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Problems of the Evolution of the Continental Crust

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Problems of the evolution of the continental crust

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In introducing this symposium I shall draw attention to some of the main problems concerning the early stages in the evolution of the continental crust.

First, we wish to know how early the continental crust began to form, how fast it accumulated, when and by what processes the granitic and granodioritic rocks which form so much of it separated from the mantle, and whether these granitic materials were added continuously to the crust or in sharply defined phases.

A few years ago it was thought that by studying $^{87}Sr/^{86}Sr$ initial ratios, it would be possible to discriminate between granitic rocks derived directly from the mantle and those derived from pre-existing crust. This method has shown that increments of granitic material are still being added to the crust from the mantle, as in Iceland (Moorbath & Walker 1965). But recycling of crustal material through the mantle by way of subduction zones, depletion of the deeper parts of the crust in rubidium and the smaller difference between mantle and crustal strontium isotope ratios early in the Earth's history make this method less decisive when it is applied to very old granites such as those which surround the greenstone belts of the Rhodesian craton and the Barberton mountain land. It has been proposed that the greenstones in these regions accumulated in an island arc environment on a thin crust (Anhaeusser, Mason, Viljoen & Viljoen 1969) and that most of the enveloping granites represent subsequent additions to the crust (White, Jakes & Christie 1971). Yet before 2550 Ma ago when the Great Dyke of Rhodesia was intruded, the granites which constitute much of the Rhodesian craton had already been emplaced and the crust there must have reached almost its present thickness. Structural and stratigraphic evidence suggest that much of the granitic material represents reactivated continental crust older than the rocks of the greenstone belts (Shackleton 1970). Isotopic studies of lead from the Tanzanian, Rhodesian and Canadian cratons indicate that a protocrust existed at least 4000 Ma ago in these regions (Robertson 1970).

Another method of investigating the rate of accumulation of the continental crust is to estimate its thickness at various times. This may be derived from determinations of the pressure–temperature conditions in which ancient metamorphic assemblages crystallized and from comparisons of the chemistry of ancient volcanic rocks with that of modern volcanics erupted through crust of known thickness. Metamorphic assemblages imply a crustal thickness of 25 km over 3000 Ma ago (Saggerson & Owen 1969), while the chemistry (alkali index) of volcanic rocks suggests a crustal thickness of 10 to 25 km and an island arc environment 2700 Ma ago (Condie & Potts 1969) or 15 km 2700 Ma ago from the trace element chemistry (Hart *et al.* 1970). Such estimates refer to specific local areas, and since crustal thickness must have varied it is to be expected that estimates differ. However, it is not justifiable to extrapolate from the present to the early Precambrian in interpreting the chemistry of volcanic rocks because this depends primarily upon depth of origin of the melt which in turn depends on the thermal gradient. Since this was much steeper early in the Earth's history, melts such as are now erupted where the crust is thin could then have come from beneath continental crust nearly as thick as it is now.

Because of the rapidity with which erosion reduces the land surface nearly to sea level, the thickness of the crust is itself controlled by the relative volumes of continental crust and oceans. In some of the old cratons in Africa, Canada and Australia, which have been undeformed for 2500 Ma, it is possible to estimate the amount by which the crust has been thickened since that time. This thickening must, in such areas, have been by underplating, the addition of material to the crust from below. Veneers of very old flat-lying sediments deposited near sea level remain on some of these cratons and their subsequent elevation indicates appreciable thickening of the crust by underplating (Shackleton 1970).

A second problem is to specify the thermal régimes in the early stages of the evolution of the crust. The rate of radiogenic heat production 2700 Ma ago was twice its present value and before then still greater. Although the heat-producing elements were not then so highly concentrated in the upper crust, the thermal gradients must have been much higher. In continental crust the gradient is now about 10 °C/km. The characteristic metamorphic association of the Archaean is the low-pressure facies series from greenschist through amphibolite to granulite which indicates thermal gradients in excess of 30 °C/km. Saggerson & Owen (1969) suggest a reduction of the geothermal gradient from about 75 to 10 °C/km in the last 3000 Ma. Kyanite, typical of the intermediate pressure facies series, is uncommon in rocks over 2000 Ma old and glaucophane, typical of the high pressure facies series which is developed at gradients less than 18 °C/km, is extremely rare. An association of glaucophane and pumpellyite in the Archaean Dharwar schists of the Nuggihalli belt in Mysore (Uadarajan 1968) appears to be unique. Diamonds, which reach the surface in regions where the thermal gradients in the crust and upper mantle are low, are increasingly restricted in distribution earlier in the Earth's history but are found in rocks more than 2000 Ma old in West Africa and South America (Knopf 1970). There are thus indications that despite a generally higher thermal gradient there were some regions where it was low.

An estimate of the geothermal gradient in the Archaean may also be derived in areas such as West Australia, from seismic refraction determinations of the depth below greenschists of the Conrad discontinuity which represents the upper surface of granulite facies rocks that have presumably survived since Archaean times. From thermal gradients it is possible to infer the thickness of the lithosphere. It is often suggested that this was thin early in the Precambrian but quantitative estimates are needed.

A third problem is the chemistry of the primitive crust. The presence of detrital uraninite in ancient fluvial deposits (Witwatersrand, Blind River) and the absence of continental red beds, suggests a lack of free oxygen in the atmosphere before about 2300 Ma ago. Early Precambrian volcanic rocks, which in a well-sampled area in Canada were found to consist of 58 % basalt, 18 % andesite, 11 % dacite, 9 % rhyodacite and 4 % rhyolite, a typical calcalkaline series (Goodwin 1968), were characterized by low K/Rb and K/Cs ratios, suggesting that the upper mantle source region of low-K tholeiites has become progressively depleted in Rb and Cs. Systematic sampling shows that the early crust had less K, Ti, U and Th and more Na, Cr, Ni and perhaps Au, than younger crust. Older swarms of basic dykes which cut continental crust have less K_2O and TiO_2 and slightly more Na_2O than younger ones. Thus the chemical evolution of the crust is becoming well established. It is also clear that deeper crustal levels, represented by rocks of granulite facies, were already at an early stage depleted in SiO_2, K_2O, H_2O, Th and U (Eade & Fahrig 1971).

A fourth problem is to determine the depositional environment in which the early Precambrian sediments accumulated. Sediments in the ancient greenstone belts have been interpreted as trench/arc associations, as miogeoclinal and as deposited in microgeosynclines. The oldest well-defined sedimentary basin for which there is detailed sedimentological and stratigraphic data is the 2500 Ma Witwatersrand basin (Brock & Pretorius 1964). A thickness of 5 km of mainly shallow-water sediments accumulated in the interior of this basin and in this case it is clearly established that the various formations thinned to feather edges near the present margins of the basin and that sediment transport was towards the basin. In the case of the older greenstone belts there is very little reliable evidence to show whether these generally synclinal belts represent original basins of deposition, although this is suggested by inconclusive evidence from the Slave structural province of Canada (McGlynn & Henderson 1970).

Finally there are the fascinating problems of the kinematics of the early Precambrian crust. Should we interpret the structures in terms of plate tectonic theory, modified to take account of a thinner crust and lithosphere and higher thermal energy or were other processes of crustal deformation then operative?

Palaeomagnetic studies are beginning to give polar wandering curves from the early Precambrian (McElhinny, Briden, Jones & Brock 1968). Intercontinental and intercratonic comparisons of these curves should soon indicate whether large-scale plate motions and consequent subduction zones were responsible for the volcanism deformation metamorphism and granite intrusions in the Archaean. At present it seems that no large intercratonic movements are indicated by the palaeomagnetic data (McElhinny *et al.* 1968; Briden, Piper, Henthorn & Rex 1971). Another way of studying these problems is to use the distribution patterns of specific and unusual rocks or minerals. Plots of the distribution of Precambrian anorthosites (Herz 1969), diamonds (Knopf 1970) banded ironstones and granulite facies rocks tend to show that when plotted on pre-Mesozoic drift reconstructions, these distributions form simple patterns which would be difficult to explain if the supercontinents had been assembled from numerous previously dispersed plates. The majority of Precambrian fold belts appear to have formed on pre-existing continental crust. Older structures can be traced from the cratons into the fold belts and occasionally through them and into the craton on the other side. There are two examples in central Africa where an orogenic belt crosses an older one obliquely without apparently offsetting it, which implies that there has been little, if any, crustal shortening in the younger orogenic belt and thus little, if any, relative motions of the plates on either side. The best examples are the crossing of the Irumide belt by the Zambesi belt (Cahen & Snelling 1966, fig. 17.4) and the crossing of the Ubendian–Ruzizi fold belt by the Kibaran. The sutures which should exist if Precambrian fold belts, such as those which separate and surround the Archaean cratons in Africa are to be interpreted as collision orogenies, appear to be lacking. The Precambrian orogenic belts in Africa surround subcircular cratons in a pattern which is difficult to reconcile with a plate-tectonic interpretation. Thus there are many reasons to doubt whether the plate-tectonic hypothesis is generally applicable to the early history of the crust. On the other hand, there is also evidence suggesting plate movements. The glaucophane–pumpellyite–epidote assemblage in Mysore (Uadarajan 1968) suggests a subduction zone; the great wrench faults which are found in many ancient Precambrian terrains should, by analogy with wrench faults now active, be transform faults. Geochemical data have been used to support an

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interpretation of Archaean greenstone belts in terms of sea-floor spreading and subduction zones (White *et al.* 1971). Palaeomagnetic evidence shows that large, rapid and irregular motions of the continental crust were occurring throughout Precambrian times. Many uncertainties concerning the kinematics of the primitive crust thus remain.

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