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Growth and Differentiation of the Continental Crust

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Growth and differentiation of the continental crust

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Declining radiogenic heat production since the Archaean has resulted in a secular evolution from a régime of numerous fast-moving small thin torsionally weak plates to the present régime of larger thicker torsionally stronger plates moving at an average rate of less than one-sixth of the Archaean rate; this has been accompanied by episodic changes in geological effects. By 2500 Ma B.P., about 85% of the present crustal mass had grown by the addition and amalgamation of mafic and calc-alkaline rocks in oceanic arcs at an average rate of 11.17 Pg/a. During the early Proterozoic, the first large cratons were stabilized and, locally, thickened and differentiated; the Proterozoic was an era of little continental growth, falling average sea level, and intra-continental deformation. By 700 Ma B.P. cratons had become much more stable, marginal accretionary terrians had begun to develop with an average Phanerozoic growth rate of 1.64 Pg/a, and blueschists and ophiolites *sensu stricto* witness the advent of the extant plate tectonic régime.

1. Introduction

The differentiation of the continental crust, enriched in light, large-ion-lithophile and radioactive elements, from the mantle since 3600 Ma B.P. has been a consequence of global heat loss, which has been accomplished by the creation, cooling, thickening and subduction of a boundary conduction layer segmented into an evolving mosaic of lithosphere plates. The continental crust is the expression of the most efficient mode of global heat loss, the concentration of the heat-producing elements in an outer layer. Differentiation from the mantle in a plate tectonic setting has occurred in the following way: incorporation of oceanic basalts and minor amounts of ultramafic rock to form the basement and core of ensimatic volcanic arcs and minor ophiolite slivers in subduction–accretion prisms, arc basalts and andesites and their silicic differentiates produced by partial melting of ultramafic rocks and basalts of the subducting slab and the overlying asthenosphere wedge (Ringwood 1974), and basalts underplated in intracontinental graben systems. Further intracrustal differentiation occurs in the continents by the upward migration of the large-ion and radioactive elements in zones of crustal shortening and extension to produce an upper crust enriched in these elements and a more mafic and refractory granulitic lower crust depleted in minimum-melting components (Dewey & Burke 1973). Thus the large-ion-lithophile elements are concentrated into the sedimentary veneer and are therefore preferentially involved in shallow recycling of the continental crust through arc systems by subduction–accretion, subduction underplating and the possible contribution of subducted sediments to arc magmas. Furthermore, any sediment subducted and homogenized with the mantle by convection will possess higher concentrations of K, U, Pb, Rb, etc. than their average abundances in continental crust.

The oceanic crust is cycled by plate-accretion and minor intraplate vulcanism, balanced by subduction, in approximately 170 Ma at a rate of about 20 km^3/a (580×10^{14} g/a). This rapid recycling rate is due to the ease with which oceanic lithosphere is subducted; although the oceanic crust is gravitationally stable, most of it remains firmly attached to the negatively buoyant ultramafic part of the lithosphere as it sinks into the asthenosphere. Furthermore, the short-term strength of the oceanic lithosphere is shown by its low proportion of intraplate seismicity and vulcanism, by well defined narrow intraoceanic plate boundaries and by rifts rarely developing in the oceans, the Nazca–Cocos boundary and the Caroline–Truk Plateau being the only apparent extant exceptions.

In contrast, the continental crust contains rocks as old as 3800 Ma and, if it has occurred, recycling has proceeded at an order of magnitude slower, at an average rate, since 3600 Ma B.P. of 1.46 km^3/a (4.02 Pg†/a). This stems from the greater thickness, buoyancy and weakness of the continental crust, illustrated by the greater width and complexity of plate boundary zones in continents, a higher density of intraplate seismicity and vulcanism, the origination of most rift systems and ridges within continents, and the general ease with which continental crust is structurally rejuvenated and remobilized by shortening/thickening and stretching/thinning. However, continental deformation is clearly inversely related to the age of the continental crust. The pre-2500 Ma B.P. shield areas that were not affected by early and mid-Proterozoic remobilization are singularly free from later reactivation and seismicity and also have lower heat-flow values than remobilized Archaean and younger crust. Beneath the continental crust must lie a layer of variable thickness, of depleted mantle, the refractory residue, relatively poor in clinopyroxene, garnet, spinel and plagioclase and depleted in large-ion-lithophile and radioactive elements, from which the continental crust has been extracted. At shallow mantle levels, this olivine/orthopyroxene-enriched lithosphere is more bouyant than undepleted mantle and acts as a mechanically strong 'floater' (Jordan 1978) that may help resist subduction of superjacent continental crust. Such a thick depleted mantle lithosphere may help to insulate the old shield areas from the ravages of continental deformation and perhaps to provide protected mantle source regions that allow strontium inheritance from ancient lithosphere (Brooks *et al.* 1976). Only where the lithosphere is thinned by extension zones can undepleted mantle regain access to the base of the crust.

Greatly disparate views have been expressed on the basic problem of whether the continental crust has increased or decreased its mass and/or area or perhaps maintained a constant mass since about 2500 Ma B.P. Hargraves (1976), Lowman (1976) and Shaw (1976) argued for constant post-Archaean mass, and Armstrong (1968 and this symposium) advocated such models with large-scale deep recycling of continental materials through the mantle. Constant mass was supported by Wise (1974) from freeboard (elevation of continents with respect to sea level) arguments. Hurley *et al.* (1962) argued a quasi-continuous steady growth in continental mass and Ringwood (1974) suggested that such growth is a natural consequence of calc-alkaline magmatism in arc systems. Fast early growth with a peak in, and with about two-thirds of the present continental mass generated by the end of, the Archaean, followed by steady growth, was supported by McCulloch & Wasserburg (1978), O'Nions & Pankhurst (1978), Tarney & Windley (1977) and Veizer & Jansen (1979), mainly from isotope systematics. Sutton (1963) Moorbath (1978) and Moorbath & Taylor (1980) advocated, from Sr and Pb systematics, an irreversible but episodic mass gain, with short periods of accelerated mass growth, with the largest gains in the Archaean. Hurley & Rand (1969) proposed a model of accelerating con-

† 1 Pg = 10^{15} g.

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tinental growth. More recently, Fyfe (1978) outlined a model of progressive mass loss from the continents.

Models involving no addition of new material from the mantle but only shallow recycling of Archaean materials may be rejected out of hand on isotopic and regional tectonic evidence. The central question addressed in this paper is whether the mass of the continents has increased, decreased or remained constant since the period of late Archaean rapid growth of almost 85% of their present mass. Although it is now generally agreed that mechanisms exist by which continental crust is differentiated from, and returned to, the mantle, the extent, balance, timing and rates of these processes are in dispute.

It is appropriate at this stage briefly to list terms and define their meanings as used in this paper.

(i) *Addition*: the transfer of mass from the mantle to the continent.
(ii) *Subtraction* (*deep recycling*): the return of mass from the continents to be homogenized with the mantle.
(iii) *Growth*: net gain of continental mass.
(iv) *Diminution*: net loss of continental mass.
(v) *Accretion*: area increase by addition or redistribution at edges; *subduction–accretion* (Karig *et al.* 1978) is an important form of this process.
(vi) *Decretion*: area decrease by removal of material from edges; *subduction erosion* or *tectonic erosion* is an important method.
(vii) *Overplating* and *underplating*: increase in crustal thickness by emplacement at the top (e.g. ophiolite obduction) or base, respectively, of the crust. *Overaccretion* and *underaccretion* (Wells 1980) are a magmatic version of this process.
(viii) *Stretching and shortening*: area increase and decrease, accompanied by tectonic thinning and thickening, respectively.
(ix) *Enlargement*: increase in continental area.
(x) *Reduction*: decrease in continental area.
(xi) *Redistribution*: lateral transfer and rearrangement of crustal materials by the erosion–sedimentary cycle or by structural means (e.g. thrust stacking).
(xii) *Shallow recycling*: subduction of continental materials to subcontinental depths and reincorporation into the continent mainly in arc magmas.
(xiii) *Remobilization* (reworking): tectonic reactivation or rejuvenation, of previously stabilized continental crust.
(xiv) *Reconstitution*: pervasive recasting of one lithology to another.
(xv) *Cratonization*: stabilization of a continental crust sufficiently to allow the accumulation of thin epicontinental sedimentary sequences.
(xvi) *Differentiation*: the development of an upper potassic crust and a lower more refractory crust from a more uniform protocrust (maturation and ripening are terms that have also been used).

2. Addition, modification and subtraction processes

Figure 1 is a schematic summary of the processes contributing to addition, subtraction, accretion, decretion, redistribution, shallow recycling, remobilization and differentiation.

(a) *Arcs*

Accretion takes place dominantly in extensional and neutral arc systems (figure 2; Dewey 1980) such as those of Mesozoic/Cainozoic age in the Caribbean and west Pacific regions. Considerable parts of such arc systems have a modified ophiolitic basement or core (the Nicoya Complex of Costa Rica, the Bermeja Complex of Puerto Rico and the Demisseaux Complex of Haiti) and a total crustal thickness varying from 12 to 25 km. Basically, such arcs are mafic/intermediate volcanic assemblages with a subjacent comagmatic plutonic infra-

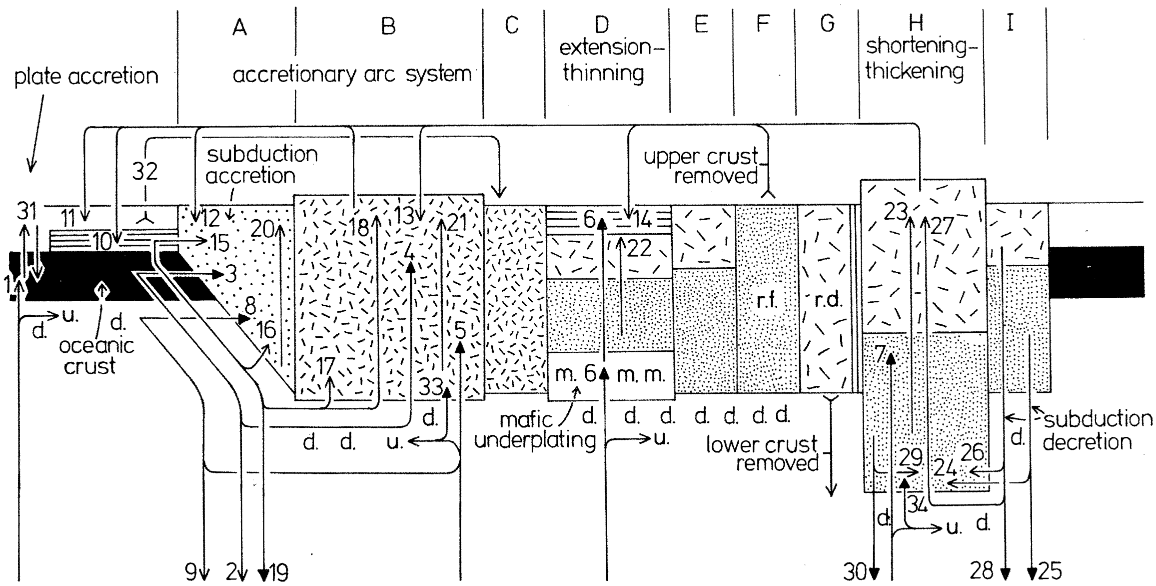

FIGURE 1. Schematic illustration of the processes that add to, subtract from and redistribute continental mass: A, subduction–accretion prism; B, magmatic arc and associated oceanic basins; C, undifferentiated Archaean crust; D, stretched and differentiated crust underplated by mafic material (m.) and overplated by sediment in graben systems; E, differentiated crust returned to normal thickness by erosion and subcrustal delamination; F, differentiated crust with upper crust removed by erosion to leave refractory granulitic crust of normal thickness (r.f.); G, differentiated crust with lower crust removed by delamination to leave a radiogenic upper crust removed by delamination to leave a radiogenic upper crust rich in incompatible elements (rd.); H, crust involved in shortening and thickening in Archaean or collision orogenic belt with differentiation into a radiogenic upper crust and a refractory lower crust; I, crust suffering decretion/delamination. Key to shading: solid black, oceanic crust; coarse stipple, subduction–accretion prism; horizontal lines, sediment; fine random lines, magmatic arc terrains; coarse random lines, differentiated radiogenic upper crust; fine stipple, differentiated refractory lower crust. Key to numbers: 1, partial melting and extraction of basalts to leave ultramafic rocks (u.) as a depleted harzburgite uppermost mantle (d.); 2, oceanic crust subducted into mantle; 3, oceanic crust slivers incorporated into subduction–accretion prism; 4, contribution to arc magmas from subducted oceanic crust; 5, contribution to arc magmas from mantle; 6, basalt addition to graben systems; 7, basalt addition to Andean-type arcs; 8, ultramafics incorporated into subduction–accretion prisms; 9, depleted ultramafics returned to asthenosphere, 10, pelagic oceanic sediments; 11, mass in solution in sea water; 12, coarse subduction–accretion sediments; 13, continent-derived sediment in arc-related oceanic basins; 14, sediment in graben systems; 15, pelagics incorporated into subduction–accretion prisms; 16, pelagics partially subducted and added to base of subduction–accretion prism; 17, pelagics underplated to base of magmatic arc; 18, pelagics contributing to arc magmas; 19, pelagics subtracted; 20, upward diapiric flow of sediment in thickest part of accretionary prism; 21, minor differentiation in magmatic arc; 22, differentiation in graben system; 23, differentiation in Andean or collisional systems; 24, underplating by decreted lower crust; 25, subtraction of lower crust; 26, underplating by decreted upper crust; 27, contribution to magmas from decreted upper crust; 25, subtracted upper crust; 29, structural redistribution of delaminated thickened lower crust; 30, subtraction of delaminated thickened lower crust; 31, exchange between sea water and the upper oceanic crust; 32, removal of chemical and biogenic sediments from sea water.

structure built upon an oceanic foundation, which comprises about one-third of the crustal thickness. Volcaniclastic dispersal fans spread quartz-poor detritus into adjacent fore-arcs, trenches and intra/inter- and rear-arc basins. Thus, additions in the extensional and neutral arc environments consist of mafic and minor ultramafic rocks of the arc foundation, generated by earlier plate accretion (figure 1, 1), possible andesitic partial melts from the basalts of the subducting lithosphere (figure 1, 4), basaltic/andesitic partial melts of the ultramafic rocks of the subducting lithosphere and the sub-arc asthenosphere wedge (figure 1, 5) and ophiolite fragments incorporated into subduction–accretion complexes (figure 1, 3 and 8). Some extensional/neutral arcs, particularly those with crusts from 25–40 km thick, have a partial, if not complete, older continental basement in the magmatic arc.

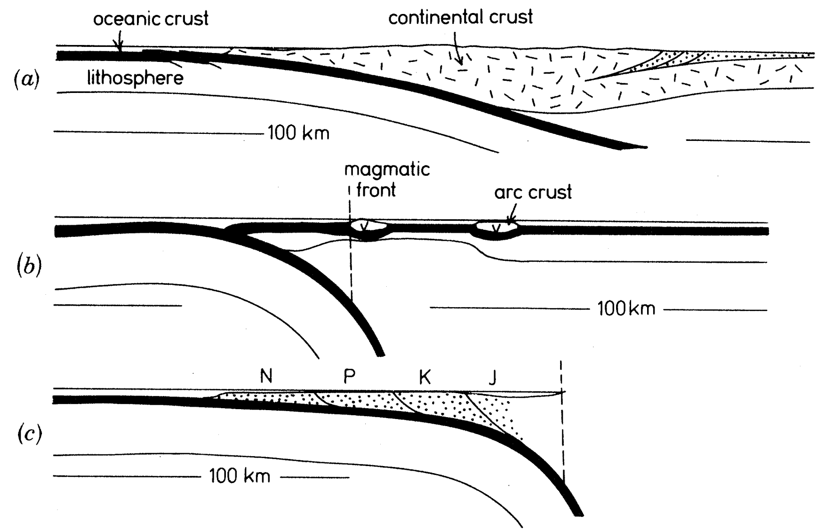

FIGURE 2. Simplified true-scale cross sections of (*a*) central Peruvian Andes, (*b*) Marianas arc and Parece Vela Basin, and (*c*) Alaskan accretionary prism.

Extensional arcs, in particular, lay down a lace-like web that forms a trap for the redistribution of older continental material as sediment piles in intra-, inter- and rear-arc basins (e.g. Irrawaddy fan in the Andaman Sea), and in subduction–accretion prisms. Subduction–accretion wedges are composites of a minor component of oceanic pelagic sediment accreted directly (figure 1, 15), a possibly larger pelagic component partially subducted and added to the base of the prism, a dominant fraction of coarser clastics accreted from trench turbidites, and upper slope basin sediments (figure 1, 12). Crustal thicknesses up to at least 30 km can be built by subduction–accretion alone (figure 2*c*). Subsequent shortenings of accretionary arc complexes by entrapment in collisional zones or back-arc basin closing (Dalziel *et al.* 1974) lead to telescoping, crustal thickening and further differentiation of a complicated collage of added and redistributed mass. Most post-Archaean accretionary terrains appear to be latest Proterozoic and Phanerozoic; the best examples are the Tasman Belt of New Zealand and eastern Australia, the Cordilleran Belt of western North America west of the Nevada Palaeozoic carbonate platform, and the late Proterozoic greenstone belts of Saudi Arabia and north Africa.

Andean arc/orogenic systems (figure 2; Dewey 1980) differ from extensional west-Pacific arcs in their thick (up to 75 km) largely older continental crust, more silicic magmatism, redistributed silicic clastics in foreland basins over which the back-arc region is thrusted, and a fore-arc with continental basement rocks. Evidence for decretion (subduction erosion) is fairly clear in the Peruvian segment, where subduction zone dip is very shallow, where the continental forearc crust is as little as 10 km thick (figure 2*c*) and where the late Palaeozoic accretionary terrains are missing between Ecuador and Chile. Subduction-erosion, particularly that resulting from subduction dip flattening (Dewey 1980), preferentially decretes the lower crust, but upper crustal components must also be involved in large-scale decretion. The post-decretion fate of tectonically eroded crust is not clear: whether some is underplated to aid formation of a thick crust (figure 1, 24; figure 2, 2*b*), is subtracted back into the mantle reservoir (figure 1, 25, 28), or contributes to the silicic magmas of the arc. The role of the asthenosphere ultramafic wedge in contributing directly or indirectly to arc magmatism is, however, clear. In the Andes, a very gently dipping Benioff zone with no asthenosphere wedge in central Peru has no overlying extant volcanic arc (Barazangi & Isacks 1976). Thus, even if large amounts of decreted crust and perhaps subducted sediment are being carried beneath central Peru, they do not appear to be currently contributing to arc magmatism, as neither does the subducting oceanic crust, unless sill-like batholithic sheets are being injected at depth. The nature and volumes of additions to Andean-style arcs are much more difficult to determine than those to extensional arcs. Even if all Andean arc magmas are pure additions from mantle or oceanic crust, with no contribution from pre-existing continental crust, the shape and extent of subvolcanic plutonic complexes are in sufficient dispute to prevent accurate assessments of overaccretion/underaccretion ratio. Current thinking (Francis & Rundle 1976; Brown 1977; Thorpe & Francis 1979; Thorpe, this symposium) is that a rate of volcanic production of about 1.5 $km^3/(km\ Ma)$ is a small percentage of the total rate of magma production. The crustal root beneath Peru has a volume of about 7000 km^3 per kilometre of arc, giving a minimum formation rate of about 44 $km^3/(km\ Ma)$ since the early Jurassic. If the whole of the root is composed of Mesozoic/Cainozoic batholiths, the plutonic/volcanic ratio is therefore about 28. However, part of the root may have resulted from shortening and thickening of an earlier crust during Cainozoic times (Dewey 1980). Also, about 3500 km^3 per kilometre of arc has been removed from the deep and intermediate crust beneath southwestern Peru (Dewey 1980). Some of this may have contributed to crustal root underplating, as may also have some of the 7000 km^3/km decreted from Peru since early Jurassic times. In view of these great uncertainties, addition rates in Andean-type arcs are very difficult to determine. However, it is difficult to escape the conclusion that subtraction has exceeded addition in the Peruvian segment since the early Jurassic. Some 66 $km^3/(km\ Ma)$ have been removed by subduction erosion, whereas the maximum possible addition rate has been about 44 $km^3/(km\ Ma)$, and therefore subtraction must outstrip addition here by at least 50 %.

(*b*) *Graben systems*

Substantial addition and differentiation occurs in intracontinental extensional terrains such as the Rhine Graben, the Aegean, the Basin and Range system and those that formed the loci of continental fragmentation and separation. In these graben complexes, rapid stretching and crustal thinning is accompanied by consequent isostatic subsidence (McKenzie 1978), the overplating of very thick, rapidly deposited sedimentary sequences (figure 1, 14), and an increase in geothermal gradient. Progressive cooling and return to a normal geothermal gradient yields a further slow and declining subsidence, producing more widespread sedimentary basins without much fault control and with thinner sedimentary sequences, During the early stretching phase, basaltic addition is considerable (figure 1, 6). Flood basalts and associated dyke swarms at shallow levels in the crust imply very considerable injection of gabbro and underplating in the lower crust (figure 1) to thermally 'pave the way' for basalt migration to the upper crust by preventing freezing at deeper levels. The ferrogranite (riebeckite, aegerine, and fayalite-bearing) and undersaturated alkaline syenite suites, sometimes with carbonatites, are the bimodal partners of mafic rocks in rift systems and may imply substantial crustal differentiation by partial melting of the lower crust and the consequent supra- and prebasaltic rise of buoyant silicic/intermediate liquids to enhance the alkalinity of the upper crust and the refractory nature of the lower crust (figure 1, 22).

All transitions in the sequence from rifts with minor crustal thinning and mafic intrusion, through severely attenuated crust, to oceanic plate accretion *sensu stricto* exist in the extensional régime. Up to about 15 km of sediment can accumulate on a thermally subsided and loaded oceanic lithosphere or a lithosphere with continental crust so thinned and stretched that it approaches oceanic behaviour. Thus, the extensional environment can generate crustal sections ranging from an unmodified 35 km crust to sections 25 km thick consisting of about 10 km of mafic lower crust with minor boudins and slivers of older continental crust overlain by 15 km of sediment. Crustal sections of this type are developed in a wide range of tectonic settings, such as rifted continental margins, minor pull-aparts on irregular intracontinental transforms in convergent régimes, and filled remnant oceanic basins between irregular, colliding continental margins.

(*c*) *Collision*

Continental collision results in the thrust-stacking of the commonly thinned edges of colliding continents. The contracted ocean may have been a wide oceanic tract generated by plate accretion or a narrow pseudo-oceanic region floored by a severely attenuated continental crust or thinned continental slivers and xenoliths in a mafic/ultramafic complex too little ordered to be called an ophiolite complex *sensu stricto*. Thrust stacking of an attenuated continental crust with oceanic or pseudo-oceanic remnants (Milnes 1978) leads to crustal thickening and the telescoping and juxtaposing of hotter over colder wedges of crust, giving the common granulite/amphibolite over greenschist/blueschist, the decretion or shaving, and shallow subduction of giant slices of continental crust (e.g. Sesia zone of the Italian Alps (Dal Piaz *et al.* 1972)), and the emplacement of subcontinental mantle and lower crust over high-pressure terrains (e.g. Ivrea zone, Ronda Massif and Beni Bousera massifs (Kornprobst 1976). Large-scale structural redistribution of crustal components occurs, therefore, with individual intracontinental thrusts having up to 150 km of displacement (Gee 1975). Crustal low-velocity zones may aid intra-crustal flaking and detachment as thrust root zones and aid delamination of the lower crust (Bird 1978, 1980) and perhaps its subduction (Molnar & Gray 1979).

Minor minimum-melting granites such as the Bergell–Novate body of the Pennine root zone and the 'hair' granites of the Himalayas suggest small amounts of crustal differentiation. Extensive crustal differentiation (figure 1, 23) is suggested by the widespread high-potash dacites of the Tibetan Plateau, which are probably the result of partial melting of lower crustal rocks in the double-thickness crust produced by 50% shortening (Dewey & Burke, 1973) since collision with India in the late Eocene (Molnar & Tapponnier 1975).

Large-scale foreland reactivation results in graben systems at a high angle to the collisional interface and lateral escape of transform-bounded wedges towards the nearest oceanic 'free face' (Molnar & Tapponnier 1975). The transforms are major zones of shear reconstitution and also have compressional locking segments, where minor crustal thickening occurs, and

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extensional pull-apart segments, such as Lake Baikal, where severe extensional thinning or even separation of the continental lithosphere has occurred. In these and in remnant oceanic nodes, a 25 km crust could develop consisting of a 15 km sedimentary section over a 10 km mafic or attenuated continental crust.

The only clear additions to the continental crust of collisional orogenic systems are the basalts in graben systems at a low angle to the convergence direction and in pull-aparts along transforms; these basalts appear to be, volumetrically, a very minor component.

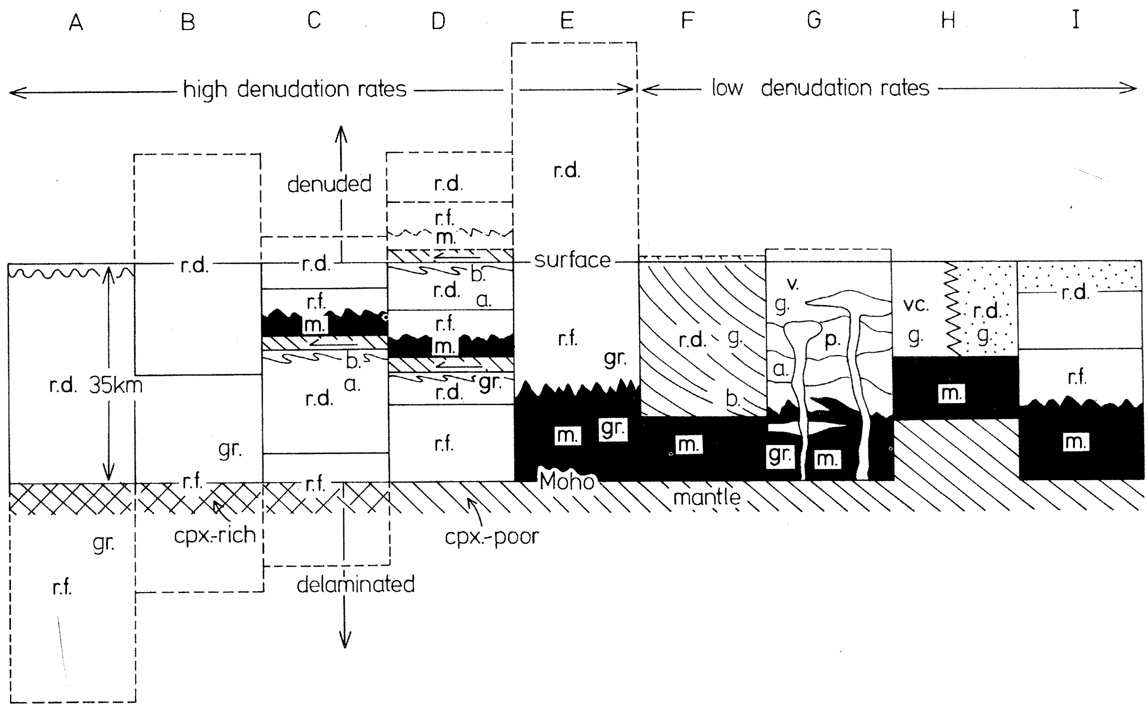

FIGURE 3. Schematic sections of various types of continental crust discussed in the text: A–E, crust thickened by shortening (A, B, E) and thrusting and (C, D) returned to normal thickness by erosion (B–E) and/or delamination (A–C); F, subduction–accretion crust (cf. figure 2*c*); G, magmatic arc crust, with plutonic (p.) and volcanic (v.) components; H, crust of rear or intra-arc basin with arc-derived volcaniclastic sediment (vc.) and/or continent-derived radiogenic sediment (rd.); I, crust of graben system. Key to shading and abbreviations; solid black shading (m.), mafic lowest crust; rd., radiogenic upper crust and sediments; r.f., refractory lower crust; a., amphibolite facies; b., blueschist facies; g., greenschist facies; gr., granulite facies.

(*d*) *Crustal trends*

Rogers & Novitsky-Evans (1977) argued for a general thickening of the crust in central Oregon from a 14 km oceanic plus sediment crust in late Triassic times to the present 35 km crust. Such a crustal thickening trend is consistent with the growth of Cretaceous/early Cainozoic arc complexes on a late Palaeozoic oceanic foundation, followed by minor shortening and the late Cainozoic overplating of 2 km of Columbia River basalt, that is, the roughly *in situ* thickening and/or cratonization of a continental crust. Such terrains are low-potash, low-metamorphic-grade with abundant supracrustals and calc-alkaline plutons and volcanics, show little differentiation, have crustal thickness of about 30 to 35 km and appear to have suffered little erosion. They are reminiscent of the granodiorite/greenstone belts of the Archaean (Tarney & Windley, this symposium) and, in particular, of the late Proterozoic of Saudi Arabia

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(Gass 1979). Characteristic of such terrains (figure 3*c*) in the Phanerozoic, such as the Tasman Belt, the Oregon–Idaho segment of the Cordilleran Belt, and the Notre Dame Bay region of central Newfoundland, is the mixing of components of addition and redistribution in accretionary systems and the growth of calc-alkaline arc complexes on and in obducted ophiolite sheets such as those of the Klamath Mountains in southwest Oregon. Thus the magmatic arcs and their ophiolite foundations are low, and the subduction-accretion terrains high, in radiogenic elements (figure 3).

Shortening in collision zones and, perhaps, in Andean-type orogens causes extensive crustal thickening and differentiation (figure 1, 23). Return to a normal thickness of crust is probably usually achieved by uplift and erosion but may occasionally or locally result from delamination and subduction of the lower crust (Bird 1978, 1980; Molnar & Gray 1979), particularly where the lower crust is refractory and mafic (figure 3). The mode of crustal thickening determines the degree to which supracrustals are metamorphosed and the nature of detritus eroded from the thickened crust. Roughly homogenous thickening by a vertical plane strain (e.g. Tibetan Plateau; Dewey & Burke 1973) involves either a remnant lower crust exposed by the erosion of the radiogenic upper crust (figure 3, E) or a remnant upper crust with preserved low metamorphic-grade supracrustals (figure 3, A). Thrust stacking of a previously thinned crust, as in the Alps (Milnes 1978), leads to a complicated interleaving of radiogenic and refractory crust with mafic lower crust and ultramafic mantle rocks (figure 3, C and D). Supracrustals are carried to great depths to become involved in blueschist, amphibolite and granulite facies metamorphism.

Crustal trends in extensional environemts are towards a thinned differentiated crust underplated by mafic igneous rocks and overplated by a radiogenic sediment blanket (figure 3, I) or, following plate accretion, towards a 25 km crust (figure 3, H) consisting of a mafic lower crust buried by radiogenic sediments and/or low-potash arc-derived volcaniclastics.

All these processes of generating, evolving and modifying the crust are overlaid and interplayed in complex temporal and spatial sequences resulting in a general crustal trend towards differentiation and the redistribution of upper crustal products rich in incompatible radiogenic elements because high denudation rates occur dominantly in areas of differentiated crust. Therefore, if substantial deep recycling of crustal rocks occurs by sediment subduction and minor delamination, radiogenic crustal materials are selectively returned to the mantle from the differentiating crust.

3. Rates of addition and subtraction

If the average thickness is taken to be 35 km, the continental crust has a volume of *ca.* 5.25×10^9 km^3 and, with an average density of $2.75\ g/cm^3$ ($2.75\ Pg/km^3$), a mass of 1.44×10^{25} g. We estimate that about 70 % of this mass consists of pristine Archaean terrains such as the Superior, Slave, Swazi and Yilgarn provinces, and remobilized Archaean terrains such as much of the Hudsonian Province of Manitoba. The remaining 30 % was accreted after 2500 Ma B.P. and consists of additional and redistributed material; we estimate that a maximum of about half of this material is redistributed Archaean crustal material. Therefore, about 85 % of the present mass of the continents grew during the interval 3600–2500 Ma (Moorbath & Taylor 1980), giving an average Archaean growth rate of $4.06\ km^3/a$ or 11.17 Pg/a. The average post-Archaean minimum addition rate of 15 % of the present continental mass is 0.88 Pg/a. We

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estimate that of that 15% only a small part is Proterozoic; there are very few accretionary granodiorite greenstone terrains of early to mid-Proterozoic age (for a few examples see Tarney & Windley (this symposium)) and they are volumetrically very small. The bulk of the 15% is of late Proterozoic and Phanerozoic age (700–0 Ma B.P.), giving an average addition rate during that interval of 3.16 Pg/a and an accretion rate of 6.27 Pg/a (figure 4). All the rates so far quoted are minimum values because possible subtractions are very uncertain for pre-Jurassic

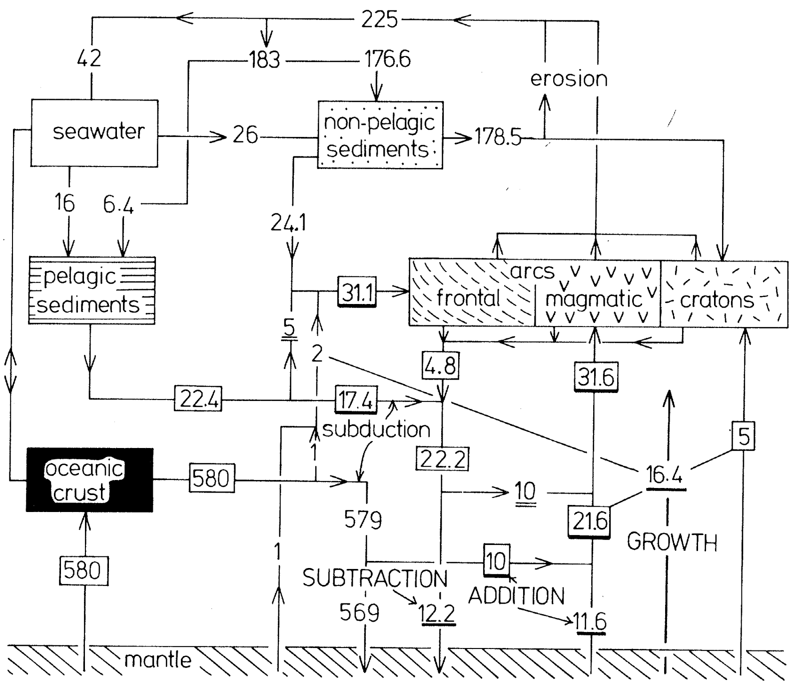

FIGURE 4. Schematic illustration of annual mass flux among the mantle, crust and sea water. Flux values are in units of 10^{14} g/a. Sedimentary redistribution values are from Garrels & McKenzie (1971); other values are discussed in the text.

times and additions to Andean-type volcanic arcs are difficult to estimate and have not been included. Other estimates of addition rates based upon arc volcanic production rates and estimates of volcanic/plutonic ratios are of the same order although somewhat less: 0.3–1.5 Pg/a (Fyfe 1978), 1.38 Pg/a (Brown 1977), and 2.76 Pg/a (Francis & Rundle 1976). We therefore take 3.16 Pg/a as the average Phanerozoic arc accretion rate and 6.27 Pg/a as the average total accretion rate (figure 4), recognizing that the arc rate, in particular, is a minmum.

To determine whether growth, diminution, or steady-state constant volume has occurred, a knowledge of subtraction rate is required. Although redistibution and shallow recycling have clearly occurred (Armstrong 1977; Kay *et al.* 1978) by decretion, delamination and sediment subduction, it is less obvious from isotopic arguments (O'Nions & Pankhurst 1978) to what extent continental mass is subtracted to be homogenized and mixed with the mantle in its convective system. Any material, whether subtracted continental or mantle material, cannot retain a long-term coherent identity below the lithosphere except as multiply interleaved and refolded laminae due to the large finite strains resulting from mantle convection (McKenzie 1979).

The value of 0.48 Pg/a (figure 4) is a maximum for the subduction based upon the decretion of $8 \times 10^5/\mathrm{km}^2$ from western South America since early Jurassic times. The dominant contributor to subducted continental mass, however, must be oceanic pelagic sediment. The oceanic crust is created at a rate of about 58 Pg/a and subducted at about the same rate less a minor fraction of ophiolite fragments clipped off in subduction–accretion prisms. This oceanic crust carries a layer of pelagic sediment with an average thickness of 0.4 km to the subduction zones at a rate of about 2.24 Pg/a (figure 4). To judge by the constitution of dissected subduction–accretion wedges such as the Franciscan assemblage of California and the Southern Uplands of Scotland (Leggett *et al.* 1979), coarse clastics in the form of trench turbidites and upper slope basin sediments dominate a minor pelagic contribution in a ratio of about 4:1. Thus, of the 3.11 Pg/a, we judge that only about 0.5 Pg/a is accreted pelagic sediment and therefore that 1.74 Pg/a of the pelagic sediment is subducted to be added to the base of the subduction–accretion prism, shallow-recycled in arc magmas or subtracted. Subduction of the bulk of the pelagics is probably aided by entrapment in the topographic irregularities at the top of the oceanic basalts.

Ophiolite basements appear to comprise about one-third of the crustal section of west Pacific and Caribbean arcs and therefore we take a contribution of 1 Pg/a from the oceanic crust to the addition rate. Of the remaining 2.16 Pg/a, it is inconceivable that more than about half comes from shallow-recycled continental mass, to judge from the high proportion of mafic and intermediate igneous rocks in extensional arcs, and most workers would regard our value of 1 Pg/a as much too high. This leaves 1.16 Pg/a coming by partial melting of mantle ultramafics. Further contributions to addition come from a minor (about 0.2 Pg/a) component of ophiolites incorporated into subduction–accretion prisms and a more significant mass of about 0.5 Pg/a of basalts underplated beneath rift complexes. We therefore derive a Mesozoic/Cainozoic addition rate of 2.86 Pg/a, a subtraction rate of 1.22 Pg/a and a growth rate of 1.64 Pg/a ($0.6\ \mathrm{km}^3/\mathrm{a}$). The subtraction rate may be as high as 2.2 Pg/a, but this does not affect the growth rate because additional material from the mantle and the oceanic lithosphere is needed to make up the arc accretion rate of 3.16 Pg/a. Although these values and rates are subject to considerable uncertainties we can see no obvious way of avoiding a Mesozoic/Cainozoic growth rate of a minimum 16.4 kg/a.

Continental freeboard has been used as an argument against post-Archaean growth (Wise 1974). Continental average thickness has been about the same since Archaean times (Burke & Kidd 1978; Windley 1977) because minimum-melting granites are rare in Archaean terrains and shallow-water sedimentary sequences of early Proterozoic through Phanerozoic age are preserved as thin sheets on the continents. Consequently, it is argued that freeboard has been roughly constant on a long-term average basis and, therefore, that, given a constant volume hydrosphere, contraction of the area of ocean basins by the accretionary growth of continents would necessitate a progressive flooding of the continents with a progressive secular evolution towards deeper water epicontinental sediments. Apart from glacial control, there are three important ways in which changes in sea level and therefore freeboard can occur on at least a short-term basis (figure 5). Accretion (figure 5, B) or stretching of 1% causes a sealevel rise of a little over 20 m. Similarly, a 20 m drop in sea level results from a 1% decretion (figure 5, C) or shortening. Valentine & Moores (1970) have pointed out the close connection between times of major continental fragmentation and transgression and between times of major collisional orogeny and regression. Late Silurian to Devonian collision in the Caledonides and northern

Appalachians was accompanied and followed by fluviatile red-bed deposition and the assembly of Pangaea in the late Carboniferous and early Permian was accompanied and followed by a major regression.

Substantial changes in sea level, however, are accomplished by changes in the rate of relative plate motion. Increased plate accretion rates lead to larger ridge volumes and therefore sea-level rise (Pitman 1978); conversely, decreased rates of plate accretion result in a drop in sea level. Some forms of tectonic episodicity such as blueschist metamorphism and ophiolite obduction may be a consequence of increased rate of relative plate motion. Sealevel drops (figure 5, C, E, G) are buffered by the resulting increased continental denudation and oceanic sedimentation, which gradually re-establishes average freeboard (figure 5, H). Sealevel rises (figure 5, B, D, F), however, are unbuffered.

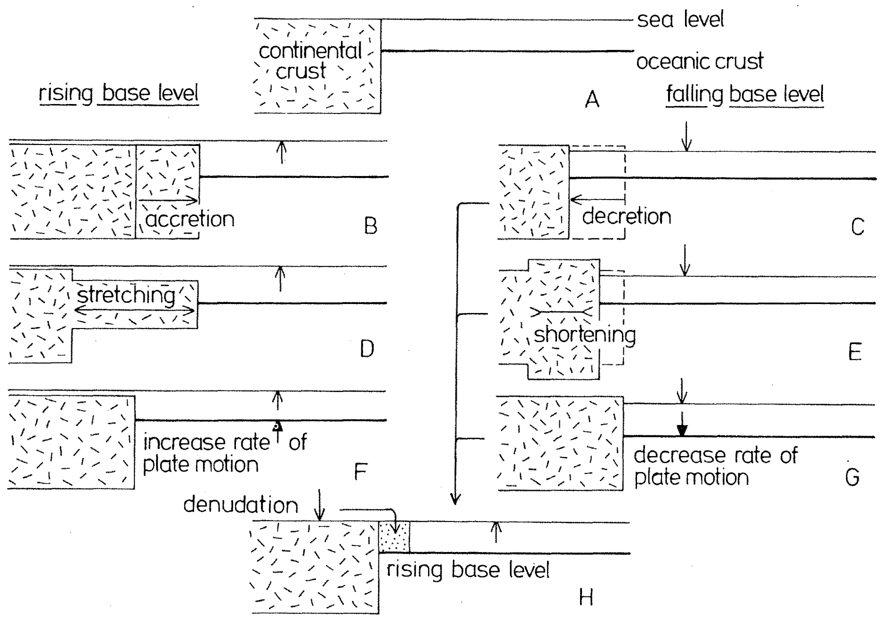

FIGURE 5. Non-glacial mechanisms by which sea level is raised or lowered; description and discussion in text.

A 15% growth of continental mass during the last 700 Ma would produce a sealevel rise of about 320 m, which could be balanced by a decline of 50% in the average rate of plate accretion. A 50% decline extrapolated linearly implies a minimum late Archaean rate of relative plate motion some six times as fast as the present rate, a conclusion similar to that reached by Bickle (1978) by totally different reasoning. We believe that the freeboard argument demands continental growth given a decline in rates of relative plate motion as radiogenic heat production diminished.

4. Secular changes

Knowledge of the first 800 Ma of earth history is hampered by lack of a geological record. Nevertheless, Smith (this symposium) uses constraints from the Moon's crust and the surviving physicochemical properties of the Earth's crust to derive a model for early crust–mantle evolution. He places emphasis on heavy meteorite bombardment up to *ca*. 3900 Ma, and a multi-level rapid convective mechanism that promoted upward transfer of basaltic material and downward transfer of volatile-rich material during a continuous melting–growth process.

Relative motion among a mosaic of plates is an essential consequence of the mantle convection system by which heat is lost and mass is added to and subtracted from the crust. Boundary plate motion rates therefore must reflect the rate of radiogenic heat production and loss, and must have been at least six times faster (Bickle 1978) during the interval of fast Archaean crustal growth from about 3600 to 2500 Ma B.P. (Burke & Kidd 1978). The rapid rate of Archaean plate motion was probably controlled by small-scale convective overturns (McKenzie & Weiss 1975); oceans were much shallower because their average age was less than 10 Ma compared with about 50 Ma at present. Archaean tectonic styles indicate clearly that large cratons did not exist and that the rapidly growing and evolving crust had a permobile character (Burke *et al.* 1976). We visualize an Archaean tectonic scheme of rapid motion among a series of much smaller mainly oceanic plates with average oceanic depths of less than 3 km and numerous sea-mount and oceanic-plateau complexes. The resulting fast-growing volcanic arc systems amalgamated rapidly into nuclei that became the cores of late Archaean continents.

A root zone or tectosphere 200 km thick (Jordan, this symposium) acted as a thermal buffer between the Archaean continental crust and the hot underlying convecting mantle (Davies 1979). Thus thermal gradients in the lower continental crust did not exceed 20–40 °C/km (Tarney & Windley 1977) and, because moderately high-pressure mineral assemblages are at the present surface of the continents, those parts of the Archaean sialic crust must have been at least 60–80 km thick by 2800 Ma ago (Wells, this symposium). Isotopic data indicate that such crustal thickening took place by short-lived episodes (100–200 Ma maximum) of irreversible chemical differentiation of the mantle (Moorbath 1978).

Such crustal thickening took place either in the granulite–gneiss belts (Windley 1977), which consist predominantly of tonalitic, granodioritic and trondhjemitic orthogneisses in granulite or in high amphibolite facies terrains such as west Greenland (McGregor 1973). Wells (this symposium) suggests that this massive crustal thickening and accretion took place by recumbent magma sheet intrusion (associated with thrusting), which helped to retard conductive heat supplies from the upper mantle. The tonalite material was probably produced by partial melting of metabasalt (Barker *et al.*, this symposium) or quartz–eclogite (Arth & Hanson 1972), most likely produced by rapid subduction.

In the modern Earth, we see comparable crustal thickening and structural and magmatic processes taking place in active Cordilleran continental margins (Tarney 1976; Windley & Smith 1976; Wells, this symposium; Brown, this symposium; Hamilton, this symposium). But we must go to the deep-level sections of such Cordilleran orogenic belts to observe comparable geological relationships, such as the British Columbia batholith, which has been eroded down to 35 km depth (Hollister 1979). Tarney and B. F. W. have recently discovered, in southern Chile, that, in a catazonal level of the Andes (sillimanite and kyanite are stable in schists in the batholiths), an early subhorizontal tectonic régime of accretionary thrust stacks with schists and gneisses was followed by sheet intrusion of tonalite commonly deformed to orthogneiss, little different in principle from the Archaean geological relationships established by McGregor (1973) in west Greenland.

The second dominant type of Archaean terrain comprises greenstone–granodiorite belts that consist of a lower unit of oceanic-type ultramafic and mafic lavas, a central unit of island-arc-type calc-alkaline andesites and dacites, and an upper sequence of volcanogenic greywackes, clastic conglomerates and quartzites and chemically precipitated iron formations and chert. Although many models have been proposed to account for greenstone belt formation (Windley 1977), we consider the most viable to be based on comparison with modern intra-, inter- and rear-arc basins (Burke *et al.* 1976; Tarney *et al.* 1976; Tarney & Windley, this volume). The major zones of crustal growth in modern orogenic belts of subduction-related arc systems (the Himalayan belt contains negligible evidence of tonalitic-granite production during continent–continent collision) are little different from the major tectonic régimes of the Archaean, and the magma products of the Archaean were broadly similar to those of the Phanerozoic (Tarney 1976). By the end of the Archaean, almost 85% of the present volume of the continent had developed, at an average growth rate of 11.7 Pg/a.

The Archaean–Proterozoic boundary, about 2500 Ma ago, was the most important threshold in geological history (Salop & Scheinmann 1969). It represents the time of change from the Archaean low-torsional rigidity plate régime, based on small-cell convection, to a régime of larger, more torsionally rigid, plates. The massive thickening of continental crust towards the end of the Archaean, between about 3000 and 2600 Ma ago (Taylor, this symposium), gave rise to rigid, thick and stable continental portions of plates underlain by tectospheric roots. From this time, it was possible for modern-style plate tectonics to operate. This view is based upon the appearance in the early Proterozoic, between about 2500 and 1800 Ma B.P. of the following features, which we believe to be indicators that plate tectonics was affecting a lithosphere consisting of oceans and continents little different in size from those of today:

(i) large continental areas showing a high degree of torsional cohesion evidenced by the deposition of little-deformed cratonic sequences and widespread dyke swarms with fairly uniform orientations;

(ii) continental shelf – rise sequences on and aulacogen re-entrants in continental margins, often with early alkali complexes associated with the rifted margin (Hoffman 1973);

(iii) linear/arcuate orogenic belts with thrust sheets over little reactivated basement with continental margin sequences (Hoffman 1973);

(iv) widespread basement reactivation adjacent to linear/arcuate thrust belts; reactivation involves the generation of high-potash minimum-melting granites and major shear zone tectonics on intracontinental transforms similar to the present tectonic behaviour of the western part of the Peoples Republic of China (Dewey & Burke 1973; Molnar & Tapponnier, 1975). Much of the Hudsonian terrain of Canada north of the Thompson Nickel Belt of Manitoba and east of the Slave Province is of this origin (Gibb 1978). Deep sections of such shear zones are prominent in west Greenland (Watterson 1978).

However, although a clear case can be made for the operation of the Wilson cycle of continental rupture, oceanic growth and continental collision in a plate tectonic framework during the Proterozoic (Hoffman 1973), an equally strong argument can be made for substantial and significant differences from the late Proterozoic/Phanerozoic (700 Ma B.P. to present) geologic effects of plate tectonics. First, although cratons were much larger and more stable than during the Archaean, extensive intracontinental rifting resulted in the widespread injection of anorthosites and rapakivi ferrogranites, probably the products of lower crustal melting induced by massive mafic magma injection in a stretched and thinned crust (Emslie 1978). Secondly, there is some geological (Kröner 1977) and palaeomagnetic (Briden 1976) evidence suggesting that many Proterozoic orogenic belts, particularly those of Africa, were the result of minor separation and resuturing. We believe that many Proterozoic orogenic belts were of classic Alpine type,

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whose history was that of stretching and extensive thinning of continental crust, with minor separation to form narrow oceanic lenses, followed by shortening and thrust restacking of the thinned crust. Thirdly, the rarity of granodiorite/greenstone belts of Archaean type or mafic/intermediate extensional arcs of west Pacific type, in regions evolved from 2500–700 Ma B.P., is striking, as is the total absence of Phanerozoic-type ophiolite complexes and blueschists. Tarney & Windley (this symposium) cite some possible Proterozoic greenstone belts, but in total volume they are very small. This suggests to us a dominance of Andean arc and collisional tectonics over extensional arc tectonics and may indicate substantial Proterozoic continental margin compression and, possibly, decretion (Dewey 1980). Engel *et al.* (1964) have drawn attention to the rapid increase in magmatic potash/soda ratio at about 2500 Ma B.P. and an even more dramatic increase in sediments. Schwab (1978) has similarly demonstrated a rapid decrease in potash, silica and lime, with a concurrent decrease in soda, alumina and magnesia. We take this, as did Engel *et al.* (1979), to reflect the widespread establishment, by 2500 Ma B.P., of continental crust, which underwent rapid differentiation. The dominance of high-potash minimum-melting granites over calc-alkaline silicic/intermediate igneous rocks from 2500 to 700 Ma B.P. implies a dominance of crustal differentiation over crustal growth, perhaps with decretion/subtraction balancing or exceeding accretion/addition.

5. Conclusions

Plate tectonics is the shallow boundary conduction layer expression of heat loss by mantle convection and the fractionation of heat-producing elements into the buoyant, weak continental crust is achieved by multi-stage processes inherent in plate tectonics. Late in the Archaean, total radiogenic heat production was at least twice as great as it is now; from that time it has exponentially declined to its present level (Lubimova 1969). Higher conductive heat loss with higher Archaean geothermal gradients cannot account for the higher heat flux (Burke & Kidd 1978) and, in consequence, convection overturn and the resultant plate motion must have been faster. We, like Bickle (1978), argue for plate motion rates, at the end of the Archaean, that were more than six times greater than at present.

The principal feature of this decline in heat production and consequent plate motion rate has been, we believe, a secular change from an Archaean tectonic régime of fast-moving, numerous, small, thin, torsionally weak plates with an average age of less than 10 Ma and much intraplate volcanism to the present regime of slower-moving, fewer, larger, thicker, torsionally stronger plates with an average age of about 50 Ma and much less intraplate volcanism. Increasing average age of the ocean floor resulted in steadily increasing depth of the ocean basins and consequently a falling sea level, countered, however, by intermittently rising sea level caused by episodic continental growth. A late Archaean plate motion rate six times the present rate produces a sea level about 1 km higher than at present, if one assumes a constant-volume hydrosphere; 15% growth since 2500 Ma B.P. has caused an average sea level rise of 320 m. Therefore, late Archaean sea level was probably some 700 m higher than that at present. During the Proterozoic, average sea level steadily fell to the present average freeboard, which was established by about 700 Ma.

Perturbations in the declining rate of plate motion have occured, however, and are particularly well documented for the Cretaceous (Pitman 1978) when increased rates were accompanied by rising sea level, much ophiolite obduction and an increased production of granodiorite and blueschist. Plate mosaic geometry and continental distribution may be important in determining these departures from steady change. During the Cretaceous, lengths of Pacific subduction systems were greater by at least 50% (Briden *et al.* 1974, fig. 6), with more cold lithosphere entering the asthenosphere, speeding convection and therefore plate motion. Fast subduction resulted in sediments and slices of continental crust dragged deep into subduction zones to enjoy blueschist facies metamorphism and in increased calc-alkaline magma production. The Cretaceous was a period of great rifting and ocean opening, the resultant stretching of continental margins enhancing the sealevel rise. Ophiolite obduction may have resulted from a lithosphere that was hotter, thinner and weaker by virtue of its lower average age. Higher sea level caused lower erosion rates and, therefore, fewer clastics and subduction–accretion prisms, a higher proportion of shallow-water limestones and anoxic deep-ocean conditions. The Ordovician was a period of high sea level and, like the Cretaceous, was one of ophiolite obduction and continental dispersal (Scotese *et al.* 1979, fig. 9). In contrast, the assembly of Pangea-like supercontinents, as during the late Carboniferous and early Permian, drastically reduced the length of subduction zones by blocking them with continental crust, thereby suddenly reducing the volume of subducting lithosphere. Reduction in plate motion rate causes a sealevel drop that is enhanced by a drop caused by continental shortening. Such periods are characterized also by widespread red-bed sequences and high erosion rates with rapid redistribution of high-potash detritus from the radiogenic upper crust resulting from collisional differentiation.

The high Archaean average crustal growth rate of 11.17 Pg/a resulted from a high calc-alkaline magma generation rate, itself a consequence of greater plate motion rate. The Archaean was dominated by ensimatic arc and back-arc tectonics (Taylor 1979; Tarney & Windley, this symposium) with a rapid amalgamation of those arcs and microcontinents to form about 85% of the present crustal mass by 2500 Ma. Although rapid addition occurred during the Archaean, crustal thicknesses were moderate until late in that period. Consequently, little crustal differentiation, uplift, erosion and redistribution occurred until late Archaean times, with the result that coarse sediments and subduction–accretion prisms are apparently absent during much of the Archaean.

During the Proterozoic, the cratons amalgamated by the consolidation of Archaean arcs, increased in internal stability and were subject to rupture, separation and recombination (Hoffman 1973), although a dominant style of orogenesis was a lacing network of remobilization zones resulting from intracontinental transforms produced by collisional reactivation, and possibly the collapse of para-oceans formed by continental attenuation.

The Proterozoic was a period of slow growth and dominance of crustal differentiation with high potash/soda ratios in igneous and sedimentary rocks and europium depletion in the upper crust (Taylor 1979). By about 700 Ma B.P., the geologic expression of tectonic regime seems little different from the Mesozoic/Cainozoic with the renewal of growth by ensimatic arc addition, the generation and obduction of ophiolites with sheeted complexes and, by about 600 Ma B.P. the first clearly subduction-related blueschists (Wood 1974).

The changing tectonic styles of the Archaean/Proterozoic are consistent with the secular evolution of a slowing plate tectonic régime; that is, the same basic cause produces an evolutionary sequence of different effects. We believe that it is more fruitful to work back from a solidly based Mesozoic/Cainozoic plate tectonic framework to discover different Proterozoic and Archaean effects than to use the differences to erect untestable *ad hoc* hypotheses.

We thank our many friends and colleagues with whom we have argued profitably for many years. The values for Archaean and post-700 Ma B.P. continental mass are based upon a long-term continuing cooperative study, at Albany, of continental tectonics.

References (Dewey & Windley)

Armstrong, R. L. 1968 *Rev. Geophys.* **6**, 175–199.
Armstrong, R. L. 1971 *Earth planet. Sci. Lett.* **12**, 137–142.
Arth, J. G. & Hanson, G. N. 1972 *Contr. Miner. Petr.* **37**, 161-174.
Barazangi, M. & Isacks, B. L. 1976 *Geology* **4**, 686–692.
Bickle, M. J. 1978 *Earth planet. Sci. Lett.* **40**, 301–315.
Bird, P. 1978 *J. geophys. Res.* **83**, 4975–4987.
Bird, P. 1980 *J. geophys. Res.* **84**, 7561–7571.
Briden, J. C. 1976 *Phil. Trans. R. Soc. Lond.* A **280**, 405–416.
Briden, J. C., Drewry, G. E. & Smith, A. G. 1974 *J. Geol.* **82**, 555–574.
Brooks, C., James, D. E. & Hart, S. R. 1976 *Science, N.Y.* **193**, 1086–1094.
Brown, G. C. 1977 *Nature, Lond.* **265**, 21–24.
Burke, K., Dewey, J. F. & Kidd, W. S. F. 1976 In *The early history of the Earth* (ed. B. F. Windley), pp. 113-129. London: Wiley.
Burke, K. & Kidd, W. S. F. 1978 *Nature, Lond.* **272**, 240–241.
Chase, C. G. 1972 *Geophys. Jl R. astr. Soc.* **29**, 117–122.
Crawford, A. J. & Keays, R. R. 1978 *Earth planet. Sci. Lett.* **41**, 197–208.
Dal Piaz, G. V., Hunziker, J. C. & Martinotti, G. 1972 *Mem. Soc. geol. ital.* **11**, 433–466.
Dalziel, I. W. D., DeWit, M. J. & Palmer, K. F. 1974 *Nature, Lond.* **250**, 291–294.
Davies, G. F. 1979 *Earth planet. Sci. Lett.* **44**, 231–238.
Dewey, J. F. 1981 *Proc. geol. Ass. Can.* (In the press.)
Dewey, J. F. & Burke, K. 1973 *J. Geol.* **81**, 683–692.
Drury, S. A. 1978 In *Archaean geochemistry* (ed. B. F. Windley & S. M. Naqvi), pp. 3–23. Amsterdam: Elsevier.
Emslie, R. F. 1978 *Precambr. Res.* **7**, 61–98.
Engel, A. E. J., Itson, S. P., Engel, C. G., Stickney, D. M. & Cray, E. J. 1974 *Bull. geol. Soc. Am.* **85**, 843–858.
Francis, R. & Rundle, C. 1976 *Bull. geol. Soc. Am.* **87**, 474–480.
Fyfe, W. S. 1978 *Chem. Geol.* **23**, 89–114.
Garrels, R. & McKenzie, F. 1971 *Evolution of the sedimentary rocks.* New York: Norton.
Gass, I. G. 1979 In *Evolution and mineralization of the Arabian–Nubian Shield* (ed. A. M. S. Al-Shanti), pp. 11–20. Oxford: Pergamon.
Gaudette, H. E. & Hurley, P. M. 1979 *Tectonophysics* **54**, 211–230.
Gee, D. G. 1975 *Am. J. Sci.* A **275**, 468–515.
Gibb, R. A. 1978 *Nature, Lond.* **278**, 50–52.
Hargreaves, R. B. 1976 *Science, N.Y.* **193**, 363–371.
Hargreaves, R. B. 1978 *Nature, Lond.* **276**, 459–461.
Hoffman, P. 1973 *Phil. Trans. R. Soc. Lond.* A **273**, 547–581.
Hollister, L. S. 1970 *Episodes I.U.G.S.*, no. 3, pp. 3–8.
Hudson, T., Plafker, G. & Peterman, Z. E. 1979 *Geology* **7**, 573–577.
Hurley, P. M., Hughes, H., Faure, G., Fairbairn, H. W. & Pinson, W. H. 1962 *J. geophys. Res.* **67**, 5315–5334.
Hurley, P. M. & Rand, J. R. 1969 *Science, N.Y.* **164**, 1229–1242.
Jordan, T. H. 1978 *Nature, Lond.* **274**, 544–548.
Karig, D. E., Cardwell, R. K., Moore, G. F. & Moore, G. D. 1978 *Bull. geol. Soc. Am.* **89**, 265–276.
Kay, R., Sun, S. S. & Lee, H. 1978 *Geochim. cosmochim. Acta* **42**, 263–273.
Kornprobst, J. 1976 *Bull. geol. Soc. Fr.* **18**, 607–618.
Kröner, A. 1977 *Tectonophysics* **40**, 101–135.
Leggett, J. K., McKerrow, W. S., Morris, J. H., Oliver, G. J. H. & Phillips, W. E. A. 1979 In *The Caledonides of the British Isles – reviewed* (ed. A. L. Harris, C. H. Holland & B. E. Leake), pp. 489–498. London: The Geological Society.
Lowman, P. D. 1976 *J. Geol.* **84**, 1–26.
Lubimova, E. A. 1969 *Am. geophys. Un. Mem.* **13**, 63–77.
McCulloch, M. T. & Wasserburg, G. J. 1978 *Science, N.Y.* **200**, 1003–1011.
McGregor, V. R. 1973 *Phil. Trans. R. Soc. Lond.* A **273**, 343–358.
McKenzie, D. P. 1969 *Geophys. Jl R. astr. Soc.* **18**, 1–32.
McKenzie, D. P. 1978 *Earth planet. Sci. Lett.* **40**, 25–32.
McKenzie, D. P. 1979 *Geophys. Jl R. astr. Soc.* **58**, 689–715.

McKenzie, D. P. & Weiss, N. 1975 *Geophys. Jl R. astr. Soc.* **42**, 131–174.
Milnes, A. G. 1978 *Tectonophysics* **47**, 369–392.
Minster, J. B., Jordan, T. H., Molnov, P. & Haines, E. 1974 *Geophys. Jl R. astr. Soc.* **36**, 541–576.
Molnar, P. & Gray, D. 1979 *Geology* **7**, 58–62.
Molnar, P. & Tapponnier, P. 1975 *Science, N.Y.* **189**, 419–426.
Moorbath, S. 1978 *Phil. Trans. R. Soc. Lond.* A **288**, 401–413.
Moorbath, S. & Taylor, P. N. 1981 In *Precambrian plate tectonics* (ed. A. Kröner). Amsterdam: Elsevier. (In the press.)
O'Nions, R. K. Pankhurst, R. J. 1978 *Earth planet. Sci. Lett.* **38**, 211–236.
Pitman, W. C. 1978 *Bull. geol. Soc. Am.* **89**, 1389–1403.
Ringwood, A. E. 1974 *J. geol. Soc. Lond.* **130**, 183–204.
Rogers, J. J. W. & Novitsky-Evans, J. M. 1977 *Geophys. Res. Lett.* **4**, 347–350.
Salop, L. & Scheinmann, Y. M. 1969 *Tectonophysics* **7**, 565–597.
Schwab, F. L. 1978 *Geology* **6**, 532–536.
Scotese, C. R., Bambach, R. K., Barton, C., Van der Voo, R. & Ziegler, A. M. 1979 *J. Geol.* **87**, 217–277.
Shaw, D. M. 1976 In *The early history of the Earth* (ed. B. F. Windley), pp. 33–53. London: Wiley.
Smith, A. G. 1976 *Tectonophysics* **33**, 215–285.
Sutton, J. 1963 *Nature, Lond.* **198**, 731–733.
Tarney, J. 1976 In *The early history of the Earth* (ed. B. F. Windley), pp. 405–417. London: Wiley.
Tarney, J., Dalziel, I. W. D. & De Wit, M. J. 1976 In *The early history of the Earth* (ed. B. F. Windley), pp. 131–146. London: Wiley.
Taylor, S. R. 1979 In *The Earth* (ed. M. W. McElhinny), pp. 353–376. London: Academic Press.
Thorpe, R. S. & Francis, P. W. 1979 *Tectonophysics* **57**, 53–70.
Valentine, J. W. & Moores, E. M. 1970 *Nature, Lond.* **228**, 657–659.
Veizer, J. & Jansen, S. L. 1979 *J. geol.* **87**, 341–370.
Watterson, J. 1978 *Nature, Lond.* **273**, 636–640.
Wells, P. R. A. 1980 *Earth planet. Sci. Lett.* **46**, 253–265.
Windley, B. F. 1977 *Nature, Lond.* **270**, 426–428.
Windley, B. F. 1977 *The evolving continents.* London: Wiley.
Windley, B. F. 1981 In *Plate tectonics in the Precambrian?* (ed. A. Kroner). Amsterdam: Elsevier. (In the press.)
Windley, B. F. & Smith, K. V. 1976 *Nature, Lond.* **260**, 671–675.
Wise, D. V. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 45–58. Berlin: Springer.
Wolery, T. & Sleep, N. 1976 *J. Geol.* **84**, 249–275.
Wood, D. S. 1974 *Spec. Publs Soc. econ. Palaeont. Miner., Tulsa* **19**, 334–344.
Zeil, W. 1979 *The Andes.* Berlin: Borntraeger.