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Precambrian Geologic History

Continents grew and emerged from beneath the primordial sea.

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A considerable body of opinion concerning the early history of the earth favors extensive melting and prodigious degassing, either associated with the initial accretion of the earth, or occurring soon after its formation (*1–3*). Such a molten earth hypothesis would suggest, as Hess (*4*) stated, "the initial formation of a thin continental or sialic layer uniformly over the earth, with a very thin uniform world encircling water layer above." Hess, however, postulated the development of the present bilateral crustal asymmetry very early in earth history, and indeed many geologists seem to assume that continents and oceans, more or less in their present form, have existed for most of the geologic past. This strictly uniformitarian view is supported by the presence of sedimentary sequences (including detrital components) overlying deeply eroded metamorphic rocks in the oldest terrains on the earth (*5*). Occurrences such as these throughout the stratigraphic record testify that the classic "rock cycle" of uplift, erosion, sedimentation, burial, and metamorphism has been going on throughout geologic time. This evidence does not say, however, that continents and oceans have always been present, nor does it imply that the effects of erosion and deposition were always as widespread as they are now.

The thesis of this article is the contrary view: that the separation of the primordial shells into continents and oceans began only about 3.7 billion years ago and was a slow process, occupying most of Precambrian time; subaerial erosion and the deposition of its products have increased as more and more land has emerged. This idea occurred to me originally when I was considering Precambrian paleomagnetic results, and I will discuss these data and their apparent implications first. The early-earth crustal model of Fyfe (*1, 6, 7*), which was founded on thermal considerations stemming from the higher radiogenic heat production in the Precambrian (*8, 9*), also seems to favor globe-encircling sialic and hydrospheric shells. This view sets the stage for an isostatic mechanism for the formation of Archean greenstone belts, and leads to a general hypothesis for the stabilization of cratons and the progressive generation of thicker continental crust segments and deeper ocean basins. I will appraise this hypothesis, and the timing of the events involved, in light of some of the major aspects of the Precambrian sedimentary rock spectrum. Finally, I will attempt to assess the implications of the model for the overall Precambrian geologic record.

Precambrian Paleomagnetic Data

Briden (*10*) has pointed out that the Precambrian paleomagnetic data available today are comparable in quantity and quality to the data available for the Phanerozoic in 1960. The paleomagnetic polar wander curves for the Phanerozoic of Europe and North America available in 1957 (*11*), after only about 7 years of data accumulation, are compared in Fig. 1 with an equivalent synthesis of paleomagnetic data in 1973 (*12*). The curves are negligibly different, even though considerably more data were included in the more recent estimate. Briden's argument is that the apparent implications of Precambrian paleomagnetic data should not be dismissed. These are outlined below.

McElhinny *et al.* (*13*) evaluated late Precambrian to early Paleozoic paleomagnetic data from the southern continents. They found the data to be consistent with a common polar wander path when the various continents were reassembled into Gondwana. The distribution of glacial tillites formed during that interval is consistent with this polar wander curve (Fig. 2). They consider the data to indicate that the Gondwana supercontinent dated from at least 800 million years ago, and that orogenic or mobile belts [such as the Pan African orogeny of 450 to 680 million years ago (*14*)] by some means developed within the supercontinent.

Briden (*15*) and Piper *et al.* (*16*) found that paleomagnetic data from South America, West Africa, and southern Africa, for the period 2300 to 1950 million years ago are also consistent with a common polar wander curve (Fig. 3). This implies that this major segment of Gondwana was essentially a single entity from at least the beginning of the Proterozoic until its breakup in Triassic to Jurassic time.

Although the validity in detail of the Precambrian paleomagnetic data can be questioned (*17*), there is a trend in the data which supports two conclusions.

1) At least some of the Precambrian orogenic belts presently located within continental cratons were actually formed there. [This is also the conclusion of many geologists who have worked in these terrains (*18*).] The orogenic belts are not the result of collisions of subcontinental blocks by the closure of major oceans (although closure of small narrow oceans is a possible factor).

2) The Precambrian cratons may now be much more fragmented and dispersed than they were in Precambrian time (*19*).

Presumably, forces similar to those which today drive lithospheric plates

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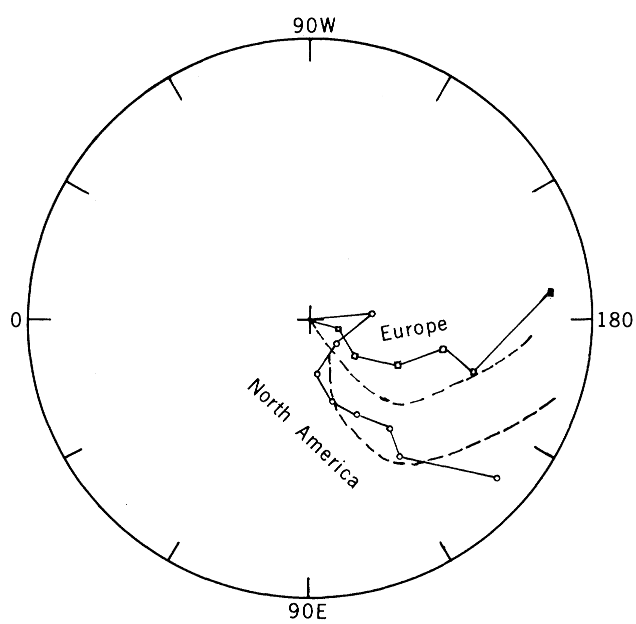
were also operating in Precambrian time. This conclusion is favored by evidence for compression revealed in the mobile belts within continents (*20*). Yet, if the relative areas of continents and oceans were the same then as now, why did complementary tensional stresses not completely fragment and disperse the supercontinent? A possible explanation for this relative cratonic coherence is related to the fact that sialic crust, being less dense, cannot be readily subducted. If the area of continental crust was much larger and the area of oceans much smaller than at present, there would have been no room on the globe for continents to drift around in (*21*).

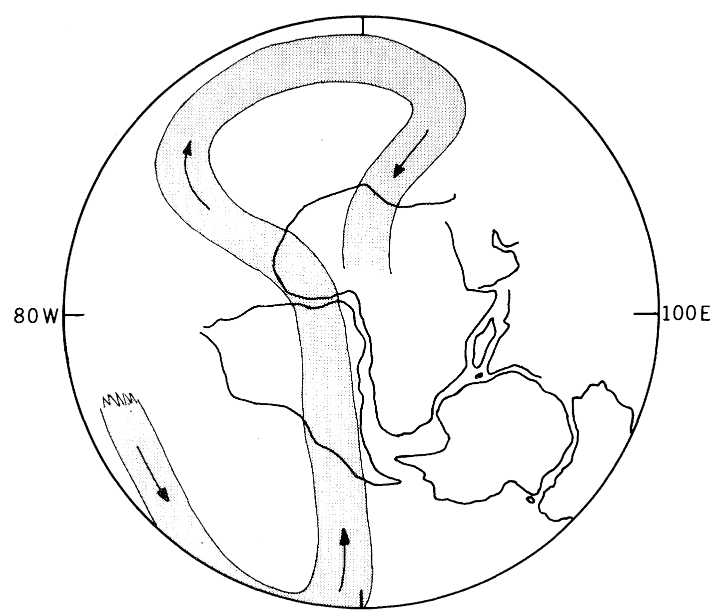
If we assume isostasy and a constant volumetric proportion of continental crust to hydrosphere in Precambrian time, then in order to have continents cover a larger area of the globe, it is necessary to postulate that some of the hydrosphere flooded the continents. This points to the ancient concept (*22*) of a primitive earth with a globe-encircling sialic crust covered by a primordial sea.

Primordial Earth Model

The models of thermal evolution of the earth derived by McKenzie and Weiss (*8*), assuming an initial temperature sufficient to ensure mantle-wide convection, show a smooth variation with time, depending on the initial temperature after formation and subsequent radioactive decay (Fig. 4). Their tectonic evolution model depends on their postulate of two convective flow regimes: small-scale flow throughout the upper mantle above the phase transition from the spinel to the postspinel form of olivine, and large-scale flow throughout the deep mantle. McKenzie and Weiss attribute the Archean tectonic style primarily to control by the small-scale flow, whereas contemporary large-plate tectonics is associated mainly with the large-scale flow. The interplay of these controls on evolving tectonic style is, in these authors' view, a function of the thermal history of the earth. This evolution should also be manifest in the products of more uniquely crustal, or surface, geologic processes such as igneous activity, faulting, erosion, sedimentation, and metamorphism.

Fyfe (*1, 6, 7*) pointed out that if most of the sial was segregated to the surface of the earth early in its history, when thermal gradients were particularly high,





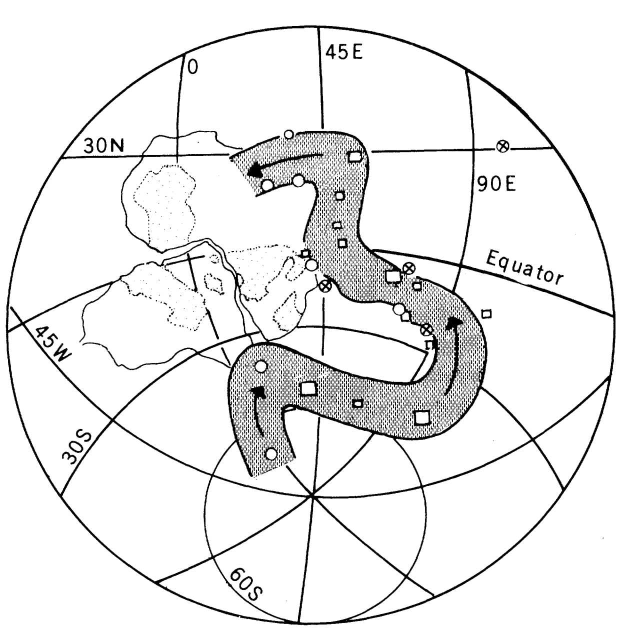


Fig. 1 (top left). Polar projection comparing the paleomagnetic polar wander curves for Europe and North America as reported by Creer *et al.* (*11*) in 1957 (dashed lines) and McElhinny (*12*) in 1973 (solid lines). Fig. 2 (top right). Polar wander path common to the reassembled Gondwana continents for the period 800 to 500 million years ago (*13*). Fig. 3 (bottom left). Polar wander path common to reassembled South America and Africa for the interval 2300 to 1950 million years ago (*15, 16*); poles are from sites in southern African cratons (open circles), the West African craton (squares), and South America (crossed circles).

then the maximum thickness of solid sialic crust would be constrained by the granite minimum-melting isotherm (see Fig. 5). Although the oldest rocks preserved on the surface of the earth appear to be granitic in composition (*5*), they are only about 3.8 billion years old, so there is no assurance that they represent the primordial crust (*23*). Differentiation after early melting of the outer part of the moon led to the formation of an anorthositic crust (*24*), and such a beginning cannot be excluded for the original crust of the earth. A molten layer undergoing differentiation on the earth, however, would have contained considerably more water and alkali metals than the molten layer on the moon. As a result, a relatively more calcalkaline fractionation trend would be expected (*25*). Being less dense, such "sialic" residual liquids would tend to rise and intrude or transform any preexisting crust, which may have formed by freezing of the primitive molten liquid or upward flotation of early formed crystals.

The continental crust today, if distributed over the entire surface of the earth, would form a shell about 12.0 kilometers thick. The present oceans are equivalent to a hydrosphere 2.8 km deep. If segregation of a substantial fraction of this present sial occurred during the early part of earth history, then the outermost crust could not have become completely consolidated at its base until the earth cooled to the granite solidus. In the presence of even a small amount of water, this temperature is not higher than about 750°C, and hence a geothermal gradient of 65°C per kilometer or less would have been required. According to the thermal models depicted in Fig. 4, this could only have occurred between about 0.3 billion years [the so-called Wasserburg model (*9*)] and 1.5 billion years [the chondritic model (*2*)] after the formation of the earth. Until then, as suggested by Bridgwater and Fyfe (*7*), there would have been a molten or partially molten layer, of gradually diminishing thickness, beneath a thin crust. With the tectonic turmoil resulting from vigorous mantle convection beneath, the thin lithospheric crust would bend and break. Folds and slabs could emerge from the sea and be eroded, and the products subside and be engulfed, remelted, and reequilibrated with the "mantle" beneath. Eventually, the sialic crust would be completely welded to the mantle beneath, and this tempestuous tectonic style would be calmed. The ancient gneiss terrain with associated supracrustals might be the surviving relics of this phase of earth history (*7*).

Archean Crust and the Origin of Greenstone Belts

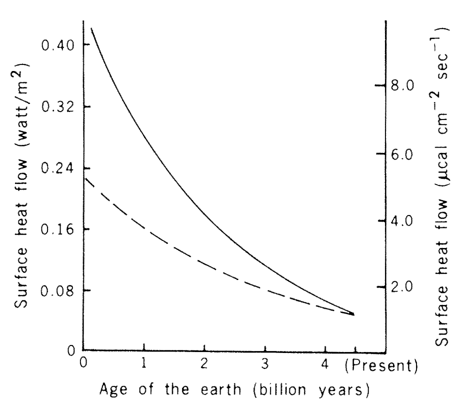
Much consideration has been given to the origin of the Archean greenstone belts and the granitic terrain between them (*7, 26*). Debate has focused on the nature of the crust on which the greenstones were erupted, and on whether the greenstones themselves or high-grade quartzofeldspathic gneisses (*27*) in the intervening granitic terrain are the most likely representatives of the primitive crust of the earth.

Windley and Bridgwater (*28*) suggested that the high-grade and low-grade Archean terrains represent different levels of erosion of the primitive crust. Anhaeusser (*29*) proposed that the terrains may be relics of Archean continental and oceanic crust, respectively. Windley and Smith (*30*) suggest that the high-grade gneiss terrains are ancient analogs of the modern deep-seated granite plutons such as those along the cordillera of western North America and South America. In the model proposed here, the high-grade gneisses are the favored candidate for oldest crust available, whatever their specific mode of origin may be.

Considering the greenstone belts alone, Anhaeusser *et al.* (*31*) favored their formation in "downwarps or fault-bounded troughs on an unstable, thin, primitive sialic crust." Windley (*32*) suggested an analogy with mid-ocean ridges. However, Goodwin (*33*), Anhaeusser (*29, 34, 35*), and others have been particularly impressed by the similarities between greenstone piles and contemporary island arcs, and favor a similar "subduction" mechanism for the formation of both. Most recently, specific plate tectonic analogies have been drawn between greenstone belts and the deposits that accumulate in marginal basins behind arcs (*30, 36*).

The model presented here invokes mainly vertical motions, akin to what Anhaeusser (*35*) has described as "gravitational slumping and downfolding of the troughs of lava and sediments on a thin unstable crust," although the globe-encircling sea is considered to play a particularly important role. The ubiquity of pillow structures in the greenstone volcanics shows that the lavas were erupted under water, but the isostatic control imposed by the depth of water on the thickness of the volcanic pile and the lithologic sequence to be expected as every greenstone pile builds toward the surface of the sea have not been emphasized.

As stated above, if the present hydrosphere and continental crust were distributed uniformly over the surface of the earth, they would form shells about 2.8 and 12.0 km thick, respectively. If the early melting, degassing, and differentiation achieved 70 percent segregation in the first billion years the idealized primitive earth 3.5 billion years ago would have an 8.4-km-thick sialic shell overlain by a 2.0-km-deep sea. (The precise percentage is not critical; only the presence of a sea is important in the following model.) As demonstrated in Fig. 6a, if basalt erupted through this sialic shell onto the floor of the sea, and isostasy was maintained, about 15.5 km of volcanics could accumulate before the pile reached sea level. Furthermore, as the volcanic pile built up, the likelihood of volcanoes breaking through the sea surface would increase. A progressive upward increase in the proportion of clastic sediment could be produced, as is



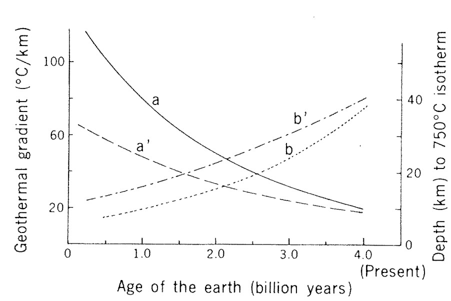


Fig. 4 (left). Surface heat flow in a model of the earth with an initial temperature sufficiently great to permit convection throughout the mantle. The solid curve is for a model with chondritic abundances of radioactive elements (*2*). The broken curve is for a model with heat flux equivalent to present values but with a K/U ratio derived from measurements of crustal rocks—the Wasserburg model (*9*). [Adapted from McKenzie and Weiss (*8*, figure 7)] Fig. 5 (right). (Curves a and a') Geothermal gradients as a function of time calculated from data in Fig. 4, assuming a solid sialic crust with a conductivity equivalent to that of average granite, 8.0×10^{-3} cal cm^{-1} sec^{-1} °C^{-1}. The effect of upward concentration of radioactive elements in the crust has been ignored. (Curves b and b') Depth to the 750°C isotherm as a function of time. Curves a and b are for the chondritic model, curves a' and b' are for the Wasserburg model.

actually observed (*31*). With sufficient isostatic downwarping (*1, 6, 7*), the sialic shell would eventually start to melt and would mix to varying degrees with the basalts. Spreading laterally, extruding, and intruding into and around the borders of the volcanic-sedimentary pile above, these anatectic granitic hybrids would cause further deformation and lead to more subsidence (Fig. 6b). With constant isostasy, the depth of the primordial sea would control the maximum height to which this pile could build (anything projecting would be rapidly eroded) and hence dictate the maximum thickness of the greenstone belt. The 15.3 km shown in Fig. 6a, even without the additional accumulation allowed by melting of the sialic layer, is consistent with the thickest sequences reported (*31*). The pile accumulated beneath a 1.0-km-deep sea, however, would be only about 7.5 km deep, which may be more consistent with the results of a recent gravity survey (*37*). On the other hand, the maximum thickness achievable might also be constrained by anatexis and erosion of the roots of the greenstone piles themselves.

Mechanism for Growth of Continents and Ocean Basins

In the small-scale convection mode of McKenzie and Weiss (*8*), there would have been areas of mantle upwelling and downwelling. Once the sialic shell became frozen to the mantle beneath, greenstone piles most likely began to form, as a result of tensional cracking of the crust above areas of upwelling. In this sense, greenstone belts are more closely analogous to mid-ocean ridges (*32*) than to island arcs. While spreading was beginning with the intrusion and extrusion of basalt and andesite in the greenstone belts, compression and thickening of the sial was taking place above the cooler downwelling areas. Thus the sial grew and thickened (Fig. 6c) and the Archean cratons were generated.

Initially, the greenstone basaltic additions to the sial increased its thickness. Eventually, however, the increasing "continental" crustal thickness permitted by the slowly decreasing geothermal gradients, coupled with the tectonic tendency to thicken continental crust (as long as it is submerged), required compensation by the formation of true oceanic crust. This may have coincided with the stage when, as the earth cooled, the lithosphere thickened and strengthened to the point where individual plates straddled the small-scale convective upwellings. Crustal tensions and compression were concentrated in longer, more widely spaced zones or belts related geometrically to the large-scale convection pattern (*8, 9*). The tectonic interaction along these zones progressively increased in scale and intensity (*14*). Plate tectonics resembling the contemporary style began; as true oceanic crust was still relatively undeveloped, however, the plates for the most part were capped by continental crust and remained submerged. Rifting (and spreading) in one place was matched by plate convergence elsewhere, and compressional mountain welts formed. These may have projected above the sea surface, been eroded, and shed thick clastic deposits. All the while, roots of these sialic mountains tended to intersect the 750°C isotherm below (Fig. 5), causing anatexis. Continued convective downwelling (subduction?) facilitated lateral spreading of these partial melt products and promoted thickening of continental crust by sialic underplating (*38*).

By some unknown means the convective pattern driving plate tectonics in the mid-Precambrian tended to maintain the coherence of the slowly dwindling fraction of the earth's surface underlain by continental crust. One or at most a few supercontinents (*10, 15, 16*) survived until late in the Precambrian, despite the increasing proportion of oceanic crust. This segregation process was continuously constrained, however, by the prevailing geothermal gradients (*1, 6, 7*). If the nominal maximum thickness of continental crust is defined as the depth to the 750°C isotherm, the thickness could only increase with time, as shown in Fig. 5. Based on complete segregation of the sial and hydrosphere before 3.5 billion years ago, with progressive separation into oceans and continents thereafter, Fig. 7 and Table 1 show the increasing fraction of the earth's surface formed by true ocean and the decreasing depth of the continental seas. In these terms, general emergence of the continental plates could not occur until about Grenville times, 3.1 billion years (Wasserburg model) to 3.6 billion years (chondritic model) after the earth formed.

Evidence for this postulated progressive emergence of land and the timing of its culmination should be present in the Precambrian sedimentary record (*39*).

Precambrian Sedimentary Record

Detrital sediments. Ronov (*40*) has estimated the proportions of different sedimentary rock types as a function of age (Fig. 8). The predominance of volcanogenic sediments and graywackes early in earth history favors the model proposed in this article, in which sedimentation initially resulted from subaqueous eruptions and the subaerial erosion of only scattered volcanic areas.

Arkoses became relatively more abundant in the mid-Proterozoic (1700 million years ago; see Fig. 8), but have decreased relative to quartz sands, toward the present. This could reflect the first

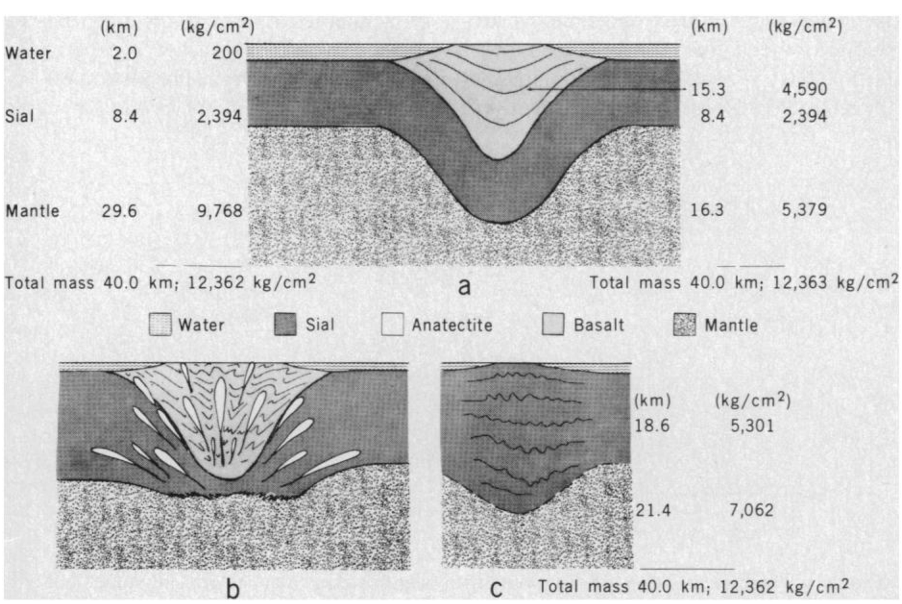


Fig. 6. Crustal models drawn approximately to scale, showing the thickness of different rock columns isostatic with a hydrosphere 2.0 km deep [density (ρ) = 1.0 g/cm^2] overlying a sial shell (ρ = 2.85 g/cm^3) 8.4 km thick on a peridotite mantle (ρ = 3.3 g/cm^3). The depth of compensation is 40 km. (a) The basaltic greenstone pile (ρ = 3.0 g/cm^3) at sea level is 15.3 km deep. (b) Postulated remelting of depressed sialic shell. (c) The sialic crust, thickened to reach sea level, is 18.4 km thick.

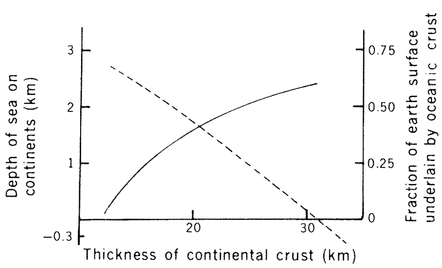


Fig. 7. (Dashed line) Depth of sea on continents as a function of thickness of continental crust isostatic with present oceanic crust (*4*). The present volumes of hydrosphere (1.37×10^{24} cm^3) and continental crust (6.3×10^{24} cm^3) are assumed. (Solid line) Fraction of the earth's surface that would be underlain by oceanic crust in such an isostatic model. These data together with the data in Fig. 5 predict the evolutionary sequence shown in Table 1.

emergence of compressional mountain welts formed of tectonized granitic terrain, as proposed in the evolutionary model. Deposition of the eroded debris on the still submerged continental platform away from the welts preserved its arkosic character. Erosion and redeposition of these arkoses, as the platforms became increasingly exposed, promoted the winnowing needed to produce quartz sands.

Banded iron formations. Banded iron formations (or jaspilites) are a characteristic, if relatively minor sedimentary rock type in association with greenstone belts in southern Africa (*31*), but occur more abundantly in the younger Canadian greenstones (*41*). They reached their maximum relative abundance as stable-platform-type chemical deposits about 2 billion years ago (Fig. 8). As stated by Bayley and James (*42*), "the overriding factor in the deposition of iron-formation almost certainly was the chemistry and biochemistry of the world's hydrosphere and atmosphere at that particular stage of earth's history."

The source of iron and silica in these sediments has long been debated, the two favored hypotheses being volcanic exhalations (*41*) and deep chemical weathering in the source area (*43*). These contrasting views derive, at least in part, from the intimate association of banded iron formations with Archean greenstone in the older deposits (*41, 44*) and with carbonates and minor clastics in early Proterozoic "continental shelf" type sedimentary accumulations (*45*).

Considering the masses of chemical precipitate involved, and the capacity of the potential source reservoirs, Borchert (*46*), Holland (*47*), and Drever (*48*) have suggested that the oceans themselves were the source of the iron and silica. I consider that all three of these views can be understood in terms of the model proposed here.

The primordial sea, activated by subaqueous volcanic exhalations since its formation, would also interact with the sea floor and any volcanogenic deposits formed on it during the early stage of crustal evolution; the water could thus dissolve considerable amounts of iron and silica (*49, 50*). This may have been the situation for more than 1 billion years before the emergence of the first substantial greenstone belts. Iron and chert precipitates would form in near-surface waters due to interaction with the atmosphere, but would redissolve in the more reducing waters at depth. Shallow stable platforms were required to accumulate and preserve the deposits, and such platforms were provided for the first time by the rising greenstone piles. As lateral segregation of sial and hydrosphere progressed, the cratons thickened still further. Eventually (~2300 million years ago) the sea above many parts of them shallowed to the point where sandbanks and shoals emerged. The development of these broad shallow platforms provided a more widespread environment where iron minerals and silica, precipitated by interaction of the sea with the increasingly oxygenated atmosphere (*51*) could accumulate and be preserved.

Never again in geologic history would circumstances be so favorable for iron formation. Sedimentary iron deposits have formed in restricted continental environments at various times since, but these younger Phanerozoic deposits typically contain a significant detrital component, and are generally considered to be distinct from classic Proterozoic banded iron formations (*52*). This difference is readily understood in terms of the proposed model, wherein large areas of land and restricted sedimentary basins only developed much later, in Phanerozoic times.

Table 1. Evolutionary sequence based on data shown in Figs. 5 and 7.

Thickness of sialic crust (km)	Fraction of earth's surface underlain by oceanic crust	Time before present (billion years)	
		Wasserburg model	Chondritic model
16	0.25	3.5	2.5
20	0.40	2.9	2.0
24	0.50	2.4	1.5
31	0.60	1.5	0.9

Evaporites. Few evaporite deposits are known in rocks much older than late Precambrian, although, as stated by MacKenzie (*49*), casts of gypsum nodules are found in rocks as old as 1.8 billion years. In view of the susceptibility of evaporites to dissolution and recycling, their restricted age distribution (Fig. 8) is no surprise. But in the model proposed there would be few environments suitable for the formation of evaporite deposits until well into the Proterozoic.

Total sediment mass versus age. Garrels and MacKenzie (*53*) estimated the total amount of sedimentary rock preserved as a function of age (Fig. 9). To explain this distribution, they considered two models: (i) constant sediment mass and (ii) linear sediment accumulation. In the first model a mass of sediment equal to that existing at present is assumed to have formed very early in earth history. In the second, the amount of sediment increased linearly as H_2O, HCl, and CO_2 were progressively degassed from the interior. For both models, the best fit

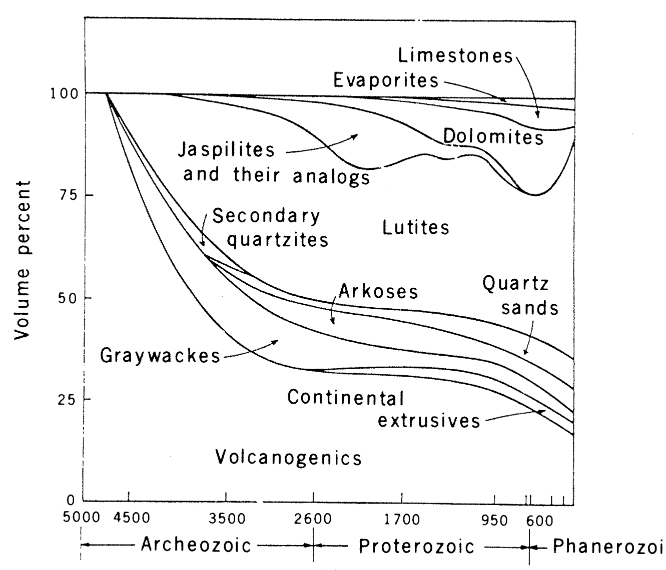


Fig. 8. Ronov's estimate (*40*) showing the relative proportion of different sedimentary rock types as a function of age. [Modified from MacKenzie (*49*)]

with the observed distribution (Fig. 9) is achieved if, throughout geologic time, the amount of sediment cycled through erosion and redeposition is equivalent to five times the present sedimentary mass. In support of these models, Garrels and MacKenzie (*53*) pointed out that if the average rate of sediment accumulation for the Tertiary (50×10^{14} grams per year) (*54*) was extrapolated for 3.5 billion years, the amount accumulated would be about five times the present sedimentary rock mass [$32,000 \times 10^{20}$ g (*53*)]. They acknowledged that most theories favor a high early degassing rate rather than a linear rate, and hence prefer their constant mass model.

An alternative explanation, implicit in the model proposed here, involves a linear rate of emergence of land, following extensive early degassing and complete crustal flooding. All the HCl and CO_2 dissolved in the hydrosphere was presumably neutralized by interaction with the sea floor, forming a deeply weathered regolith (nascent sedimentary rock). But formation of sedimentary rock proper (excluding chemical precipitates) required erosion and redeposition of this regolith; this, it is proposed, would only occur as land emerged.

This explanation yields numerical results closest to those obtained with the linear degassing model of Garrels and MacKenzie. However, it is unrealistic to picture a linear increase in sediment mass due to a linear emergence of land. If ocean basins proper grew as continental crust thickened (Fig. 7), a time would come when, relatively suddenly, the waters would drain from even the lowest, flattest parts of the continental rafts, and the area of erodable land (equivalent to total sediment mass) would increase considerably.

One could consider the distribution in Fig. 9 as a composite of the effects of two entirely different regimes. The first is a period when sediment mass increased slowly and relatively linearly until a brief interval when the continents emerged completely, the last of the paleoregolith was cycled, and the present-day sedimentary rock mass was achieved. The second is the period since then, during which the cycle of erosion, deposition, subduction, and so on has maintained roughly constant freeboard (*55*) and a more or less steady state.

If this simple two-stage history is correct, then the discontinuity in Fig. 9 that is most likely to record the changeover from linear accumulation to constant mass appears to occur at about the end of the Precambrian. Figure 10 shows a sediment mass-age distribution based on the two-stage model (*56*). The sedimentary rock erosion rate used in Fig. 10, a and b, which gives the best fit to the data in Fig. 10c is 14 percent of the accumulated sedimentary rock per 100 million years. This can be converted to an average absolute land erosion rate, for all of geologic time, as follows: During the constant-mass stage of the model the amount of sedimentary rock eroded per 100 million years is $(0.14)(32,000 \times 10^{20}$ g$) = 4,480 \times 10^{20}$ g. Correcting for the fact that only 66 percent of the present land area is underlain by sedimentary rock (*57*) gives a total eroded mass of $6,720 \times 10^{20}$ g or $2,585 \times 10^{20}$ cubic centimeters, assuming a density of 2.6 g/cm^3. The present area of land is 1.49×10^{18} cm^2, and therefore the average rate of denudation of land area throughout geologic time is $(2.585 \times 10^{15}$ cm^3/year$)/(1.49 \times 10^{18}$ cm$^2) = 0.017$ millimeter per year. This is in surprisingly good agreement with estimates of present erosion rates: 0.06 mm/year for the United States (*58*), 0.03 mm/year (*54*), and upward from 0.02 mm/year for different mid-latitude basins (*59*). In this sense at least, uniformitarianism applies.

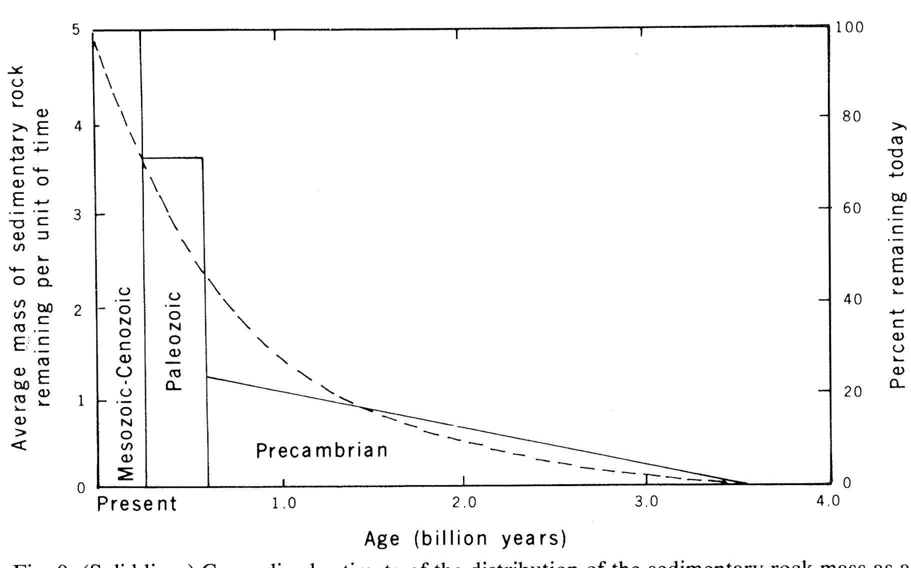


Fig. 9. (Solid lines) Generalized estimate of the distribution of the sedimentary rock mass as a function of geologic age. (Dashed line) Predicted distribution of rock mass according to either the "constant mass" or the "linear accumulation" model of Garrels and MacKenzie (*53*), assuming that the total sediment mass has been cycled five times throughout geologic time. [Modified from MacKenzie (*49*)]

Applications of the Theory

The earth-moon system. It has been pointed out to me (*60*) that the model of a globe-encircling sea prevailing for most of earth history may provide a solution for a problem concerning the evolution of the earth-moon system. Extrapolating the present rate of recession of the moon from the earth (*61*) backward in time brings the moon to the Roche limit (the distance of closest approach of a satellite to its planet, within which the satellite would disintegrate; for the earth-moon system this is ~2.89 earth radii) between 1.0 and 2.0 billion years ago. Although there is geologic evidence for strong tidal currents at various times in the Precambrian (*62*), there is no obvious evidence of approach that close in the geologic record on either the earth or the moon. The energy loss in the earth-moon system that causes the present recession of the moon is attributed primarily to tidal friction in shallow seas, such as the Bering Straits or the Irish Sea (*61*). There is no good geologic reason for assuming that the average rate of tidal dissipation of energy was any lower in the past, hence this "time scale problem" (*63*). However, if the sea were globe-encircling and relatively deep for much of Precambrian time, there would be much less energy loss due to tidal friction. After an early recession due to body tides, the earth-moon distance would increase very slowly, until continents became shallow or emergent. This would only occur toward the end of Precambrian time.

The strontium isotope problem. If the bulk of the sial was segregated early in earth history, and the crust was initially globe-encircling, then the material of the younger, apparently intracontinental, mobile belts must include (if not altogether consists of) reworked old "continental" crust. Reworking of preexisting basement is sometimes inferred by geologists (*7, 18, 26, 64*) but some geochemists consider that it is ruled out by isotopic data, particularly the $^{87}Sr/^{86}Sr$ ratio. Moorbath (*23, 65*) argues that reworking of old sialic crust with a high Rb/Sr ratio (>0.2, as in granites) should be revealed by a high initial $^{87}Sr/^{86}Sr$ ratio in the metamorphosed products. Most data show that the initial $^{87}Sr/^{86}Sr$ ratios are low (~0.700 to 0.702) in crystalline rocks of the Archean and early Proterozoic mobile belts, which suggests that the precursors of the metamorphic rocks involved were freshly derived from the mantle not long before their final metamorphism. Moorbath (*65*) interprets the isotopic evidence to indicate that the material of the continental crust has been progressively accreted by primary additions from the mantle, the converse of what is assumed in this article.

A solution of this problem (*66*), if one is possible, seems likely to be related to the fact that continental crust today is inhomogeneous. Rather than a uniform high Rb/Sr ratio (*67*), there is a vertical gradient in this and other geochemical parameters. Measurements of the heat flow in continents (*68*) indicate a near-surface concentration of the heat-producing elements, and a marked depletion in uranium and rubidium (and thorium) has been found (*69*) in the progression from amphibolite to granulite regional metamorphic facies. Zartman and Wasserburg (*70*) interpreted their data on the lead isotopic composition of 1-billion-year-old rocks partly in terms of a stratified continental crust. They proposed a model in which the Rb/Sr ratio decreases exponentially from an average value of 0.3 in the upper 10 km of a 35-km crust to less than about 0.05 in the lower third of the crust. The latter value is close to the ratio in the mantle, 0.03 (+0.02, −0.01) (*71*), and hence reworking of old, deep (granulite facies) crust should not be reflected in high initial $^{87}Sr/^{86}Sr$ ratios (*7*).

Considering the high early geothermal gradients (Fig. 5) in the model proposed in this article, I suggest that the upward concentration of incompatible elements was even more severe then than it is now. As the crust was thinner, the uppermost layer with the high Rb/Sr ratio may have been only a few kilometers thick. The plate tectonic analogy implies that the compressional phase of a mobile belt orogeny was preceded by tensional rifting, but with minor sea-floor spreading. Erosion of the upwarped rift margins (*72*), if they emerged above sea level, would thin the high-Rb/Sr crust. Volcanism and degassing attending this stage would contribute material with a low $^{87}Sr/^{86}Sr$ ratio to the mobile zone. With the thickening of the crust during the compressional phase, the upper, Rb-enriched zone would be preferentially eroded. The granulite facies (low Rb/Sr) rocks exposed would likely form the basement for subsequent sedimentary accumulations and would be involved in later metamorphism. The erosional debris, high in Rb, would have the opportunity to equilibrate with the primordial seawater, which had an $^{87}Sr/^{86}Sr$ ratio that diverged only slowly from that of the mantle (*73*). Thus, a low initial $^{87}Sr/^{86}Sr$ ratio may have been maintained in the bulk of the supracrustal deposits and in the reworked preexisting sialic crust basement. During orogenesis, partial melting of garnet granulites in such a crust might favor the production of anatectic melts with low $^{87}Sr/^{86}Sr$ ratios and relatively depleted in the heavy rare earth elements (*23*).

General. As the bulk of sediment today is provided by erosion in mountainous areas, it might be considered incorrect to equate the rate of sediment production simply with the rate of emergence of land. One would expect mountains to project above any shallow sea, and hence sediment production might

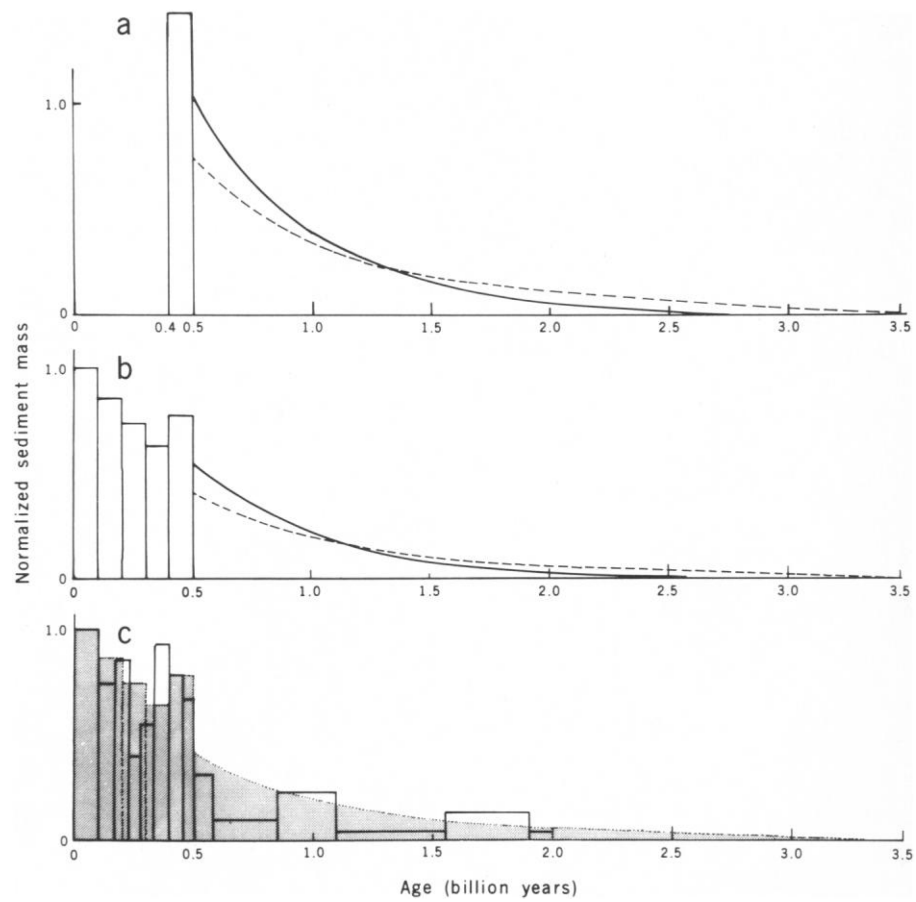


Fig. 10. Distribution of sediment mass with age according to the postulated two-stage model (*56*). The distribution was calculated on the basis of 35 intervals of 100 million years, and a sedimentary recycling rate of 14 percent of the accumulated sedimentary mass per 100 million years. The recycling rate was selected by trial and error. For the Phanerozoic it is equivalent to approximately 45 × 10^{12} g/year; in the best-fit models of Garrels and MacKenzie (*53*) the value is ~ 50 × 10^{12} g/year. In (a) and (b) two curves are drawn for the mass distribution before 0.5 billion years ago. The solid line is based on a direct correlation between land area and the amount of sedimentary rock eroded. The dashed line is based on the assumption that the proportion of sedimentary rock exposed on land increased from zero to the present value of 66 percent (*57*) by 0.5 billion years ago and has remained constant since. This assumption implies that an even higher percentage (as much as 11 percent) of the sedimentary rocks deposited 3.5 billion years ago still survive today. (a) Mass distribution 0.4 billion years ago, after the first of the two stages. A linear increase in the sediment mass to 80 percent of the present total as land emerged from 3.5 to 0.5 billion years ago is followed by the final stage of emergence and accelerated production of the remaining 20 percent of the sediment mass between 0.5 and 0.4 billion years ago. (b) Present mass distribution. During the last 400 million years the total sediment mass has remained constant at 32,000 × 10^{20} g, the fraction in Precambrian sediments being reduced to 14,000 × 10^{20} g (*53*). (c) Best estimate of the actual mass distribution for the past 2.0 billion years according to Garrels and MacKenzie (*53*). The shading corresponds to the model indicated by the dashed line in (b).

Table 2. Postulated stages in earth history.

Stage 1: 4.5 to ~3.5 billion years ago. Early segregation of the bulk of the sial and hydrosphere to form concentric shells as a result of extensive melting and vigorous mantle convection. The extremely unstable crustal regime continued until the average geothermal gradient became less than about 65°C per kilometer, when the sialic shell froze onto the mantle. Ancient, high-grade gneisses and associated supracrustals are relics of this stage.

Stage 2, Archean: ~3.5 to ~2.5 billion years ago. The rise of greenstone piles beneath globe-encircling sea followed freezing of the sialic shell to the mantle. Tectonics was dominated by the small-scale convective regime and led to the development of Archean cratonic nuclei surrounded by high-grade metamorphic mobile belts.

Stage 3, Proterozoic: ~2.5 to 0.6 billion years ago. Progressive segregation of the sial and hydrosphere into continents and oceans continued through separation and convergence of lithospheric plates induced by large-scale mantle convection; the scale of plate interaction and intercontinental drift increased with the growing oceanic area. Land area increased as the sea shallowed until continents of approximately present-day dimensions finally emerged.

Stage 4, Phanerozoic: 0.6 billion years ago to the present. Wilson cycle of plate tectonics and continental drift. The total area of continents and oceans is maintained relatively constant by subaerial erosion and subduction.

better be correlated with the rate of mountain building or orogeny. Higher heat flow and faster convection in earlier times would imply more vigorous tectonism and mountain building; hence one would predict more rapid sediment production then, the converse of what is proposed here.

The thermal constraint on thickness of continental crust (Fig. 5), however, also restricts the maximum height of mountains. Isostatic modeling shows that an 18.7-km sialic mountain column would be required just to reach the surface of a 2.0-km-deep sea (Fig. 6c). The critical 750°C isotherm (Fig. 5) would only reach that depth between 1.3 and 2.3 billion years after the earth formed. A mountain range 2 km above sea level requires a root more than 30 km deep, which would not be possible (Fig. 5) until at least 3.0 billion years after the earth formed. In other words, any sialic mountain belts projecting above sea level in Archean and early Proterozoic times would be puny by contemporary standards, and very short-lived, primarily because of subcrustal rather than surface erosion. The maximum possible isostatic height of mountains would thus gradually increase as the earth's average heat flow declined. This factor, I argue, counteracted the importance of mountains as a source of sediment in Precambrian times.

The timing of the changeover from predominantly submerged to emerged continents is crucial, yet most uncertain. A fundamental source of the uncertainty is lack of knowledge about the amounts and proportions of sial and hydrosphere after the early melting stage. They may not have been in the same proportions as today, and their separation may have proceeded episodically [associated with major orogenic episodes (*74*)] and unevenly, leading to periods of subsidiary emergence and submergence during the Proterozoic. The restriction on the thickness of continental crust imposed by higher geothermal gradients in the past (Fig. 5) implies that the continents emerged between 1.4 and 1.0 billion years ago. The shallowing of the sea and emergence of land presumed to be associated with the widespread deposition of banded ironstone between 2.3 and 1.7 billion years ago favors emergence during that interval. Yet the argument leading to the two-stage sediment accumulation model of Fig. 10, as discussed in the previous section, suggests that it occurred near the end of Precambrian time (*75*). This interpretation is incorporated in the postulated Precambrian crustal history summarized in Table 2.

In this simplified evolutionary model, well-known early Proterozoic detrital sequences such as the Witwatersrand system in South Africa and the Huronian in Canada cannot readily be explained. Although these deposits are areally insignificant on a global scale, there is no doubt that the continental crustal floor on which they accumulated was once near sea level; there is abundant evidence of fluvial to shallow marine deposition (*76*). In the Witwatersrand system up to 8 km of medium- to coarse-grained detritus was apparently deposited in a basin 350 km long by 200 km wide on a granitic crust (*76*); the tectonic circumstances allowing such an accumulation are an enigma. If these deposits were once much more widespread, then continental emergence must have occurred much earlier than 2.3 billion years ago. The critical question is whether such old detrital deposits are the exception or the rule. This would be one test of the evolutionary model proposed here.

The most appealing attribute of the late Precambrian emergence model (Fig. 10) is its implication that the total amount of sediment deposited and reeroded from the surface of the earth throughout geologic time has been much less than that required by a single-stage history, assuming recent rates for these processes. This indicates that we may have a fairly representative, although severely depleted record of Precambrian crustal history since 3.5 billion years ago—one that is much more complete than is implied by uniformitarian models. Attempts to decipher details of Precambrian geologic history need not necessarily be frustrated by the preconception of major gaps in the record.

References and Notes

W. S. Fyfe, *Philos. Trans. R. Soc. London Ser. A* **273**, 457 (1973); *Tectonophysics* **17**, 273 (1973).
2. F. Birch, *Geol. Soc. Am. Bull.* **76**, 133 (1965).
3. T. C. Hanks and D. L. Anderson, *Phys. Earth Planet. Inter.* **2**, 19 (19 F. P. Fanale, *Chem. Geol.* **8**, 79 (1971); F. M. Flasar and F. Birch, *J. Geophys. Res.* **78**, 6101 (1973); G. Arrhenius, R. DeBibhas, H. Alfven, *Mar. Chem.* **5**, 839 (1974); M. Ozima, *Geochim. Cosmochim. Acta* **39**, 1127 (1975).
4. H. H. Hess, in *Petrologic Studies: A Volume in Honor of A. F. Buddington*, A. E. J. Engel, H. L. James, B. F. Leonard, Eds. (Geological Society of America, New York, 1962), p. 599.
V. R. McGregor, *Philos. Trans. R. Soc. London Ser. A* **273**, 343 (1973); S. Moorbath, R. K. Onions, R. J. Pankhurst, *Nature (London) Phys. Sci.* **243**, 24 (1973).
W. S. Fyfe, *Philos. Trans. R. Soc. London Ser. A* **280**, 655 (1976).
7. D. Bridgwater and W. S. Fyfe, *Geosci. Can.* **2** (No. 3), 7 (1974).
8. D. McKenzie and N. Weiss, *Geophys. J. R. Astron. Soc.* **42**, 131 (1975).
9. G. J. Wasserburg, G. J. F. MacDonald, F. Hoyle, W. A. Fowler, *Science* **143**, 465 (1964).
J. C. Briden, *Philos. Trans. R. Soc. London Ser. A* **280**, 405 (1976).
K. M. Creer, E. Irving, S. K. Runcorn, *ibid.* **250**, 144 (1957).
12. M. W. McElhinny, *Paleomagnetism and Plate Tectonics* (Cambridge Univ. Press, Cambridge, England, 1973).
13. ———, J. W. Giddings, B. J. J. Embleton, *Nature (London)* **248**, 557 (1974).
14. T. N. Clifford, in *African Magmatism and Tectonics*, T. N. Clifford and I. G. Gass, Eds. (Oliver & Boyd, Edinburgh, 1970), p. 1.
15. J. C. Briden, *Nature (London)* **244**, 400 (1973).
16. J. D. A. Piper, J. C. Briden, K. Lomax, *ibid.* **245**, 244 (1973).
17. A. Brock, in *Implications of Continental Drift to the Earth Sciences*, D. N. Tarling and S. K. Runcorn, Eds. (Academic Press, New York, 1973), p. 33.
R. M. Shackleton, *Philos. Trans. R. Soc. London Ser. A* **273**, 317 (1973).
19. This conclusion was also reached by P. M. Hurley and J. R. Rand [*Science* **164**, 1229 (1969)] and A. E. J. Engel and D. L. Kelm [*Geol. Soc. Am. Bull.* **83**, 2325 (1972)] on the basis of geologic evidence.
20. J. Sutton and B. F. Windley, *Sci. Prog. London* **61**, 401 (1974).
21. If this interpretation is correct, the apparent Precambrian polar wander curves reflect mainly true polar wander and not continental drift.
22. A. Winchell, *Sketches of Creation* (Harper, New York, 1870); see also *Genesis*, chap. 1, verses 7 and 9.

23. S. Moorbath, *Proc. Geol. Assoc.* **86**, 259 (1975).
24. D. S. Burnett, *Rev. Geophys.* **13** (No. 3), 13 (1975).
25. E. F. Osborn, *Am. Mineral.* **47**, 211 (1962).
26. J. Sutton and B. F. Windley, *Philos. Trans. R. Soc. London Ser. A* **273**, 315 (1973).
→ D. Bridgwater, J. Watson, B. F. Windley, *ibid.*, p. 493; D. R. Hunter, *Univ. Witwatersrand Econ. Geol. Res. Unit Inf. Circ. No. 83* (1974); *Trans. Geol. Soc. S. Afr.* **73**, 107 (1970).
28. B. F. Windley and D. Bridgwater, *Spec. Publ. Geol. Soc. Aust. No. 3* (1971), p. 33.
→ C. R. Anhaeusser, *Philos. Trans. R. Soc. London Ser. A* **273**, 359 (1973).
30. B. F. Windley and J. V. Smith, *Nature (London)* **260**, 671 (1976).
31. C. R. Anhaeusser, R. Mason, M. J. Viljoen, R. P. Viljoen, *Geol. Soc. Am. Bull.* **80**, 2175 (1969).
→ B. F. Windley, *Philos. Trans. R. Soc. London Ser. A* **273**, 321 (1973).
33. A. M. Goodwin, *Proc. 23rd Int. Geol. Congr.* **1**, 69 (1968).
34. C. R. Anhaeusser, *Annu. Rev. Earth Planet. Sci.* **3**, 31 (1975).
35. ———, *Univ. Witwatersrand Econ. Geol. Res. Unit Inf. Circ. No. 98* (1975).
36. J. Tarney *et al.*, in *Early History of the Earth*, B. F. Windley, Ed. (Wiley, London, in press).
37. B. W. Darracott, *Trans. Geol. Soc. S. Afr.* **78** (part 1), 123 (1975).
38. A. E. J. Engel, *Univ. Witwatersrand Econ. Geol. Res. Unit Inf. Circ. No. 27* (1966); in *Adventures in Earth History*, P. Cloud, Ed. (Freeman, San Francisco, 1970), p. 431.
39. J. Veizer, *Contrib. Mineral. Petrol.* **38**, 261 (1973).
40. A. B. Ronov, *Geochemistry* **8**, 715 (1964).
41. A. M. Goodwin, *Econ. Geol.* **68**, 915 (1973).
42. R. W. Bayley and H. L. James, *ibid.*, p. 934.
43. H. L. James, *ibid.* **49**, 235 (1954).
44. N. J. Beukes, *ibid.* **68**, 960 (1973).
45. A. Button, *Univ. Witwatersrand Econ. Geol. Res. Unit Inf. Circ. No. 89* (1975).
46. H. Borchert, *Inst. Min. Metall. Trans.* **69**, 261 (1960).
47. H. D. Holland, *Econ. Geol.* **68**, 1169 (1973).
48. J. I. Drever, *Geol. Soc. Am. Bull.* **85**, 1099 (1974).
49. F. T. MacKenzie, in *Chemical Oceanography*, J. P. Riley and G. Skirrow, Eds. (Academic Press, New York, ed. 2, 1975), vol. 1, p. 309.
50. R. M. Garrels and E. A. Perry, Jr., in *The Sea*, E. D. Goldberg, Ed. (Interscience, New York, 1974), vol. 5, p. 303.
51. H. D. Holland, in *The Origin and Evolution of Atmospheres and Oceans*, P. J. Brancazio and A. G. W. Cameron, Eds. (Wiley, New York, 1964), p. 86; P. Cloud, *Econ. Geol.* **68**, 1135 (1973); R. M. Garrels, E. A. Perry, F. T. MacKenzie, *ibid.*, p. 1173.
52. H. P. Eugster and I-Ming Chou, *ibid.*, p. 1144.
53. R. M. Garrels and F. T. MacKenzie, *Evolution of Sedimentary Rocks* (Norton, New York, 1971).
54. C. B. Gregor, *Proc. K. Ned. Akad. Wet. Ser. B* **71**, 22 (1968).
55. D. U. Wise, *Geol. Soc. Am. Mem. 132* (1972), p. 87; in *The Geology of Continental Margins*, C. A. Burk and C. L. Drake, Eds. (Springer-Verlag, New York, 1974), p. 45.
56. I would like particularly to acknowledge the valuable discussions with and considerable help given by J. L. Means in the construction of this model.
57. H. Blatt and R. L. Jones, *Geol. Soc. Am. Bull.* **86**, 1085 (1975).
58. S. Judson and D. F. Ritter, *J. Geophys. Res.* **69**, 3395 (1964).
59. F. Ahnert, *Am. J. Sci.* **264**, 243 (1970).
60. M. G. Rochester, personal communication.
61. W. Munk, *Q. J. R. Astron. Soc.* **9**, 352 (1968).
62. P. E. Cloud, Jr., *Science* **160**, 729 (1968); R. B. Hargraves, in *Palaeogeophysics*, S. K. Runcorn, Ed. (Academic Press, New York, 1970), p. 471; P. M. Merifield and D. L. Lamar, in *ibid.*, p. 31; *J. Geophys. Res.* **73**, 4767 (1968); D. L. Lamar and P. M. Merifield, *Geol. Soc. Am. Bull.* **78**, 1359 (1967); W. S. Olson, *Am. Sci.* **54**, 458 (1966).
63. W. M. Kaula and A. W. Harris, *Rev. Geophys. Space Phys.* **13**, 363 (1975); W. M. Kaula, *ibid.* **9**, 217 (1971).
64. K. D. Collerson, C. W. Jesseau, D. Bridgwater, in *Early History of the Earth*, B. F. Windley, Ed. (Wiley, London, in press).
65. S. Moorbath, *Nature (London)* **255**, 391 (1975); *ibid.* **254**, 395 (1975).
66. J. Tarney, in *Early History of the Earth*, B. F. Windley, Ed. (Wiley, London, in press).
67. G. Faure and J. L. Powell, *Strontium Isotope Geology* (Springer-Verlag, New York, 1972).
68. R. F. Roy, D. D. Blackwell, F. Birch, *Earth Planet. Sci. Lett.* **5**, 1 (1968); A. H. Lachenbruch, *J. Geophys. Res.* **75**, 3291 (1970).
→ K. S. Heier, *Philos. Trans. R. Soc. London Ser. A* **273**, 429 (1973).
70. R. E. Zartman and G. J. Wasserburg, *Geochim. Cosmochim. Acta* **33**, 901 (1969).
71. P. W. Gast, in *The Nature of the Solid Earth*, E. C. Robertson, Ed. (McGraw-Hill, New York, 1972), p. 19.
72. D. J. J. Kinsman, in *Petroleum and Global Tectonics*, A. G. Fischer and S. Judson, Eds. (Princeton Univ. Press, Princeton, N.J., 1975), p. 84.
73. Z. E. Peterman and C. Hedge, in *Handbook of Geochemistry*, K. H. Wedepohl, Ed. (Springer-Verlag, Berlin, 1974), vol. 2, part 4, p. 38.
74. G. Gastil, *Am. J. Sci.* **258**, 1 (1960).
75. The final stage of the emergence of land postulated in this model apparently coincides with the rapid spread of life at the end of Precambrian time [S. M. Stanley, *Am. J. Sci.* **276**, 56 (1976)].
76. D. A. Pretorius, *Miner. Sci. Eng.* **7**, 18 (1975).
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Trophic Regulation of Nerve Sprouting

Neuron-target interactions and spatial relations control sensory nerve fields in salamander skin.

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Studies on developing nervous systems have revealed that nerve cells can change their shape by sprouting axonal branches and reabsorbing old ones. The physiological connections which these branches make with their target cells are the basis for the establishment of circuitry in the nervous system. In studying nerve sprouting, we need to consider both the stimulus for its initiation and the way it is controlled. This control applies to the area over which endings are distributed and to the density of the endings within it. These two parameters, density and area, define the terminal field of a nerve. The control of sprouting then is a principal means of determining how nerve fields develop. This development could be regulated entirely by a rigid genetic program intrinsic to either the target tissue or the nerve, but a potentially more interesting mechanism would have a competence to respond to internal and external environmental demands. If so, then the dynamic regulation of terminal fields, including sprouting and possibly regression of endings, may be a normal feature of both central and peripheral neurons, even in the mature organism. This article deals with investigations of the regulation of terminal fields in a readily accessible peripheral system. It appears that an interaction between the nerve and the target tissue controls the density of the endings, while the area of a terminal field is more determined by spatial relations.

Sprouting During Development

In his observations on the genesis of epithelial innervation, Ramón y Cajal detected an important influence from the target tissue (*1*). He observed that the incoming fibers often grow relatively long distances to reach the epithelial tissues, but only after arriving at them do the nerves start sprouting collateral branches, each growing to a territory devoid of nerves. This sprouting eventually stops, and Ramón y Cajal noted the absence both of any vast aneuritic spaces and of any excessive collection of nerve fibers. He suggested that there are growth-pro-

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