix 5.

The present value for the total length of ridge, $L = 5.6 \times 10^7 \text{ m}$ is taken from Reymer and Schubert (1984,

Table 1).

The oceanic area calculated from equation (4) shows good agreement (11 %) with present oceanic surface area.

Assuming that oceanic lithosphere acts as a thickening thermal boundary layer, the oceanic surface heat loss, Q_{oc} , is related to the mean age of the crust at subduction, t_* by

$$Q_{oc} = \frac{2k}{(\pi\kappa)^{1/2}} \Delta T_* \frac{A_{oc}}{t_*^{1/2}} \quad (5a)$$

(derivation in Appendix 1A)

which may be expressed as

$$t_*^{1/2} = \frac{2k}{(\pi\kappa)^{1/2}} \Delta T_* \frac{A_{oc}}{Q_{oc}} \quad (5b)$$

where t_* is the mean age of oceanic crust at subduction. The temperature difference between the sublithospheric upper mantle (SLUM) and the top of the crust is denoted by ΔT_* , and is assumed constant over the time span of a Wilson cycle.

The composite, $k\kappa^{-1/2}$, is left as an adjustable parameter to fit the calculated function to the present mean age of subduction, t_* . Assigning the coefficient of thermal diffusivity a reasonable value for igneous rocks, $1.2 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ (Stacey, 1977), allows calculation of k . Thus $k = 3.95\text{--}4.01 \text{ W m}^{-1} \text{ K}^{-1}$ for the mantle (chondritic models), which fall within the acceptable range of values (Table 5). The K-depleted earth model requires a thermal diffusivity of $1.4 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ to achieve a reasonable coefficient of thermal expansion, $k = 3.83\text{--}3.89 \text{ W m}^{-1} \text{ K}^{-1}$.

In order to define t_* numerically, the cooling function of the mantle must be known. Assuming that the mantle-core boundary

has maintained a constant depth of 2886 km throughout time, the temperature of SLUM is derived by extrapolation, according to the average mantle adiabat, of the mantle geotherm from the core-mantle boundary to the top of the mantle, ignoring mantle structure (see Figure 5), using

$$T_{\text{SLUM}}(t) = T_S / \exp \left(\frac{\alpha g}{c} * 2690 \right)$$

where T_S is the mantle-core boundary temperature. The derivation of this function is given in Appendix B. The temperature of the SLUM through time is calculated at an arbitrary constant depth of 200 km, which presently represents the top of the mantle adiabat (Stacey, 1977, Fig. 7.4). The average adiabatic mantle geotherm, $\frac{dT}{dz}$ is given by

$$\frac{dT}{dz} = \frac{\alpha g}{c} T$$

(Turcotte and Schubert, 1982)

The volume coefficient of thermal expansion, α , is used as an adjustable parameter to fit the present geotherm at 200 km. The temperature at this depth is estimated as 1600 K by Ranalli and Fischer (1983). Using an average mantle specific heat, c , equal to $1.26 * 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$ (Bott, 1982), gives an average mantle coefficient of thermal expansion of $2.24 * 10^{-5} \text{ K}^{-1}$, which falls within the acceptable range of values (Bott, 1982: Table 7.3).

Spreading Rate and Length of Ridge

Since continental geothermal gradients have remained approximately constant in time, it follows that the rate of

Mantle Adiabats

Figure 5.

The mantle adiabat, relating the present temperatures of the sub-lithospheric upper mantle, $T_{SLUM}(0)$, and the mantle-core boundary, $T_s(0)$, to the past values, $T_{SLUM}(t)$ and $T_s(0)$. The mantle adiabat given by

$$\frac{dT}{dz'} = \frac{\alpha g}{c} T$$

oceanic heat loss was higher in the past. As stated previously, this increased heat loss implies an increased spreading activity. The spreading activity is given by (equation 4)

$$v \ell = A_{oc} / t_*$$

No geologic evidence exists to define reasonable Precambrian values of either velocity of spreading or total ridge length. Plate velocities lower than present values are mathematically feasible given a proportionately increased ridge length, and vice versa. However, increased spreading activity suggests increased v and ℓ . Therefore current values of rate of spreading and total ridge length are proposed as absolute minima throughout geologic time. Thus fixing one of the parameters at present value in equation (4) permits calculation of the maximum value of the other.

$$\ell_{MAX} = \frac{A_{oc}}{t_* v(o)}$$

$$v_{MAX} = \frac{A_{oc}}{t_* \ell(o)}$$

Mantle Density, and Thickness of Oceanic Lithosphere

Two additional tectonic parameters may be calculated based on the upper mantle temperature: the density of the upper mantle, and the thickness of the oceanic lithosphere.

The thickness of oceanic lithosphere, d_{li} , as a function of age, is given by

$$\frac{T_{scum} - T}{T_{scum} - T} = \text{erf}\left(\frac{d_{li}}{2(\pi t_*)^{1/2}}\right) \quad (6)$$

(Turcotte and Schubert, 1982)

from which the thickness of the lithosphere at subduction (d_{li} at t_*) may be calculated. The surface temperature, here denoted by T , is 300 K. The mean temperature of the sub-crustal upper mantle (SCUM), T_{SCUM} , is left as an adjustable parameter to fit d_{li} at t_* to thicknesses determined from Rayleigh wave dispersion data. Leeds et al. (1974) cite the oceanic thickness at 80 Ma as 75 km (their Figure 1). This is approximated by the Turcotte and Schubert (1982) model as 80 km. The above correction gives a temperature of 1200 K, which is a reasonable value for T_{SCUM} .

In light of negative-buoyancy subduction, the density of the mantle is an important tectonic parameter. Negative-buoyancy subduction will occur only if the subducting oceanic slab is denser than the underlying upper mantle. The density of SLUM therefore provides a lower limit of lithospheric density subject to spontaneous negative-buoyancy subduction.

If the density of the lithosphere in the subducting slab is less than the density of the upper mantle, then negative-buoyancy is not the main driving mechanism of subduction.

The variation of upper mantle density, ρ_{SLUM} , as a function of the increasing temperature (with increasing age), independent of pressure variation (arbitrary constant depth) may be calculated from

$$\rho = \rho(0) \exp(\alpha \Delta T)$$

where $\Delta T = T(0) - T(t)$. the derivation of this function is given in Appendix IC. The average upper mantle (SLUM) density to 400 km is $3.49 \times 10^3 \text{ kg m}^{-3}$ (Dziewonski et al., 1975). The coefficient of thermal expansion is assigned the same value

fitted for the calculation of T_{SLUM} .

The crustal thickness necessary to preclude negative-buoyancy may be calculated using

$$(\rho_{SLUM} - \rho_{cr}) d_{cr} = 0.5 \rho_{SLUM} \alpha T_{SLUM} d_{li}$$

McKenzie (1969).

It should be noted that for a given lithospheric thickness, increasing the crustal thickness tends to make the slab positive-buoyant, viz., it floats. Similarly, decreasing the asthenospheric density tends to increase the crustal thickness necessary to preclude slab negative-buoyancy. Molnar and Atwater (1978) estimate the accuracy of this method to be 30 %. The coefficient of thermal expansion is used as an adjustable parameter to fit the present-day oceanic crustal thickness, 10 km (Bott, 1982). This gives $\alpha = 1.4 \times 10^{-5} \text{ K}^{-1}$, which is a reasonable value for igneous rocks (Turcotte and Schubert, 1982). Densities of various rock types are given in Table 6.

Crust Composition

The present oceanic lithosphere is composed of 1 km of basalt, underlain by 9 km of predominantly gabbro, underlain by peridotitic SCUM (Bott, 1982), which is in turn underlain by SLUM. The crustal portion of the lithosphere may be approximated to a gabbroic composition.

The question of composition of the Archean oceanic crust has been raised in view of the apparent higher proportion of komatiites in the Archean (Nisbet, 1984). Where SLUM temperature is elevated sufficiently to intersect the komatiite liquidus, komatiitic extrusion may be assumed. Green *et al.* (1975) give

TABLE 6. Mean Densities of Various Rock Types

Rock Type	Density (kg m^{-3})
Basalt	2.74
Gabbro	2.95
Komatiite	3.22
Peridotite	3.22
Dunite	3.37

From Carmichael (1984). Densities at surface temperature and pressure.

the liquidus temperature of komatiitic melts at 200 km depth as 1970 K. This is equivalent to the minimum SLUM temperature required for generation of komatiitic melts. Oceanic komatiite extrusion would slightly increase the density of the crustal component of oceanic lithosphere. In this instance, the crustal portion of the lithosphere still may be approximated to a gabbroic composition. Densities for various rock types are listed in Table 6.

SUMMARY OF RESULTS

The results of the calculations are plotted in Figures 6-13. Tables of the results are provided in Appendix B.

Sub-Lithospheric Upper Mantle Temperature

Sub-lithospheric upper mantle temperatures were 200-300 degrees higher in the early Archean (Figure 6). The variation of SLUM temperature is very dependent on the rheological mantle model chosen.

By extrapolation of data measured by Green and Ringwood (1967) the solidus temperature of undepleted peridotite mantle at 200 km depth (65 kbar) is approximately 1850 K. The temperature of SLUM, Figure 6, achieves this value in the modelled non-Newtonian mantle, indicating a partially molten layer at this depth, until at least -3.5 Ga. No partially molten layer is implied for the Newtonian mantle postdating -4.0 Ga.

Oceanic Heat Loss

The resultant oceanic heat loss was three to six times higher than present values in the Archean (Figure 7). Within the framework of mobilist tectonics, this excess oceanic heat was necessarily dissipated by subduction of younger crust through increased spreading activity.

Mean Age of Oceanic Lithosphere at Subduction

The distinction between mantle models is less significant than the distinction between bulk earth models for t_* (Figure 8). Referring to equation (5b), the mean age of subduction is influenced by the earth model via the parameter Q_{oc} , and by the

Sub-Lithospheric Upper Mantle Temperature

Figure 6.

The temperature of the sub-lithospheric upper mantle. Obtained by adiabatic extrapolation from the mantle-core boundary. Each bulk earth composition is modelled with Newtonian and non-Newtonian mantle rheology.

Oceanic Heat Loss

Figure 7.

The extremes of oceanic heat loss for each bulk earth, calculated by subtracting each continental heat loss model from the terrestrial heat loss model.

Figure 8A.

The mean age of the oceanic lithosphere at subduction (t_*) which dissipates the oceanic heat flux. The figure contrasts variation of t_* as a function of continental thermal models with variation of t_* as a function of mantle rheology models. Mantle rheology is noted to have less effect on t_* than the continental thermal model.

A: Chondritic bulk earth composition.

Figure 8B.

The mean age of the oceanic lithosphere at subduction (t_*) which dissipates the oceanic heat flux. The figure contrasts variation of t_* as a function of continental thermal models with variation of t_* as a function of mantle rheology models. Mantle rheology is noted to have less effect on t_* than the continental thermal model.

B: K-depleted bulk earth composition.

rheology of the mantle via ΔT . At -4.0 Ga the respective model ratios are as follows:

Q_{oc}	Chondritic : K-Depleted	1.7
T_s	non-Newtonian : Newtonian	1.1

This clarifies the greater dependence of t_* on the bulk earth model than on the mantle rheology. It follows that no statement concerning mantle rheology may be made based solely on Archean tectonic style.

With reference to Figure 8C, it is further observed that the mean age of subduction is less dependent on the variability of continental heat loss than on the bulk earth model variation. Hence the extreme values of mean age of subduction are dependent only on the bulk earth models, being relatively independent of the thermal continental models and mantle rheology models. It is concluded that the thermal expression of the Archean tectonic regime(3) is dependent only on the bulk earth model, and is independent of the continental thermal regime and the rheology of the mantle.

The mean age of the oceanic crust at subduction displays an exponential decrease from 80 Ma today to 7-27 Ma at -4.0 Ga (Figure 8). The minimum for t_* corresponds to a chondritic earth, and the maximum to a K-depleted earth. Minimization of t_* will maximize the spreading activity.

Spreading Rate and Ridge Length

The range of velocities calculated for -4.0 Ga varies from present values to five times present values for K-depleted earth

Mean Age at Subduction

Figure 8C.

The mean age of the oceanic lithosphere at subduction (t_*) which dissipates the oceanic heat flux. Variation of t_* as a function of the bulk earth models and the continental thermal models. Continental thermal model is noted to produce less variation in t_* than the bulk earth model. The preferred model is chosen to maximize the spreading activity (minimum t_*).

Figure 9.

Oceanic spreading activity, defined as the product of oceanic spreading rates and total ridge length. Normalized to present to give a relative measure of the rate of convective overturn.

models, and from present values to sixteen times present values for chondritic earth models (Figure 10). Plate velocity does have a measurable geologic indicator. Polar wander velocities derived from paleomagnetic signatures may be indicative of continental plate velocities. Precambrian paleomagnetic data (Burke et al., 1976b; McElhinny and McWilliams, 1977; Chase, 1978; Dunlop, 1981; Hargraves and Oostott, 1982; Koster, 1982) show that the rate of continental plate motion was not in excess of present rates of motion.

Assuming that continental and oceanic plate velocities were comparable we may assume present rates of motion based on the paleomagnetic results. The excess Precambrian heat flux would then have been dissipated by younger subduction provided by greater ridge length. To meet the demands of heat dissipation, the total ridge length would have been five to sixteen times greater than today in the early Archean for K-depleted and chondritic earth models, respectively (Figure 11).

If we assume square plates, doubling the ridge length effectively quarters the plate area, and halves plate width. Thus, for the K-depleted earth, average plate width would be one fifth the present plate width. For the chondritic earth, plate size would be reduced to one-sixteenth the present width. This produces many, small (300 km wide) plates. This is broadly consistent with Smith's (1981) polygonal tectonics. For the chondritic earth, it is deduced to be more likely that mobilist tectonics existed with more rapid overturn. Oceanic plate velocity twice the present value would require 500 km wide plates; four times the present velocity would require 1000 km

Oceanic Spreading Rate

Figure 10.

Oceanic spreading rate. Present-day rates are proposed as the minimum value. Maximum rates for each model are calculated from the spreading activity using the present length of ridge, which is proposed as a reasonable minimum.

Figure 11.

Length of ridge. Present-day rates are proposed as the minimum value. Maximum rates for each model are calculated from the spreading activity using the present rate of oceanic spreading, which is proposed as a reasonable minimum.

wide plates.

Piper (1976), McElhinny and McWilliams (1977), Dunlop (1981) and Kröner (1981) cite the intra-cratonic coherence of paleomagnetic data, and the smoothness of apparent polar wander curves as evidence for large continental masses in the Archean. Nevertheless, oceanic plates could have been smaller, possibly giving rise to polygonal tectonics (Smith, 1981) or intra-oceanic linear tectonics.

Considering the present distribution of plate velocities (Table 7), an obvious division exists between continental and oceanic plates. The continental blocks are typified by slower motion, relative to the oceanic velocities. This is presumably due to the greater viscous drag beneath continents. The mean ratio of oceanic to continental velocities is 2.9. Archean continental lithospheric thicknesses were comparable to present values and oceanic lithospheric thicknesses were less, implying an increased ratio of oceanic to continental velocity in the Archean. This implies that continental plates acted as relatively immobile rafts surrounded by rapidly moving oceanic plates. This would result in slowly migrating, rapid subduction. Mean oceanic plate velocities four to fifteen times present values could therefore have existed in the early Archean, with ridge length approximately equal to the present total ridge length (Figure 11). Accepting Archean paleomagnetic continental plate velocity determinations as valid, produces total ridge lengths of 91,000 - 253,000 km at -4.0 Ga, corresponding to the K-depleted and chondritic earth models, respectively.

Table 7: Absolute Rotation Vectors of the Plates

Plate	Rotation Vector (Degree/Ma)
African	0.139
Antarctic	0.054
Arabian	0.388
Caribbean	0.129
Cocos	1.422
Eurasian	0.038
Indian	0.716
Nazca	0.585
North America	0.247
Pacific	0.967
South America	0.285

Taken from Minster and Jordan (1978) Table 7. Error correlates inversely with absolute velocity.

Thickness of the Oceanic Lithosphere

From Figure 12 it may be seen that in the Archean the thickness of the oceanic lithosphere at subduction did not exceed 50 km, and in the early Archean was 20-35 km. Turcotte and Schubert (1982) note the underestimation of the present-day thickness of the oceanic lithosphere by theoretical models (equation 6), citing by way of explanation the neglect of convective cooling induced by hydrothermal circulation of seawater. The calculated lithospheric thicknesses and their respective lithospheric densities are therefore minima.

Thickness of the Oceanic Crust

For the chondritic earth model, basaltic crust of greater than 3-4 km at 2.5 Ga and less than 2 km throughout the Archean would preclude negative-buoyancy subduction (Figure 13). It may be concluded that basaltic oceanic crust of a bulk chondritic earth supports a change in tectonic style of the Archean-Proterozoic boundary, from buoyant (Archean) to negative-buoyant subduction. For komatiitic oceanic crust this thickness becomes 5-7 km. Thus, for bulk chondritic earth, a komatiitic oceanic crust is more likely if negative-buoyancy subduction occurred. A basaltic crust would tend to imply buoyant subduction in the Archean if mobilist tectonics acted. The model assumes constant upper mantle composition.

For the K-depleted earth model, basaltic crust less than 5-6 km thickness would be capable of normal subduction. Komatiitic crust less than 10 km thick would be negatively-buoyant. It seems likely that for the K-depleted earth model,

Thickness of Oceanic Lithosphere at Subduction

Figure 12.

The thickness of oceanic lithosphere at subduction, modelled assuming thermal (cooling) thickening proportionate to the distance from the ridge.

Figure 13A.

The thickness of the oceanic crust which is necessary to preclude negative-buoyancy subduction (voluntary sinking of the subducting slab). Crust thicker than the values indicated will float beneath the over-riding plate. Modelled for both komatiitic and basaltic compositions.

Chondritic bulk earth composition model.

Figure 13B.

The thickness of the oceanic crust which is necessary to preclude negative-buoyancy subduction (voluntary sinking of the subducting slab). Crust thicker than the values indicated will float beneath the over-riding plate. Modelled for both komatiitic and basaltic compositions.

K-depleted bulk earth composition model.

mobilitist tectonics driven by negative-buoyancy characterized the Archean.

DISCUSSION OF EXISTING MODELS

Any internally consistent models with tectonic parameters falling within the ranges expressed here are viable tectonic models. Failure of any particular model to meet these criteria proves the model invalid.

Sub-Lithospheric Upper Mantle Temperature

Davies (1979) modelled the late Archean SLUM temperature as being 150 K higher than today, based on Newtonian convective heat transport, assuming mantle heat generation three times present values. This assumption fits well in the chondritic earth model prior to ~3.0 Ga, and also in the K-depleted earth model prior to ~4.0 Ga. Within these time frames the Davies mantle shows good agreement with the analysis presented here. Turcotte (1980) performed a similar calculation, for a Newtonian mantle model, with similar results. In 1984, Nisbet suggested SLUM temperatures 350 K higher than present values, based on the adiabatic upwelling of asthenospheric material from 75 - 80 km depth to produce komatiitic extrusives. This estimate exceeds the calculated range of SLUM temperatures corresponding to the late Archean, but it does correspond to the early Archean non-Newtonian mantle temperatures.

Composition of Oceanic Crust

Two acceptable models for the Archean oceanic crustal composition have been proposed. Adiabatic upwelling from deep

sources could produce komatiites (Nisbet and Fowler, 1983; Nisbet, 1984; Arndt, 1983) and from shallow sources would produce gabbros (Glikson, 1972; Sleep and Windley, 1982).

Upwelling of mantle material from depth would produce melts by adiabatic decompression, either for localized extrusion or more extensive extrusion to create oceanic crust. Relatively shallow plume source depths would produce low temperature, basaltic melts, whereas upwelling plumes from greater depth would contain sufficient heat to generate komatiitic melts. Mantle temperatures for the modelled non-Newtonian mantle are higher than those for the Newtonian mantle, resulting in shallower depths of origin for melts of either composition than required for the Newtonian mantle. However, the non-Newtonian mantle makes a better case for the generation of komatiitic melts throughout the Archean. This could be invoked as evidence supporting a non-Newtonian mantle.

Nisbet and Fowler (1983) demonstrate that a komatiitic oceanic crust will be negatively-buoyant at the time of mean age of subduction if it overlies a dunitic mantle. Arndt (1983) demonstrates that a partly komatiitic oceanic crust would induce negative-buoyancy. Adopting the maximum age of subduction here, it can be reasonably shown that his oceanic lithosphere model having a basaltic crust would be dense enough to be negatively-buoyant. His SLUM density is lower than those modelled by Dziewonski et al. (1975), which have been adopted here.

Mean Age of Oceanic Lithosphere at Suduction

Estimates of the mean age of subduction given in the model

of Bickle (1978) approximate the chondritic model value. However, he assumes constant continental (and therefore, oceanic) area, which divided by the rate of plate production inferred from heat loss, gives the mean age at subduction. Correcting his model to the increased oceanic area of the past would increase his estimate of t_* to a better approximation of the K-depleted earth model.

Abbott and Hoffman (1984) contend that subduction of buoyant (younger) oceanic lithosphere would induce shallower partial melting, giving rise to tonalitic and trondhjemitic compositions. Deeper melting would produce andesitic melts. They also state that deep melting in the asthenospheric wedge would generate komatiitic melts, presumably with extrusion on the overlying plate. However, this does not explain the observed compositional evolution of the melt series associated with Archean komatiites as documented by Arndt (1982).

Thickness of Oceanic Lithosphere

Oceanic lithospheric thickness has received its due attention in the literature; estimates for which range from 50 km (Nisbet and Fowler, 1983) to 20 km (Sleep and Windley, 1982), which covers the full late Archean range delineated herein. Nisbet (1984) and Nisbet and Fowler (1983) advance the hypothesis of 20 km of oceanic crust, based on isostatic equilibrium with continental extrusives. Negative-buoyancy is then enhanced in their model by assuming a dunitic oceanic SLUM of lower density than the present SLUM (Dziewonski et al., 1975). Sleep and Windley (1982) model 20 km of crust based on a greater depth of

generation of melt. They assume that all material above the melt-generating depth is either melt or residue, such that the thickness of the oceanic crust is the depth of generation times the fraction of segregated melt. Two problems are apparent. First, the thermal (cooling) thickening of the oceanic lithosphere as it moves away from the ridge is ignored. Second, although the depth of melt below a thermal boundary layer does increase with increasing mantle temperature (models 3a, 3g: Richter, 1985), the thermal structure of the mantle further dictates that it also decrease, creating a zone of melt generation (see Figure 5). The minimum depth of melt generation in the calculation of Sleep and Windley (1982) infers a decreasing crustal thickness with time, rather than the increase which they suggest. According to the present study, both of these models require positive-buoyant oceanic lithosphere which precludes the negative-buoyancy subduction inferred by the authors.

Baer (1981) assumes a lithospheric thickness of 57 km necessary to preclude negative-buoyancy subduction, from which he concludes that plate tectonics could not have been operative prior to 1.0 Ga. By comparison with McKenzie's (1969) equation, it is deduced that Baer (1981) ignores the time-variation of mantle temperature and density.

TECTONIC STYLE: DISCUSSION

Adopting the preferred model of a K-depleted earth (Bott, 1982) (which minimizes oceanic heat loss), the relation between oceanic and continental plate velocities, and paleomagnetic

results, we arrive at an oceanic velocity of 12 cm a^{-1} from 97,000 km of ridge (twice the present length), subducting at 22-33 Ma, in the earliest Archean. Since the apparent polar wander curves for the African, Australian, and Laurentian shields do not fit, these cratons were not combined as a super-continent in the Precambrian (Dunlop, 1981). Extending this to all Precambrian shields and examining the present configuration of the plates, the bulk of the excess ridge necessary to the model is accomplished by separating the cratonic blocks and surrounding each by spreading ridges. The greater ridge length may also be created, as noted earlier, by a reduction in plate size, which would maintain the asymmetry of tectonic regimes with respect to the oceanic margins.

This demonstrates that the same processes of plate tectonics acting today certainly could have been operating in the Proterozoic and Archean. There is therefore no obvious change in tectonic processes at the Archean-Proterozoic boundary. Nevertheless, models proposing positive-buoyancy as a precursor to negative-buoyancy cannot be excluded. The immediate result of this subduction of buoyant oceanic lithosphere would be shallowly dipping Benioff zones beneath the cratons, defined by the non-sinking oceanic lithosphere. This would be analogous to present observations of the descending Nazca plate beneath central Peru (Barazangi and Isacks, 1979). Since slab pull would not be the dominant driving force, the combination of ridge push and trench suction (Forsyth and Uyeda, 1975; Carlson, 1983) would have to be sufficient to overcome the mantle drag force in order to drive subduction.

Total resisting forces would be lower due to the decreased mantle viscosity, antecedent to increased mantle temperature.

Sacks (1983) notes that subduction will also proceed if the subducting slab is in contact with sufficiently hot adjacent lithosphere (over-riding slab). The Richter (1985) model of the thermal structure of the continental lithosphere, based on the hypothesis that it does not act as a thermal boundary layer, shows that the temperature at the base of the lithosphere varies by several hundred degrees in geologic history. Therefore the Sacks hypothesis could also be invoked as a driving force for normal subduction.

Subduction of buoyant oceanic lithosphere invokes an interesting mechanism to generate continental thickening where incomplete positive-buoyancy subduction could result in underplating of the proto-continent with oceanic lithosphere. This is not incompatible with continental SCUM densities (Dziewonski et al., 1975) or modelled continental SCUM compositions (Kay and Kay, 1981).

A scheme of vertical, or fixist, tectonics prior to the initiation of negative-buoyancy subduction is possible. Obviously, some mechanism other than stable continental and mobile convective oceanic heat dissipation is necessary. The alternate mechanism is necessarily extrusion of magma, which provides a subcrustal to supracrustal heat conduit. By this mechanism the heat flux is on average independent of the time since extrusion, therefore maintains its highest values. A simple calculation of heat dissipation (equation 5a) assuming a 500 Ma residence time averaged over the entire terrestrial

surface shows that this mechanism is indeed sufficient to dissipate the excess Archean heat. However, the crust generated would represent a thermal boundary layer, characterized by volcanism induced by melting under a subsiding lithosphere, which Richter (1985) has shown is unlikely for the Archean continental structure, based on geothermobarometric estimates. However, this scheme of vertical tectonics is a possible mechanism of heat dissipation prior to the accumulation of large quantities of continental volume.

The Hargraves (1981) model proposes a tectonic evolutionary sequence from a vigorously convecting mantle above which no crust survived, through vertical tectonics (surface extrusion balanced by sub-lithospheric melting), to negative-buoyancy subduction. This scheme of tectonic process evolution may apply regardless of the initiation time of negative-buoyancy subduction.

Glikson (1972) proposes one possible scenario for the transition from fixed vertical tectonics to negative-buoyancy mobile tectonics. Richter (1985) concludes that such a transition from crust defining a thermal boundary layer to chemically distinct crust of constant thickness would increase the likelihood of survival of the crust, and would be reflected in the area-age distribution of the present-day continents. Continental growth models (Figure 3) indicate an increase in the survivability of continental crust between 3.5 and 2.5 Ga. However, this may also be explained by the changing relative importance of the competing processes of crustal creation and destruction (O'Nions et al., 1979; Allègre and Jaupart, 1985; Veizer and Jansen, 1985).

CONCLUSIONS

Based on numerical modelling of Archean tectonics the following conclusions have been drawn:

1. Models of the thermal history of the earth, and therefore the characteristic tectonic style of the Archean, are dependent on the bulk earth composition modelled. These thermal models are largely independent of the following:

- a) continental thermal regime: geothermal gradient
- b) continental thermal regime: continental accretion models
- c) mantle rheology.

Better definition of these three parameters will therefore not distinguish between various Archean tectonic styles.

2. The thermal characteristics of the K-depleted earth model are completely explained by mobilist tectonics driven by negatively-buoyant subduction, regardless of the composition of the oceanic crust (komatiitic vs. basaltic). Basaltic crust could have had a maximum thickness of 5 km and komatiitic crust of 10 km for this tectonic style to have been operative in the Archean. However, this does not preclude other styles of Archean tectonics.

3. K-depleted earth: For the preferred model (maximum spreading activity), if paleomagnetic velocity determinations are accepted, the limiting cases are:

- A. The continents acted as relatively immobile rafts in faster spreading oceans, with ridge length comparable to present ridge length.

B. Oceanic velocity matched continental velocity, with five times the present ridge length in the early Archean.

4. The chondritic earth model would necessitate a change in the composition of the oceanic crust from basaltic today to komatiitic in the Archean if mobilist tectonics driven by negative-buoyancy is assumed. Otherwise, a change in tectonic style is necessitated near the Archean-Proterozoic boundary. In this case the Archean would have been characterized either by subduction of buoyant oceanic crust or by some other scheme of tectonics, for example vertical tectonics. The possibility of an Archean vertical tectonic scheme in which continental crust acts as a thermal boundary layer has been argued against by Richter (1985).

5. The higher mantle temperatures inherent to the non-Newtonian mantle rheology, relative to the Newtonian mantle rheology, are more conducive to extensive komatiite extrusion.

Appendix 1A: DERIVATION OF TOTAL OCEANIC HEAT LOSS FUNCTION

Oceanic surface heat flux as a function of the age of the lithosphere, t' , is given by

$$q_{oc}(t') = k \Delta T_* (\pi \alpha t')^{-1/2}$$

(Turcotte and Schubert, 1982)

where ΔT_* is the temperature difference between the sublithospheric upper mantle (SLUM) and the top of the crust. The total heat loss of lithosphere from age $t' = 0$ (ridge) to $t' = t_*$ (subduction), if v and l are opening rate and ridge length, is

$$Q_{oc}(t_*) = \int_0^{t_*} v l k \Delta T_* (\pi \alpha t')^{-1/2} dt'$$

Assuming v , l , k , α , ΔT_* approximately constant for a given t , then

$$Q_{oc}(t_*) = 2 v l k (\pi \alpha)^{-1/2} \Delta T_* t_*^{1/2}$$

$$\text{but } A_{oc} = v l t_*$$

therefore

$$Q_{oc} = 2 A_{oc} k (\pi \alpha)^{-1/2} \Delta T_* t_*^{-1/2}$$

Therefore

$$t_*^{1/2} = 2 \frac{A_{oc}}{Q_{oc}} k (\pi \alpha)^{-1/2} \Delta T_*$$

which gives the age of the oceanic crust at subduction for a given time, t .

Appendix 1B: DERIVATION OF DEPTH DEPENDENCE OF THE MANTLE ADIABAT

The mantle adiabat (Figure 5) is given by

$$\frac{dT}{dz} = - \frac{\alpha g}{c} T$$

which, by separation of variables, may be expressed as

$$\frac{1}{T} dT = - \frac{\alpha g}{c} dz$$

Assuming α , g , and c independent of temperature and depth, we obtain, by integration,

$$\ln T \Big|_{T_{200}}^{T_s} = - \frac{\alpha g}{c} z \Big|_{200}^{2690}$$

$$\ln T_s - \ln T_{200} = - \frac{\alpha g}{c} (2690)$$

$$\ln T_{200} = \ln T_s - \frac{\alpha g}{c} (2690)$$

$$T_{200} = T_s \div \exp \frac{\alpha g}{c} (2690)$$

Appendix 1C: DERIVATION OF TEMPERATURE DEPENDENCE OF DENSITY

The coefficient of thermal expansion is given by

$$\alpha = \frac{1}{V} \left(\frac{\partial V}{\partial T} \right)_P \quad (i)$$

Volume is defined as

$$V = \frac{m}{\rho}$$

which differentiates

$$\frac{\rho}{m} \partial V + \frac{V}{m} \partial \rho = 0$$

Substituting this into equation (i)

$$\alpha = - \frac{1}{\rho} \left(\frac{\partial \rho}{\partial T} \right)_P$$

Separating the variables, and integrating

$$\alpha (T_1 - T_0) = \ln \frac{\rho_0}{\rho_1}$$

Solving

$$\rho_1 = \rho_0 \exp (\alpha \Delta T)$$

where $\Delta T = T_0 - T_1$

$$Q = Q_{oc} + Q_{cn}$$

$$Q_R(t) = Q_R(0) \exp(-\lambda t)$$

$$Q_{cn} = k \frac{dT_{cn}}{dz} A_{cn}$$

$$A = A_{oc} + A_{cn}$$

$$t_*^{1/2} = \frac{2k}{(\pi\kappa)^{1/2}} \frac{\Delta T_*}{Q_{oc}} \frac{A_{oc}}{Q_{oc}}$$

$$Q_m(0) \exp \left[\frac{\beta T_m}{n+2} \left(\frac{1}{T_s(0)} - \frac{1}{T_s} \right) \right] = Q_{Rm} - \phi_m \dot{T}_s$$

$$T_{slum} = T_s \div \exp \left[\frac{\alpha g}{c} (2690) \right]$$

$$A_{oc} = v l t_*$$

$$l_{max} = \frac{A_{oc}}{t_* v(0)}$$

$$v_{max} = \frac{A_{oc}}{t_* l(0)}$$

$$(\rho_{slum} - \rho_{cr}) d_{cr} = \frac{1}{2} \rho_{slum} \alpha T_{slum} d_{li}$$

$$\rho_{slum} = \rho_{slum}(0) \exp(\alpha \Delta T)$$

$$\frac{T_{slum} - T}{T_{slum} - T} = \operatorname{erf} \left[\frac{d_{li}}{2(\kappa t_*)^{1/2}} \right]$$

APPENDIX 3. Tables of Results

Time	Q_R	Q	$\frac{dT}{dz}_{cn}$	A_{cn}	Q_{cn}
[Ga]	[10^{12} W]	[10^{12} W]	[10^{-3} K m $^{-1}$]	[10^{14} m 2]	[10^{12} W]
	Chondritic	Chondritic	MINIMUM	MINIMUM	MINIMUM
0.0	30.7	44.	22.80	2.015	11.49
-0.5	37.0	51.	22.80	1.930	11.00
-1.0	45.2	59.	22.80	1.822	10.39
-1.5	55.9	70.	22.80	1.664	9.485
-2.0	69.9	84.	22.80	1.449	8.259
-2.5	88.5	102.	22.80	1.177	6.709
-3.0	113.	127.	22.80	0.864	4.925
-3.5	146.	160.	22.80	0.532	3.032
-4.0	191.	205.	22.80	0.212	1.208
-4.5	252.	266.	22.80	0.000	0.000
	K-depleted	K-depleted	MAXIMUM	MAXIMUM	MAXIMUM
0.0	27.4	41.	23.84	2.015	12.01
-0.5	30.1	44.	27.58	1.999	13.78
-1.0	33.5	47.	31.91	1.969	15.71
-1.5	37.7	52.	36.91	1.924	17.75
-2.0	43.2	57.	42.70	1.832	19.56
-2.5	50.5	64.	49.39	1.685	20.81
-3.0	60.3	74.	57.14	1.495	21.36
-3.5	73.7	88.	66.10	1.191	19.68
-4.0	92.6	106.	76.46	0.748	14.30
-4.5	120.	134.	88.45	0.000	0.00

Time [Ga]	Qoc [10 ¹² W]			
	Chondritic		K-depleted	
	MINIMUM	MAXIMUM	MINIMUM	MAXIMUM
0.0	33	33	29	30
-0.5	37	40	30	33
-1.0	43	49	32	37
-1.5	52	60	34	42
-2.0	64	75	37	49
-2.5	81	95	44	58
-3.0	106	122	53	69
-3.5	140	157	68	85
-4.0	191	204	92	105
-4.5	266	266	134	134

Chondritic Earth Model

Time	Ts	TSLUM	ρ SLUM
[Ga]	[K]	[K]	[10^3 kgm^{-3}]

Newtonian Mantle Model

0.0	2800	1600	3.43
-0.5	2832	1620	3.43
-1.0	2865	1639	3.43
-1.5	2901	1659	3.42
-2.0	2938	1680	3.42
-2.5	2976	1702	3.42
-3.0	3018	1726	3.42
-3.5	3069	1755	3.41
-4.0	3227	1846	3.40
-4.5	----	----	----

non-Newtonian Mantle Model

0.0	2800	1600	3.43
-0.5	2848	1629	3.43
-1.0	2900	1659	3.42
-1.5	2956	1691	3.42
-2.0	3018	1726	3.42
-2.5	3086	1765	3.41
-3.0	3168	1812	3.41
-3.5	3281	1877	3.40
-4.0	3517	2011	3.38
-4.5	----	----	----

K-Depleted Earth Model

Time	Ts	TS _{LUM}	ρ SLUM
[Ga]	[K]	[K]	[10 ³ kgm ⁻³]

Newtonian Mantle Model

0.0	2800	1600	3.43
-0.5	2823	1615	3.43
-1.0	2847	1628	3.43
-1.5	2871	1642	3.43
-2.0	2897	1657	3.42
-2.5	2924	1672	3.42
-3.0	2956	1691	3.42
-3.5	3005	1719	3.42
-4.0	3195	1827	3.41
-4.5	----	----	----

non-Newtonian Mantle Model

0.0	2800	1600	3.43
-0.5	2836	1622	3.43
-1.0	2875	1644	3.43
-1.5	2917	1668	3.42
-2.0	2964	1695	3.42
-2.5	3019	1727	3.42
-3.0	3090	1767	3.41
-3.5	3202	1831	3.41
-4.0	3460	1979	3.39
-4.5	----	----	----

Time [Ga]	t* [Ma]			
	Minimum Oceanic Heat Flux non-Newtonian	Heat Flux Newtonian	Maximum Oceanic Heat Flux non-Newtonian	Heat Flux Newtonian
<u>Chondritic Earth Model</u>				
0.0	80.2	80.2	80.2	80.2
-0.5	65.1	64.2	60.6	59.8
-1.0	50.7	49.2	45.5	44.2
-1.5	38.0	36.3	34.2	32.6
-2.0	27.8	26.0	25.8	24.1
-2.5	20.0	18.3	19.6	18.0
-3.0	13.9	12.4	14.9	13.3
-3.5	10.2	8.7	11.4	9.7
-4.0	8.0	6.5	9.1	7.4
-4.5	-----	-----	-----	-----
<u>K-depleted Earth Model</u>				
0.0	80.1	80.1	79.8	79.8
-0.5	78.4	77.6	70.8	70.1
-1.0	75.0	73.2	62.3	60.8
-1.5	70.1	67.4	54.8	52.7
-2.0	62.9	59.6	47.7	45.1
-2.5	53.2	49.2	41.2	38.0
-3.0	42.7	38.4	35.3	31.7
-3.5	33.0	28.3	29.9	25.7
-4.0	26.8	22.2	26.7	22.1
-4.5	-----	-----	-----	-----

 MAXIMUM OCEANIC HEAT FLUX

Time [Ga]	v_{MAX} [10^{-2} my^{-1}]	l_{MAX} [10^7 m]	dlt [km]
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 Chondritic Earth Model

0.0	6.9	5.3	79
-0.5	9.5	7.3	67
-1.0	13.3	10.2	57
-1.5	18.8	14.6	48
-2.0	27.1	20.9	40
-2.5	38.9	30.0	34
-3.0	56.9	44.0	28
-3.5	84.1	64.9	24
-4.0	118.	91.1	19
-4.5	----	----	--

 K-depleted Earth Model

0.0	6.9	5.3	86
-0.5	8.0	6.2	79
-1.0	9.4	7.3	72
-1.5	11.2	8.7	66
-2.0	13.7	10.6	60
-2.5	17.0	13.1	54
-3.0	21.4	16.6	48
-3.5	27.3	21.1	42
-4.0	32.7	25.3	36
-4.5	----	----	--

 MINIMUM OCEANIC HEAT FLUX

Time [Ga]	v_{MAX} [10^{-2} my $^{-1}$]	ℓ_{MAX} [10^7 m]	dli [km]
Chondritic Earth Model			
0.0	6.9	5.3	79
-0.5	8.6	6.6	70
-1.0	11.4	8.8	60
-1.5	15.6	12.1	50
-2.0	22.5	17.4	42
-2.5	46.0	25.8	34
-3.0	52.0	40.2	28
-3.5	80.2	62.0	22
-4.0	119.	92.3	18
-4.5	---	---	--

 K-depleted Earth Model

0.0	6.9	5.3	86
-0.5	7.1	5.4	83
-1.0	7.5	5.8	79
-1.5	8.1	6.3	75
-2.0	9.3	7.2	69
-2.5	11.5	8.9	62
-3.0	15.1	11.6	53
-3.5	21.2	16.3	44
-4.0	29.0	22.4	36
-4.5	---	---	--

 MINIMUM THICKNESS OF BUOYANT OCEANIC CRUST

Time [Ga]	THICKNESS (km)	
	Basaltic Crust	Komatiitic Crust
0.0	6.3	14.
-0.5	5.4	12.
-1.0	4.7	11.
-1.5	4.1	9.5
-2.0	3.4	8.0
-2.5	2.9	6.9
-3.0	2.5	5.8
-3.5	2.2	5.3
-4.0	1.9	4.6
-4.5	---	---

 K-depleted Earth Model

0.0	6.0	16.
-0.5	6.4	15.
-1.0	5.9	14.
-1.5	5.6	13.
-2.0	5.2	12.
-2.5	4.8	11.
-3.0	4.4	11.
-3.5	4.0	10.
-4.0	4.0	10.
-4.5	---	---

APPENDIX 4: Compilation of Trench Data

Trench	Length (10^3 km)	Mean Age of Subduction (Ma)
Kuriles	2.111	100
North Honshu	0.938	112.5 *
Izu-Bonin	1.220	135 *
Marianas	2.283	150
Southwest Honshu	1.016	25
Ryukyus	1.407	60
Sumatra	2.846	65 *
Java	3.471	127.5 *
Tonga 20 S	0.751	125 *
Tonga 24 S	0.500	120 *
Kermadec	1.736	120 *
New Zealand N	0.547	120 *
New Zealand S	0.954	120 *
Chile 42-46 S	0.516	15
Chile 36-42 S	0.719	35
Chile 27-33 S	1.110	54 *
Chile 18-26 S	0.954	61 *
Petu	2.736	55 *
Central America 87 W	0.688	40
Central America 92 W	0.532	40
Mexico 102 W	1.423	25 *
Cascades	1.188	10
Alaska 150 W	1.016	37.5 *
Aleutians 180 W	2.533	56.5 *
Lesser Antilles	1.454	75 +
South Sandwich Islands	1.110	55 +

Total = 35.76×10^3 m

Length Averaged Mean = 79.77 Ma

Age data from Wortel (1947), excluding Mindanao, New Britain, Solomon Islands and New Hebrides.

Trench lengths as measured on world tectonic globe.

* Average of subduction age as given by Wortel (1947).

+ Back-arc spreading ignored.

APPENDIX 5: Compilation of Ridge Data

Ridge	Mean Rate of Spreading (Ma)	Length (10^3 km)
Arabian - African	4.1	0.26
Indian - Arabian	1.3	0.47
Eurasian - North American	7.4	0.23
African - South American	8.0	0.36
North American - Pacific	3.6	0.85
Cocos - Pacific	2.8	2.21
Cocos - Nazca	2.1	0.97
Nazca - Pacific	4.2	1.54
Nazca - Antarctic	1.3	0.60
Antarctic - South American	1.1	0.30
Antarctic - Pacific	5.3	0.96
Indian - Antarctic	7.1	0.67
Indian - African	3.9	0.64
African - Antarctic	5.3	0.15
African - North American	1.8	0.26

Total length = $59.3 * 10^3$ km

Total spreading activity = $38 * 10^5 \text{ m}^2 \text{ a}^{-1}$

Length Averaged Mean = $7.25 * 10^{-2} \text{ m a}^{-1}$

Angular velocity data taken from Minster and Jordan (1978).

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