Precambrian Perspectives

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Summary. The Precambrian record is interpreted in terms of an evolutionary progression that moves in the direction of increasing continental stability. An early, highly mobile microplate tectonics phase progressed through a more stable, largely intracratonic, ensialic, mobile belt phase to the modern macroplate tectonics phase that involves large, rigid lithospheric plates. Various phases are characterized by distinctive crustal associations. Three controls—bulk earth heat production, crustal fractionation and cratonization, and atmospheric oxygen accumulation—are viewed as the cumulative cause of the trends and events that characterize the crust at different stages of development, from its inception approximately 4.6 billion years ago to the present.

Precambrian time deals with the long period of Earth history from its inception 4.6 billion years ago to the development of macro life forms 4.0 billion years later—the start of Phanerozoic time (Fig. 1). Our knowledge of modern plate tectonics, involving global motions and interactions of large, rigid lithospheric plates, greatly influences our interpretation of the Precambrian. However, the present tectonic model could inhibit interpretation of the Precambrian record. Realistic Precambrian perspectives are needed for a balanced view of Earth history.

Five basic assumptions are made in this article about the evolution of the earth. (i) The planet began homogeneously as aggregated particles, hot or cold, initiating an early brief period of extravagantly high accretion-induced heat flow (*1*). (ii) The earth, like the moon and other solar bodies, was subjected to impacts by very large meteorites, especially in the period 4.2 to 3.9 billion years ago (*2*). (iii) The earth has maintained a nearly constant diameter during the past 3.8 billion years (*3*). (iv) The principal source of the earth's internal energy is radiogenic decay, with possible primordial supplement, steadily diminishing with time (*1*). (v) The earth's crust and lithosphere, the repository of the geologic record, have responded to this changing geothermal regime. The challenge is to locate and interpret correctly the evidence in the Precambrian crust for this unidirectional evolution.

Two aspects of the Precambrian pose particular problems.

1) The modern oceanic crust, accounting for 80 percent by surface area and 50 percent by volume of the total crust (oceanic plus continental), is transitory. It is practically all consumed on return to its mantle source. The small unconsumed sialic fraction has been added periodically to the growing continents. This small fraction alone constitutes our direct Precambrian record. However, modern oceanic crust itself has influenced our perception of current tectonic processes. The virtual absence of preserved Precambrian counterparts deprives us of comparable Precambrian insight. Nevertheless, Precambrian tectonic interpretations must allow for large dynamic Precambrian oceans.

2) Continental reconstructions commonly incorporate the paleomagnetic record. However, as a method of continental reconstruction, paleomagnetism has severe limitations, the primary paleomagnetic data normally providing only paleolatitudes and pole directions. Accurate continental reconstructions rely on closely spaced, step-by-step positionings from known continental sites. The older the rocks, the more uncertain the reconstruction.

The Precambrian record is interpreted in this article in terms of an evolutionary progression. Various phases are characterized by distinctive crustal associations. There are few, if any, sharp demarcations between phases; phase boundaries as defined (Fig. 1) are arbitrary and may even transect a continuum. Nevertheless, the Precambrian record is viewed as moving in the direction of increasing continental stability from an early highly mobile microplate tectonics phase, through a more stable, largely intracratonic, ensialic, mobile belt phase, to the modern macroplate tectonics phase involving large, rigid lithospheric plates with modern continents and ocean basins.

Crustal Controls

Three interrelated factors have controlled the unidirectional development of the continental crust: (i) bulk Earth heat production and loss, (ii) crustal fractionation and cratonization, and (iii) the rise in atmospheric O_2 level. These three factors, together with pervasive gravitational control, operated cumulatively, and all first-order crustal features have developed in response to them.

Bulk Earth heat production. The earth's crust preserves a record of mantle-derived silicate melts produced during 3.8 billion years. This continuing irreversible differentiation is closely linked to the evolution of the atmosphere, which, like the crust, is of secondary origin (*1*). Heat production by radioactive isotopes of potassium, thorium, and uranium has been the dominant factor in producing melting and consequent differentiation. Other factors may have contributed significantly (*4*). The crust in its various forms has developed in response to this internal source by transfer from the mantle of heat, of heat-producing elements, and of the major elements forming the bulk of the crust and atmosphere (Fig. 1).

The present heat production is measured at 11 picowatts per kilogram and is estimated to have declined steadily from an initial level of 53 pW/kg 4.5 billion years ago, as judged from potassium, thorium, and uranium. Although the exact timing and rates of supply of materials to the crust and atmosphere are uncertain, the bulk Earth heat production is the driving force of tectonic processes and the cause of crust and atmosphere development.

Crustal fractionation and cratonization. A consequence of mantle differentiation and transport of materials to the earth's surface has been the development of sialic or continent-building crust and its aggregation into stable cratons. Although viewpoints differ on timing (*5*), estimates of the growth of continental crust and of the cratons are provided by

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crustal fractionation (*6*) and cratonization (*7*) curves (Fig. 1). Other growth curves have also been evaluated (*8*). The continental crust during the early Archean was apparently very limited, the segments existing 3.0 billion years ago probably constituting less than 12 percent of the present-day continental area. The cratonized segments were even smaller, there being a consistent time lag between fractionation and cratonization.

According to the preferred growth curve (*8*), the process of crustal fractionation and cratonization operated at an accelerated rate in the late Archean and early Proterozoic (3.0 to 2.0 billion years ago). During this time, crustal fractionation is estimated to have increased from 12 to 53 percent of the present continental crust, and cratonization is estimated to have increased from 4 to 25 percent of its present value. This dramatic increase in the quantity and stability of the continental crust profoundly influenced the upper crust and ocean-atmosphere systems. Broad, stable continental shelves, including marginal and intracratonic basins, were widely developed. The growth of stable cratons appears to have passed a threshold about 2.5 billion years ago, leading to the development of sedimentary environments more closely comparable to modern counterparts than had hitherto existed.

Rise in atmospheric O_2 level. The onset of Proterozoic tectonic and sedimentary regimes, including relatively shallow continent-margin basins and shelves, resulted in the radiation of microbial mat communities into shallow-water environments. This biologic explosion greatly enhanced the photosynthetic production of O_2, which resulted in a dramatic increase in the level of atmospheric O_2 [for another view, see (*9*)]. According to the preferred atmospheric O_2 level curve (*10*) (Fig. 1), the level rose from an estimated 1 percent of the present level 2.0 billion years ago to 10 percent of the present level 0.7 to 0.6 billion years ago, the start of the Phanerozoic. This dramatic rise in the O_2 level had profound effects on the biologic and sedimentary processes.

Although chemical evolution gave way to biogenesis at a very early stage in Earth history, more than 3.8 billion years ago, the rapid rise in the level of photosynthetically produced O_2 had to await development of stable continental shelves. Thereafter, improved biogenic mediation, continued carbon segregation, and the filling of O_2 sinks, particularly in the form of oxide-facies banded iron formation, initiated atmospheric O_2 buildup. The consequences were essential termination of the deposition of banded iron formation, onset of red beds, and O_2 shielding of anaerobic intracellular processes, heralding the eukaryotic cell (*10*). Thus, the period 2.7 to 2.0 billion years ago apparently marks a singular conjunction of critical trends in Precambrian crustal history.

This comparatively sudden transition from a mantle-dominated crust to a substantial continental presence, about 2.5 billion years ago, is illustrated in the ratio of strontium-87 to strontium-86 in carbonate rocks (*8*) (Fig. 1). Whereas high heat flow and high mantle flux dominated Archean ocean chemistry, there was, in early Proterozoic time, an increase in the continental flux, involving recycling of continental components and leading to a transition to mixed continental and mantle flux expressed in a substantial increase in $^{87}Sr/^{86}Sr$ ratios to about 0.708—well above the mantle growth line. Similar increases are expressed in other ratios, including K_2O/Na_2O, La/Yb, and rare earth element abundances across this same critical transition (*8*).

Crustal Trends

The main resulting Precambrian crustal trends are briefly described in the following categories: orogenies and other thermotectonic events, biogenesis, rock types and tectonic setting, and metallogenesis (Fig. 1).

Orogenies and other thermotectonic events. Orogenies generally refer to the processes by which structures within mountain areas were formed—including deformation, metamorphism, and magmatism. Such processes affected rocks along restricted zones and within a limited time interval; in this sense they are space-time events. The cause of orogenies is Earth heat production. Although this has apparently declined steadily with time (Fig. 1), it is translated at the surface in episodic space-time events (*11*). The cause of episodicity is not fully understood but may be a function of thermally induced mantle convection currents and periodic changes in such systems. Linear mountain belts are generally attributed to linear plate convergence. However, Precambrian thermotectonic events have produced such diverse phenomena as linear mountain belts—in the modern sense—to broad, thermally affected domains that are magmatically intruded, yet tectonically undeformed. Consequently, the more neutral term thermotectonic event (*12*) is used for those thermally induced events that did not necessarily involve mountain building, with implications of converging plates.

The characteristic pattern of cratonic blocks bordered by mobile belts, which has been a feature of the continental crust through most of recorded geologic history, seems to have developed mainly between 2.8 and 2.5 billion years ago as a result of the first widespread phase of stabilization—involving massive granitoid emplacement—of previously commonly mobile tracts (for example, the Kenoran Orogeny) (*13*). The effects of this particular style of orogeny are widely documented in most shields of the world and provide the basis for the Archean-Proterozoic division of Precambrian time (Fig. 1). Other orogenic peaks are recorded. One occurred in the northern or Laurasian Precambrian shields 1.9 to 1.6 billion years ago (the Hudsonian Orogeny), and another, more restricted, occurred 1.2 to 0.9 billion years ago (the Grenvillian Orogeny). The Pan-African mobile belt network, a major crust-forming event in the period 0.8 to 0.5 billion years ago, affected virtually all southern or Gondwanaland continents. Other orogenic peaks have been recorded in the Phanerozoic. The nature and extent of the corresponding early Archean events are uncertain.

Biogenesis. The consequence of major outgassing and the development of the atmosphere and hydrosphere was chemical evolution, leading in stages through amino acids, polypeptides, RNA, and DNA to anaerobic heterotrophs (organisms unable to manufacture their own food) and prokaryotes (unicellular organisms lacking a nucleus) (*10*). Highly metamorphosed sedimentary rocks in the 3.8-billion-year-old Isua terrain of West Greenland yield some indications of photoautotrophic biologic activity in the form of carbon isotope ratios. The oldest known microfossils, however, occur in the 3.5-billion-year-old Warrawoona Group of Western Australia, where wavy to planar laminated cherts, probably stromatolitic, contain well-preserved filamentous microfossils (*14*). Younger fossiliferous Archean rocks include microfossils and stromatolites in the Swaziland Supergroup, South Africa, of age 3.4 billion years, and stromatolites and cyanobacteria in the Fortescue Group, Western Australia, of age 2.8 billion years. Other important Precambrian fossil assemblages include the diverse cyanobacteria, bacteria, and stromatolites in the Gunflint Formation, Ontario, of age 2.0 billion years; cyanobacteria and eukaryotes (organisms with a nucleus) of the Pahrump Group, California, of age 1.3 billion years; diverse cyanobacteria and eukaryotes of the Bitter Springs Formation, central Australia, of age 0.9 billion years; and the metazoans (multicellular organisms) of the Adelaidean System at Ediacara, Australia, of age 0.7 billion years.

With the rapid increase of the atmospheric O_2 level to about 1 percent of the present level, about 2.0 billion years ago, intracellular isolation of anaerobic vital processes became essential. This led to the development and diversification of the eukaryotic cell about 1.4 billion years ago. The next big step on the evolutionary path was the origin of sexuality. This occurred before 0.7 billion years ago, the age of the oldest known Metazoa. The first metazoans were dependent on simple diffusion for O_2; exoskeletons appeared later, perhaps 0.6 billion years ago, when increasing O_2 levels favored emergence of more advanced respiratory systems. Subsequent evolution produced the oldest trilobites 0.57 billion years ago and the oldest hominids 0.004 billion years ago.

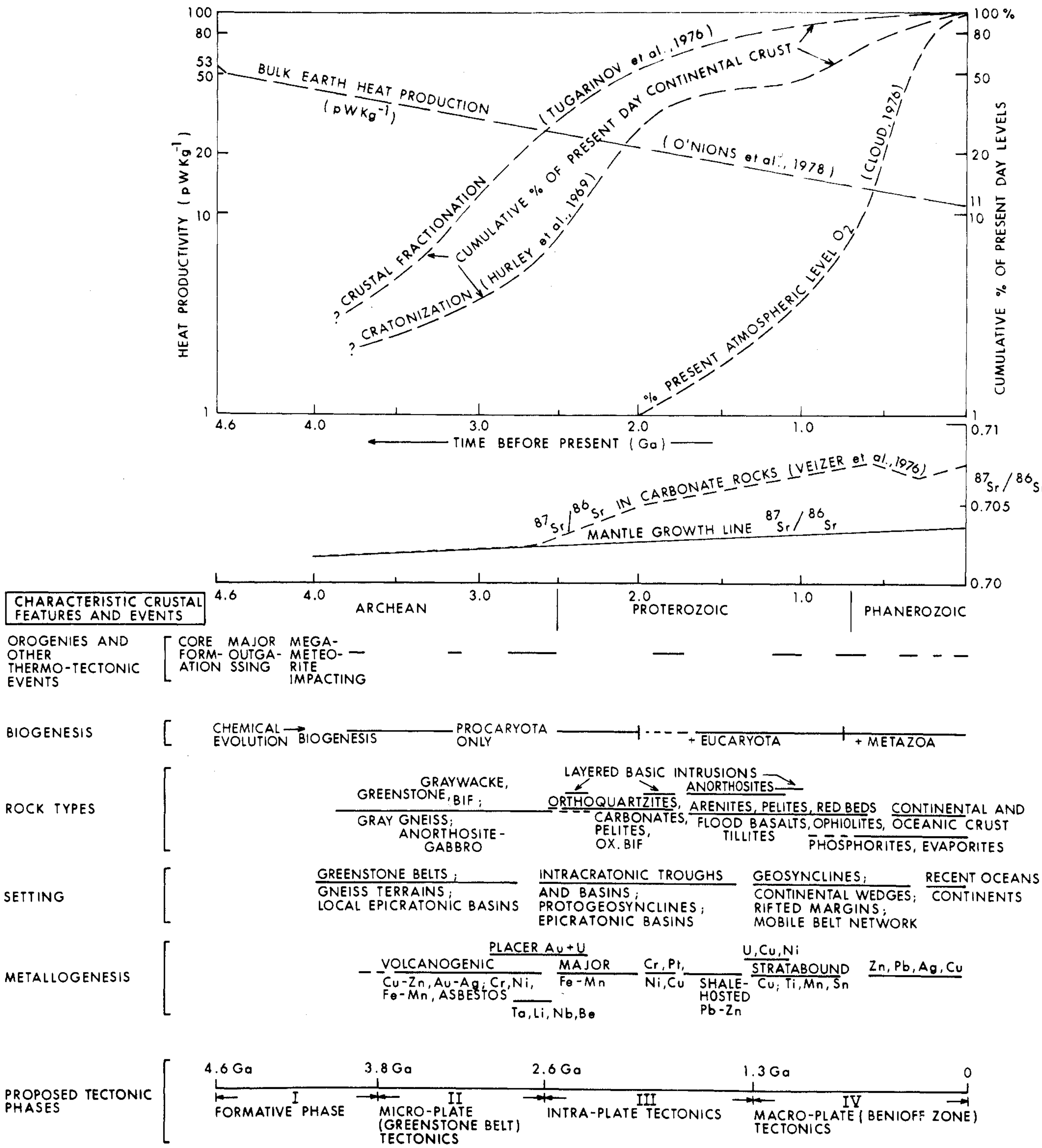

Fig. 1. Crustal controls (above) and trends (below) in development of the earth's continental crust. *Ga*, time expressed in billion years.

Rock types and tectonic settings. A three-part division of Precambrian time is used: Archean (> 2.5 billion years ago), early Proterozoic (2.5 to 1.5 billion years ago), and late Proterozoic (1.5 to 0.6 billion years ago).

1) Archean crust is present in all the main Precambrian shields. The oldest dated rocks, in West Greenland, slightly exceed 3.8 billion years (*15*), little younger than the completion of major impacting by meteorites. The main rock types—metamorphosed granitoids, clastic and chemogenic sediments, and mafic intrusions, all of otherwise unremarkable aspect—suggest that the craton-building process, parent of the future continents, was well under way 3.8 billion years ago. The geologic record then becomes increasingly voluminous. Well-preserved assemblages 3.5 to 3.0 billion years old are known in South Africa and Western Australia; and younger Archean assemblages are plentiful in most shields. Two principal Archean rock associations, granite-greenstone and high-grade gneiss, are so disposed that it is clear that each influenced the development of the other. A third association, epiclastic basin deposits, is locally important.

Greenstone belts are deformed, elongate, metavolcanic, and metasedimentary units that include substantial gneissic to massive granitic intrusions. The belts are commonly dispersed in predominantly granitic terrain. Individual belts range in size from local schist units to major belts 100,000 square kilometers in area. Most represent linear to curvilinear, steeply inclined, pluton-intruded, volcanic-rich isoclinal keels that are relics of originally much larger units. Greenstone belts range in age from 3.5 to at least 1.8 billion years, with a notable cluster at 2.7 billion years.

The belts contain various proportions of volcanic and sedimentary rocks, most being mainly volcanic. Volcanic compositions are dominantly tholeiitic (iron-bearing mafic volcanic rocks), but include komatiitic (primitive magnesium-rich mafic volcanic rocks) and calc-alkalic (more highly evolved volcanic rocks) parts. A mafic to felsic compositional progression with time is common. Komatiites are widely distributed, though not necessarily abundant. Their extrusion is considered to have required high magma temperatures at the surface, possibly 1780°C 3.3 billion years ago (*4*). The virtual absence of the peridotitic (high magnesium) variety of komatiite in post-Archean rocks suggests a substantial change at the Archean-Proterozoic boundary to lower thermal regimes. Tholeiitic basalts, the dominant greenstone component, are attributed to crystal fractionation from a mantle-derived basaltic parent. The calc-alkalic andesites are comparable in significant respects to those found in some modern island arcs. Felsic volcanic rocks are commonly clustered at volcanic centers, marking local subsidence structures.

Most greenstone belts have intrusive-migmatized margins that have obliterated the original basement relations. Prevolcanic granitic crust, not necessarily basement, has been identified in the vicinity of practically all greenstone belts. Thus most volcanic rocks apparently were extruded near or upon older sialic crust. However, such sialic crust was not necessarily basement to the volcanics, much of which probably accumulated in intervening rifts.

Low initial $^{87}Sr/^{86}Sr$ ratios, widely reported in Archean volcanic rocks and accompanying granitoids, are interpreted as indicating massive additions of juvenile, mantle-derived material to the crust without substantial involvement of older sialic crust by way of melting, assimilation, and mobilization (*16*). Ophiolites and blue schist (high-pressure) assemblages, both characteristic of modern convergent plate boundaries, are virtually absent in Archean terrains. Taken together with the rarity of kyanite and eclogite, both high-pressure products, these facts suggest a prevailing hotter, thinner lithosphere model lacking subduction (Benioff) zones during the Archean. An occurrence of kyanite in ancient Archean crust at Isua, West Greenland, however, indicates that abnormally steep gradients were not universal in the early Archean (*17*). Crustal conditions appear to have been highly variable, possibly the result of a rift-and-sag mechanism, involving rifting of older sialic crust and leading to massive accumulation of mantle-derived volcanic material in conjunction with crustal sag. Sagging could yield copious tonalite-granite intrusions by partial melting of the sagged crust, in harmony with the $^{87}Sr/^{86}Sr$ systematics mentioned above.

High-grade metamorphic terrains occur as isolated patches, as long linear belts, and as broad irregular zones of subprovince status (for example, the Ungava Belt of the Superior Province, Canadian Shield). The commonest rocks are generally granulite to upper amphibolite facies; but they contain the remains of some of the earliest volcanic and sedimentary rocks, as well as layered igneous complexes. The general complexity in these terrains makes it difficult to decipher the origin of the rocks. Archean high-grade terrains commonly lie alongside the low-grade greenstone belts. The high-grade terrains are generally attributed to regimes of comparatively high temperature and pressure, and their present crustal level is attributed to substantial vertical uplift of once deeply buried crust. The wide distribution of high-grade terrains and their proximity to low-grade terrains points to major vertical uplift in development of Archean crust, regardless of any horizontal motions.

The high-grade terrains have been interpreted as older than or synchronous with the greenstone belts; in the latter case, they are considered either deeper or lateral equivalents. In many cases, the dissimilarity in rock types and proportions between the two makes it unlikely that the one is simply the metamorphosed equivalent of the other. Thus the two are generally considered to represent complementary parts of Archean crust in growth.

Despite evidence of widespread crustal instability during the Archean, stable crust had developed locally by 3.0 billion years ago, as evidenced by the presence of epicratonic basins. Some of the oldest known, well-preserved, cratonic sediments, which occur in Swaziland and adjoining parts of southern Africa, involve a series of adjoining, northward younging, epicratonic basins. The Pongola Group, 3.0 to 2.8 billion years old, comprises older volcanic (mainly basalt) and younger sedimentary rocks. To the north, the Dominion Reef Group, about 2.8 billion years old, is a locally developed conglomerate-arenite-volcanic assemblage. It is, in turn, uncomformably overlain by the Witwatersrand Supergroup, 2.7 to 2.5 billion years old, composed of underlying shales, quartzites, banded iron formations, and mafic volcanics and uncomformably overlying fluvial conglomerates and arenites. Witwatersrand Supergroup is the world's most productive source of gold and also contains significant uranium mineralization, both distributed in conglomerate sheets developed on low-angle unconformities in the upper Witwatersrand. Following unconformably on either the Witwatersrand or adjoining basement rocks is the 2.6-to-2.5-billion-year-old Ventersdorp Supergroup, mainly subaerial basalt flows with minor felsic volcanics, volcaniclastics, and chemical sediments. Finally, the unconformably overlying early Proterozoic strata of Transvaal Basin, 2.3 billion years old, continue the epicratonic cover successions in this region.

Southern Africa thereby provides an outstanding example of early stabilization of Archean crust. Most of the granites were emplaced by 3.0 billion years ago; and shortly after that the craton had been intruded by basic dikes and planed off by erosion to receive its first sequence of extensive cratonic cover successions. This series of epicratonic cover successions does not fit the conventional Archean-Proterozoic boundary at about 2.5 billion years ago. Despite this, the end of the Archean appears to have marked the most dramatic change in the evolution of crust after major meteorite impacts. The culminating Archean orogeny, 2.8 to 2.5 billion years ago, involved major new crustal additions and introduced a new dimension of craton size and stability, which characterized the oncoming Proterozoic.

2) In response to increased stability, the early Proterozoic crust, 2.5 to 1.5 billion years ago, featured thick sedimentary-volcanic accumulation, both in large ensialic basins and in mobile belts. These locally contain, in time sequence, rich uranium-gold placer deposits, very large banded iron formations, and major shale-hosted stratiform lead-zinc (-copper) deposits. Additional characteristics include layered mafic intrusions, some possibly impact melts, with platinum and nickel deposits; diabase dike swarms; and the appearance of major red beds. The supracrustal rocks are locally flatlying or gently dipping; elsewhere deformation varies from mild warping, with faulting and intrusion, to extensive asymmetric folding, thrusting, intrusion, and metamorphism. During this period, it appears that crustal masses of continental or even supercontinental size, composed of stable cratons and mobile belts, were moving as units. These masses featured stable cratons enclosed in mobile belts.

Ensialic basins, 2.7 to 2.3 billion years old (for example, the Huronian, Canada; Witwatersrand, South Africa; Jacobina, South America), each hundreds of kilometers long, feature sandstones, quartzites, pelites, local dolomites, and paraconglomerates. They contain large quantities of uraniferous or auriferous (or both) quartzitic detritus, apparently supplied by rapid multistage erosion of the Archean granitic crust. Abundant placer pyrite and uraninite indicate a prevailing anaerobic environment. At Rum Jungle, Australia, pitchblende and pyrite mineralization of similar age is associated with copper, cobalt, and lead sulfides.

Interior belts and troughs as much as 800 km long contain orthoquartzites, dolomites, shales, graywackes, and volcanic rocks. Some are characterized by immense accumulations of oxide-facies banded iron formation. The period 2.5 to 2.0 billion years ago represents the time of greatest accumulation of sedimentary iron in Earth history. Five iron districts, each classified (*18*) as very large (about 10^{14} tons banded iron formation), together contain more than 90 percent of all known banded iron formation on Earth (Hammersley Range, Western Australia; Transvaal-Griquatown belts, South Africa; Quadrilatera-Ferrifero, Brazil; Labrador Trough, Canada; and Krivoi Rog, U.S.S.R.).

A plausible model to explain these immense banded iron formations involves a stratified Precambrian ocean with upper oxygenated layer and lower anaerobic zone (*19*). The latter is inferred to have held in solution an enormous mass of Fe^{2+}. The source of the iron may have been diverse—perhaps the greatest part from ocean-floor activities and another part from land mass weathering. The iron may have accumulated in the ocean during a very long period of Archean time, or comparatively quickly in response to heightened seafloor activity. The massive iron precipitation may be reasonably ascribed to the welling up of deeper Fe^{2+}-bearing ocean water from the anaerobic zone into the upper oxidizing layer. The worldwide appearance of the great banded iron formations 2.5 to 2.0 billion years ago may reflect a favorable conjuction of circumstances, including the development of wide continental shelves with access to deeper ocean basins; a possibly augmented supply of iron to the ocean; and greatly increased photosynthetically produced O_2. In addition to the iron, large manganese ore bodies are present in the Transvaal Supergroup, South Africa; Mount Bruce Supergroup, Western Australia; Amapa Series, Venezuela; Minas Geraes, Brazil; and the Aravalli Group, India.

Numerous interior mobile belts display recognizable mio- and eugeoclinal facies. They are commonly classified as ensialic (denoting sialic basement without discernible involvement of oceanic crust). This is supported by paleomagnetic evidence (*20*). However, the Coronation Geosyncline, situated at the northwest margin of the Canadian Shield, exhibits a depositional polarity with associated aulacogens that could reflect a continent collision development.

Younger large interior basins are typified by the McArthur Basin, 1.8 to 1.4 billion years old, in Australia. This basin is preserved as gently folded and comparatively little altered assemblages of platform cover rocks, largely pelites with some dolomites, and local volcanics. The most important ore bodies are the stratiform lead-zinc-silver ores of the McArthur River deposits. To the south, at Mt. Isa, the ensialic geosynclinal shale-sandstone-dolomite accumulations, now partly deformed and metamorphosed, include large stratiform silver-lead-zinc, copper, and uranium deposits. Still further south, at Broken Hill in the Willyama block, granulite facies meta-argillites and meta-arenites contain large lead-zinc-silver lodes. Large argillite-enclosed stratiform lead-zinc deposits at Sullivan, Canada, are of similar age (1.7 to 1.4 billion years), as are similar lodes at Black Angel, West Greenland.

Very large layered mafic complexes were emplaced in stable crust about 1.9 billion years ago. The Bushveld complex, South Africa, includes the world's largest mafic igneous body. This layered peridotite-pyroxenite-gabbro-anorthosite mass with overlying rhyolite contains very large ore bodies (chromium-platinum-titanium-tin-nickel-vanadium). The Sudbury Irruptive, Canada, a layered norite-micropegmatite body, ascribed to meteorite impact, is famous for its associated nickel-copper-platinum-cadmium lodes. The somewhat older (2.7 to 2.5 billion years) Stillwater Complex, Montana (*21*) and Great Dyke, Zimbabwe also contain substantial chromium-platinum ores. On an even larger scale, numerous diabase (dolerite) dike swarms about 2.2 billion years old, are present in the cratons (North America, Scotland, Africa, India, and Australia). All these activities attest to the presence of stable continental crust exposed to regional tensional stresses.

3) Late Proterozoic crust (1.5 to 0.6 billion years ago) featured Cordilleran-type geosynclines at continental margins, including thick continental terrace wedges, and well-developed interior geosynclines featuring miogeoclinal sedimentation. Red beds and other continental sandstones are common. Extensive basins and troughs filled with coarse continental sandstones and flood basalts, together with alkaline complexes and mafic dike swarms, point to rifting of broad continental cratons as a major process for the first time. Massif-type anorthosites were emplaced in the crust on a singular scale. A remarkable global system of late Proterozoic–early Paleozoic mobile belts marks the transition to Phanerozoic crustal regimes.

Cordilleran-type geosynclines were developed near the margins of North America and western Europe, as illustrated by the Belt-Purcell-Windermere successions (1.3 to 0.6 billion years old) of the Cordilleran Geosyncline. This geosynclinal succession represents a westward prograding continental terrace wedge derived from a constant cratonic source to the east.

The Adelaidean System of Australia, a well-developed geosyncline (1.4 to 0.6 billion years old) includes the Adelaide Geosyncline, Kimberley Basin, and Amadeus Basin. It represents an unusually extensive, long-lived platform cover succession with a wide range of mineral deposits, notably the very large paleosurface uranium-copper deposit at Olympic Dam on Roxby Downs.

Broad sandstone-filled basins (1.5 to 1.1 billion years old), many with red beds, are distributed across the Baltic, Greenland, and Canadian shields, and, mainly in the subsurface, as far west as Arizona. Large uranium-copper-nickel ore deposits of the unconformity or paleosurface type are associated with the Athabasca Basin (western Canada). Thick accumulations of continental tholeiitic to alkali basalt with local rhyolite and associated mafic intrusions (Muskox and Duluth complexes) occupy, or lie near, rifted epicratonic troughs. The Lake Superior Basin, one such trough, lies at the northern limit of the midcontinent gravity high, an immense continental rift some 2000 km long, 50 to 150 km wide, and 20 to 30 km deep, filled with great masses of basaltic magma and local rhyolite and conglomerate, all about 1.1 billion years old. Sediment-hosted strata-bound copper deposits are present at White Pine, Michigan. Similar large shale-hosted copper deposits occur in the Katangan Belt, Central Africa. Diabase dike swarms, 1.2 to 1.0 billion years old, extend across the Baltic and Canadian shields to Arizona and California, marking continental fracturing at a large scale. Alkaline complexes, including carbonatites, 1.3 to 1.0 billion years old, are distributed along fault-controlled belts and incipient rifts in the Baltic, Greenland, and Canadian shields as far west as New Mexico.

A remarkable and distinctive magmatic event in late Proterozoic time was the coeval emplacement of massif-type anorthosites, granites, and acid volcanic rocks, mainly in the period 1.6 to 1.2 billion years ago. The scale of emplacement of these anorogenic igneous bodies is astonishing and seemingly occurred only once in the history of the crust. Most anorthosites lie in two principal global belts each about 6000 km long and corresponding to the disposition of Precambrian shields in a predrift continental reconstruction (*22*). Anorthosite ages range from 3.0 to 0.8 billion years but tend to cluster around 1.3 ± 0.2 billion years. Thousands of individual round or ovoid anorthosite bodies are present, most of them in the size range 100 to 10,000 km^2. They contain more than 90 percent plagioclase. Approximately 20 percent of the Grenville Province, eastern Canadian Shield, is underlain by anorthosite massifs; and taken together with similar anorthosite massifs in the adjoining Nain Province to the northeast, this belt of massifs has been estimated to comprise perhaps 75 percent of the world's anorthosite (*23*). Closely associated rapakivi granites and coeval acid volcanic accumulations are particularly common in the subsurface of south-central and western United States. The volume of magma represented by these anorthosite-granite-acid volcanic associations is staggering for an anorogenic setting; nothing like it is known elsewhere in the earth's geological record. They have no obvious relation to any plate tectonic process. Anorthosites typically occur in a deep crustal setting; they are characteristically associated with high metamorphic-grade rocks of the granulite facies. These facts, together with their space-time concentration, suggest that the development of massif-type anorthosites on such a large scale occurred at a particular time in the cooling history of the earth, a time that roughly coincides with the changeover to the modern plate tectonics process (Fig. 1).

A remarkable global system of late Proterozoic-Paleozoic mobile belts, a veritable network of mobility, affected large parts of Gondwanaland in particular and Laurasia to a lesser extent mainly in the period 0.9 to 0.5 billion years ago. This is one of the major crust-forming events of Earth history. It includes the first appearance of ophiolites and sheeted dikes, signifying that conventional sea-floor spreading was established at least by late Precambrian time. In the Gondwanaland, or southern, supercontinent, the Pan-African mobile belt system includes the Brazilide-Pharusian, West Congo, Mozambiquian, Katangan-Damaran, Adelaidean, and Transantarctican belts. In the Laurasian, or northern, supercontinent, the corresponding belt system, which includes the Appalachian-Cadedonian, Innuitian, Cordilleran, Hercynides, and Uralides, although initiated in late Proterozoic time, did not stabilize until the late Paleozoic (0.3 billion years ago) and even later. The Pan-African event probably represents a transitional tectonic regime between 1.1 and 0.6 billion years ago, during which both intraplate (ensialic) and plate margin (Wilson-type) orogeny may have operated synchronously in different places. Undoubted ophiolites about 0.8 billion years old, probably representing pieces of obducted oceanic crust, have been identified at Bou Azzer, Morocco (*24*), and at several localities in Egypt and Saudi Arabia (*25*). These, together with associated island arc-type calc-alkalic volcanic rocks, leave little doubt that plate tectonics processes broadly similar to modern processes were operating in the late Precambrian (*25*).

Metallogenesis. Archean metallogenesis is dominated by siderophile elements of the greenstone belt association. Chromium, nickel, and asbestos deposits are associated with mafic to ultramafic rocks, iron and manganese with sedimentary banded iron formations, gold lodes with a variety of rock types, and copper-zinc massive sulfide deposits with intermediate to felsic calc-alkalic volcanic sequences. Important deposits of tantalum, lithium, niobium, and beryllium are commonly associated with granitoids, especially those emplaced in late Archean time.

Important placer gold and uranium deposits lie in epicratonic basins of late Archean to early Proterozoic time. They are synchronous with or succeeded by the widespread banded iron formation accumulations that include the largest iron and manganese deposits of the world. Several layered mafic complexes, about 2.6 and 1.9 billion years old, respectively, contain the major nickel-copper and chromium-platinum resources of the world. This event marks a major shift from siderophile to predominant chalcophile metallogenic resources.

Shale-hosted lead-zinc (-copper) deposits are of major importance in mid-Proterozoic (1.8 to 1.4 billion years ago) basins and geosynclines, as are unconformity or paleosurface type uranium (-copper-nickel) deposits in interior basins of younger Proterozoic age in Canada and Australia. Major strata-bound copper deposits in the United States and central Africa, in late Proterozoic time, represent unusual metal accumulations in stable basins dominated by interior drainage systems. Large manganese deposits are present in sedimentary sequences of the same age. Titaniferous magnetite deposits lie in massif-type anorthosites. Important tin lodes are associated with granitoid intrusions.

In brief, the several distinctive metallogenic associations of the Precambrian and the resulting metallogenic trend have developed in response to the progressive growth and stabilization of the continental crust, the change from early mantle-dominated to later mantle-continental mixed activities, and the unidirectional change from prevailing anaerobic to aerobic ocean-atmosphere systems.

Proposed Tectonic Phases

A simplified three-phase tectonic trend is proposed, which followed the initial formative stage in earth development (Fig. 1), including core formation and major outgassing, together with presumed development of some primitive crust.

Microplate tectonics. The main tectonic control exerted was the thermal regime. According to the bulk Earth heat production curve (Fig. 1), the Archean heat flow in the interval 3.8 to 2.5 billion years ago was, in general, 2.5 to 4 times its present value. As a result, all lithosphere was comparatively buoyant (*26*); there was little if any subduction of the Benioff zone type, as in modern tectonics. Instead a pattern of small scale, yet vigorous mantle convection, may have resulted in an array of many small, jostling lithospheric plates. Initially these were constructed mainly of ultramafic to mafic volcanic rocks. Later, in response to sagging and partial melting of the lithosphere, more felsic material was added—leading, in due course, to development of sialic microplates. Aggregation of sialic material, including underplating, gave rise to protocontinents (small cratons) and, eventually, cratons of continental or near-continental stature by the end of Archean time.

Intraplate tectonics. Further cooling and consequent thickening of the lithosphere, now with substantial cratons, led to a condition of limited subduction so that the thinner, hence tectonically weaker, intracratonic zones became the focus of deformation. Deformation involved attenuation and even limited separation, with synchronous or subsequent horizontal shortening through basement reactivation and ensialic piling up of crustal wedges and nappes, the so-called ensialic mobile belt-type orogeny (*25*). There was no significant ocean-floor consumption, that is, Benioff-type subduction. During this activity, the larger cratons supported epicratonic basins of increasing dimension and complexity in pace with growing cratonic stability. The Proterozoic continental crust may have been aggregated into a few supercontinents.

Macroplate tectonics. Toward the end of Precambrian time, declining heat flow produced a lithosphere approaching present dimensions and compositions. The increase in negative buoyancy of oceanic lithosphere relative to continental lithosphere, coupled with the large size and comparative rigidity of the resulting plates, led to the modern plate tectonics process with Benioff zones and ocean-floor consumption at convergent plate margins. Cordilleran- and Appalachian-type geosynclines developed at leading and trailing continental margins, respectively, and at orogenic belts along plate collision sites.

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