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Precambrian Thermal Regimes

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Precambrian thermal régimes

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Regional studies of Archaean and Proterozoic complexes provide the basis for one approach to the problem of reconstructing the thermal régimes of Precambrian eras. Such studies have a bearing both on the *PT* conditions under which metamorphism and deformation of continental crustal rocks took place and on the conditions of magma-generation at depth. Although several lines of evidence, including the wide distribution of low-pressure metamorphic facies series, are consistent with the inference that Archaean geothermal gradients tended to be steep, the characters of many complexes formed before 2700 Ma suggest *PT* ranges not far from those indicated by rocks which suffered deformation and metamorphism in Phanerozoic zones of high heat flow.

Tectonic patterns developed on a global scale in early Proterozoic times suggest that lateral variations of heat flow in continental crust did not conform to the present pattern, zones of high heat flow being developed within as well as at the borders of continental rafts. There are, nevertheless, indications that the lithosphere had locally attained thicknesses comparable with those found today in many cratonic regions soon after the end of the Archaean era.

1. THE GLOBAL CONTEXT

The probability that the heat flow to the outermost part of the Earth has declined over the past four billion years has coloured discussions of the early stages of earth history for many years. Since igneous activity, regional metamorphism and the responses of rocks to stress all depend on thermal controls, there is hardly an aspect of the geological record which is not potentially relevant in this context. The three dimensional distribution and style of the effects of regional metamorphism relate to thermal conditions in the crust and, when considered in relation to experimental data on mineral stabilities, can provide the basis on which rough estimates of crustal geotherms can be made. The extent and distribution of igneous activity and the geochemistry of igneous suites reflect, among other controls, the conditions of partial melting under which the parent magmas were generated and when combined with the results of experimental studies allow one to make rough estimates of temperatures in the upper mantle. Finally, the workings of successive tectonic systems on a global scale are, and presumably always have been, adjusted to temperature dependent factors such as variations in the thickness of lithospheric plates and in rock densities. Comparative studies of geological provinces formed over successive time intervals from the Archaean eras to the present day should therefore reveal qualitative indications of any long term variations in thermal régime which have resulted from the decline of radioactive heat production or from any decline in heat flow from the core or inner mantle.

Much of the value of such investigations depends on our ability to compare like with like. Observed heat flow values at the present Earth's surface range from about 0.5 h.f.u. in continental cratons to over 3.0 h.f.u. in island arcs and other active plate margins. The corresponding regionally maintained crustal geotherms range from below 10 °C/km to about 50 °C/km.

These gradients enclose a broad field on a depth–temperature plot and one of the points which will be suggested in this paper is that virtually the whole range of former continental geothermal gradients which are suggested by the evidence from Archaean and Proterozoic provinces falls within this field. The true picture of thermal conditions today can only be appreciated by taking into account the distribution and relative areas of zones of lower and

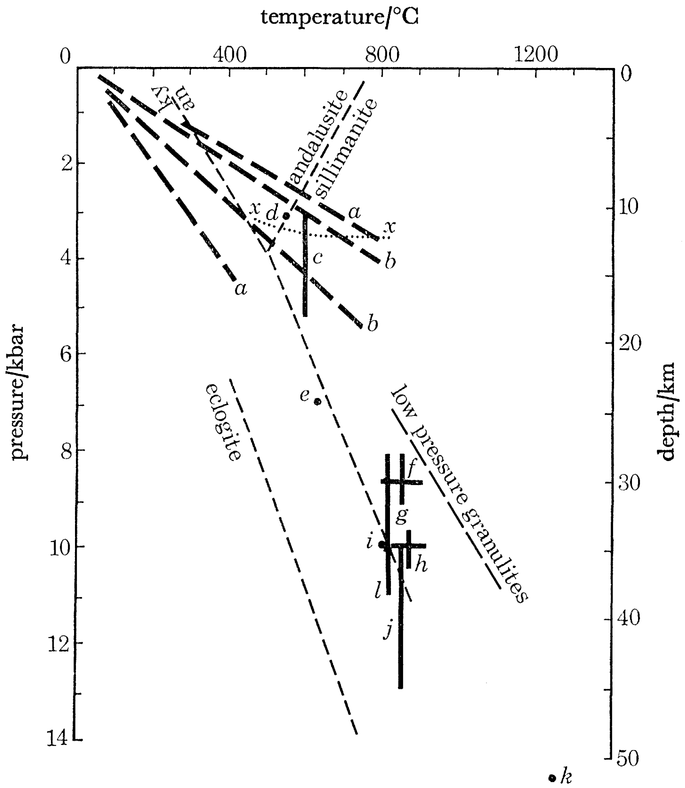

FIGURE 1. Schematic temperature–depth diagram showing published estimates of the *PT* environments of formation of Archaean metamorphic assemblages in relation to some experimentally determined stability fields. *x–x* marks possible *PT* evolution of a short-lived metamorphic environment in granite/greenstone belt provinces (p. 435). *a*, *a*, maximum and minimum gradients in Rhodesia (Saggerson & Owen 1976); *b*, *b*, maximum and minimum gradients in Wyoming (Condie 1976); *c*, Perseverence, W. Australia (Binns & Groves 1975; Martin & Allchurch 1976); *d*, Slave province (Ramsay 1973); *e*, *i*, SW Greenland (Wells 1976); *f*, *h*, Limpopo belt (Chinner & Sweatman 1968); *j*, South Harris (Wood 1975); *k*, Scourie, Scotland (O'Hara & Yarwood 1978); *l*, Scourie (Muecke 1969).

higher heat flow. High heat flows and steep crustal geotherms characterize constructive and destructive plate margins, while much lower average values are recorded over very large areas in the interiors of continental and oceanic plates. Evaluation of former thermal régimes depends not only on the estimation of gradients from individual localities but also on the correct assessment of the proportion of the crust for which such estimates may be representative. Some progress can be made along these lines for earlier Phanerozoic and Proterozoic terrains, since it is usually possible to distinguish crustal units which were relatively stable (and for which relatively low geothermal gradients may be inferred) from the contemporaneous mobile belts in which evidence of high heat flow is recorded. In Archaean terrains, where the stable/mobile distinction is blurred, we scarcely know even whether the geothermal gradient inferred from a sample locality should be regarded as falling near the minimum or the maximum limits of the range of Archaean variation. There is, of course, little direct evidence concerning heat flow in oceanic areas for any period prior to the Mesozoic era.

2. Archaean régimes (older than 2600 Ma)

Both structural and geochemical evidence seems consistent with the view that temperatures in the upper mantle were generally high during Archaean times. The structural argument rests on the absence of clearly defined stable crustal units of the kind which are readily recognizable in tectonic systems developed during or after the time band 2800–2500 Ma (p. 436). The apparent absence of continental cratons from the systems developed before 3000 Ma seems best explained by reference to an Archaean lithosphere too thin, and therefore too weak, to resist major stresses. If the base of the lithosphere represents an approximately isothermal surface, it would follow from the above considerations that the base of the Archaean lithosphere had a smaller vertical relief than the corresponding surface today, everywhere lying above the depths of *ca.* 100 km currently cited as minimum thicknesses of the lithosphere in stable regions. This arrangement might imply that heat flow at the surface of the continental crust was also more uniform than it is today, the probable minimum values being of the same order (2 h.f.u. and above) as those now characteristic of mobile regions (p. 431).

The relatively common occurrence of ultramafic extrusives and of igneous suites with komatiitic geochemistry in greenstone belts formed between 3300 and 2700 Ma suggests that large fractions of partial melt were mobilized from the source regions, possibly in very hot mantle plumes rising from depths of up to 200 km (Green 1972). The close time link observed between extrusion of the volcanic pile forming a greenstone belt and emplacement of the associated tonalitic intrusions suggests that each of the characteristically bimodal granite/greenstone-belt assemblages was formed rapidly in the aftermath of one such thermal disturbance of the mantle.

As Moorbath (1978) emphasizes, major crust-forming igneous events lasting 100–200 million years seem to have been separated by pauses of considerably longer duration. It would follow that the high mantle temperatures implied by the geochemistry of the igneous suites were anomalous in the context of the Archaean régime as a whole.

Indications of temperature variations in the crust are provided by metamorphic assemblages and structural styles in Archaean provinces (figure 1). In granite/greenstone belt provinces, where metamorphic grade ranges from low greenschist to amphibolite facies, low-pressure facies-series predominate, andalusite $\pm$ cordierite being recorded in the middle grades over, for example, much of the Kapvaal, Rhodesian, Wyoming and Slave provinces. This facies-series suggests a geothermal gradient of more than 30 °C/km if the experimentally determined limits of stability of the aluminium silicate minerals are accepted (cf. den Tex 1971). Saggerson & Owen, arguing from complex premises, infer that gradients in the Rhodesian granite/greenstone belt terrain ranged from about 30 °C/km to over 50 °C/km (1976, fig. 8), while Fyfe (1976) on theoretical grounds has envisaged Archaean crustal gradients of up to 100 °C/km. Data from mineral assemblages in granite/greenstone belt provinces of Wyoming (Condie 1976), Western Australia (Binns & Groves 1975; Martin & Allchurch 1976) and the Slave province of Canada (Ramsay 1973) fall within the *PT* field outlined by the limiting gradients of Saggerson & Owen (fig. 1): this field is essentially the same as that of the modern zone of high

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heat flow on the inner side of the Japanese volcanic arc (Uyeda 1972) and even the more extreme estimate of Fyfe (*ca.* 100 °C/km) is no higher than gradients which have been inferred for the Hercynian belt of Europe (cf. Zwart 1967) or which are observed today in Iceland.

As many writers have pointed out, geotherms of 40–50 °C/km should establish a region of partial melting of continental rocks at depths of no more than 15–20 km. Fyfe (1972) has emphasized the points, first that removal of the melt fraction in such environments might lead to the development of depleted residual granulites at comparatively shallow depths and, secondly, that the prevalence of melting should limit the thickness of the continental crust to a figure well below that characteristic of the present crust.

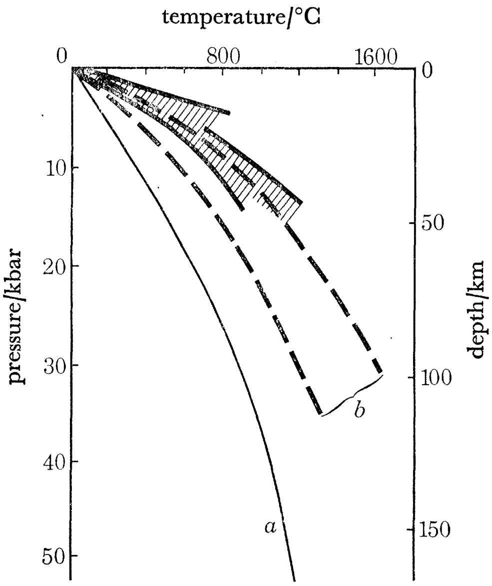

Figure 2. Temperature–depth diagram showing the fields of Archaean metamorphism inferred from figure 1 in relation to crust–mantle Archaean geotherms inferred from other evidence. The modern shield geotherm (*a*) is shown only for reference and does not imply the existence of low gradients in Archaean times; *b*, limiting Archaean geotherms (Green 1975).

In the light of these considerations, it is interesting to turn to the evidence derived from Archaean granulite and gneiss terrains (figure 1). Estimates of the *PT* environment in which these rocks were formed depend rather heavily on the interpretation of pyroxene–garnet relations and may be changed when this controversial matter is resolved. Present estimates based on samples from Scotland (Muecke 1969, cited by Lambert 1976; Wood 1975; O'Hara 1975, 1977), southwest Greenland (Wells 1976) and the Limpopo belt of southern Africa (Chinner & Sweatman 1968) fall consistently in a field bounded by geothermal gradients of 20 and 30 °C/km. None of these estimates comes within the field of low pressure granulites (characterized by the stability of olivine + plagioclase), nor does Clifford in a comprehensive study of African granulites cite any examples of low pressure Archaean granulites (1974). Eclogites, sometimes thought to be absent from Archaean provinces, are represented in gneisses probably, though not definitely, of Archaean age in the Lewisian complex of north-west Scotland (Alderman 1936) where they may have originated in the manner envisaged for the Norwegian eclogites investigated by Råheim & Green (1975).

Taking the experimental determinations at face value, it would appear that most known

Archaean granulites belong to medium to high pressure facies-series and that they originated at depths of more than 20 km in a continental crust which locally reached minimum thicknesses of 45 km (cf. O'Hara 1975). There is therefore a discrepancy between the characters of the major Archaean granulite terrains and those which might be expected of residual granulites formed in connection with partial melting at the base of granite/greenstone belt assemblages. The *PT* environments suggested by the mineralogical assemblages of the high-grade provinces are however, consistent with those expected from calculations based on the nature of Archaean basic magmatism by Green (geotherms III and IV in fig. 2, 1975) and by Lambert (1976).

The conditions of metamorphism inferred from assemblages in the granite/greenstone belt provinces fall well to the high temperature side of these estimates (figure 2). A number of lines of evidence suggest that these anomalous conditions were short-lived. The isograds tend to parallel the margins of tonalitic domes and show rapid variations of level, descending steeply in synformal greenstone belts and rising towards the granites (cf. Anhaeusser, Mason, Viljoen & Viljoen 1969). Signs of disequilibrium are provided in some regions by porphyroblasts of minerals developed during prograde metamorphism set in a finer grained and only partially reconstituted matrix (cf. Zwart 1967). These features point to the rising granites as a heat source; they suggest that the cooler greenstone belts subsided rapidly enough to distort the isograds and that when regional isotherms finally re-established themselves they did so at lower levels, leaving the irregular zonal pattern as a kind of thermal high tide mark.

The episodes of low pressure metamorphism therefore appear to record the effects of heat transfer by rising granites towards the end of the relatively brief Archaean crust-forming episodes (p. 433). The thermal gradient developed in relation to the rise of magma in such circumstances might be expected to have a curved form like that of line x on figure 1 and to diverge from any probable geotherm which can be extrapolated to depth. In the long pauses between crust-forming episodes the average geothermal gradient may have been closer to the 20–30 °C/km range suggested by evidence from the granulite and gneiss provinces.

The variations outlined above revive the often discussed question of the relations of low-grade and high-grade Archaean provinces. It is difficult to regard granulite provinces like those just discussed simply as the basal parts of the granite/greenstone belt provinces (cf. Windley & Bridgwater 1971; Glikson & Lambert 1976) if the *PT* assessments of Archaean rocks quoted are of the right order. The granulites could have been formed beneath low-grade provinces during the cooler intervals between crust-forming episodes, a possibility which does not seem to find much support from isotopic dating. An alternative possibility suggested by the fact that granulites tend to appear at the margins of granite/greenstone belt terrains (the Pikwitonei and NE Quebec granulites at the margin of the Superior province, the Limpopo belt between the Rhodesian and Kapvaal provinces, the Wheat Belt granulites at the margin of the Yilgarn block) is that lateral variations in heat flow were developed during crust-forming periods and that nuclei in which the greenstone belt sequence of events was going on were rimmed by zones in which the crust was relatively thick and the heat flow relatively low.

This possibility is envisaged by Condie (1976) for the Wyoming massif. Windley (1976) and Katz (1976) both regard high grade provinces as marking anomalous zones in the Archaean crust, Windley comparing them with modern continent margin structures like the cordilleran batholiths of western America and Katz likening them to transform structures. Both authors refer to them as zones of high heat flow whereas I suggest above that they may have been zones of average to low geothermal gradient in terms of the Archaean range of values.

The general features of Archaean granulite and gneiss provinces – the intimate interleaving of rock units of different parentage, the pervasive evidence of extreme ductile deformation, the frequency of layered anorthosites, the incorporation of supracrustal rocks in complexes subsequently depressed to great depths, and the indications that crustal thicknesses were great – are in keeping with the idea that these provinces mark specialized belts in the Archaean crust. The magnitude of the post-Archaean vertical movements required to bring the complexes to their present level of erosion suggests that their subsequent evolution followed a different course from that of those granite/greenstone belt provinces which have maintained a rather constant level in the Earth (p. 437). An initial buoyancy may have been imparted to the granulite belts by the fact that the thickness of crust was perhaps twice the Archaean average. The cause of their vertical mobility in later periods remains a major problem of Precambrian tectonics.

3. Proterozoic régimes (2600–1000 Ma)

The development of a system of continental cratons enclosed in a network of mobile belts which took place towards the end of the Archaean era marked the most important change in the working of the global tectonic system recorded in geological history. The first undoubted cratonic massifs were defined in all continents within a transitional time-band of a few hundred million years. Cratonic cover-successions showing only minor effects of subsequent disturbance and dating back to 2700–2300 Ma attest the tectonic stability of the units. Alkaline and kimberlitic igneous activity of the type normally associated with the existence of a thick lithosphere had begun at or before 2000 Ma and the occurrence of (detrital?) diamonds in part of the Birrimian formation of West Africa dated at *ca.* 2000 Ma indicates that cooling of the mantle had locally brought the base of the lithosphere within the stability field of diamond. These features, taken together, suggest that in early Proterozoic times the lithosphere beneath the cratons was at least 100–150 km in thickness and imply a cratonic geotherm not very different from that of about 10 °C/km commonly recorded in modern cratons. I know of no evidence which either favours or excludes the possibility that the lithosphere had reached thicknesses of over 200 km. The proportion of the earliest (> 1800 Ma) cratons to the mobile continental regions characterized by early Proterozoic tectonic patterns seems, from subjective estimates, to be a good deal lower than the craton: mobile belt proportions of later times although the subsequent increase in area of the cratonic units has been far from uniform (Sutton 1963).

The world-wide development of thick lithospheric slabs is usually seen as a consequence of cooling of the upper mantle. The rather short time span over which this change took place may indicate that a gradual decline of temperatures coincided with another change at depth. Strong & Stevens (1974) suggest that progressive outgassing of the mantle and consequent reduction of P_{H_2O} had simultaneously raised the freezing temperature of peridotite, so that thickening of the lithosphere was accelerated as the descending isotherms crossed the peridotite solidus.

The changes set in train towards the end of the Archaean era led to an increase in vertical relief of the base of the lithosphere from the maximum of 50–60 km possible in the generally thin early lithosphere to a figure perhaps two to three times this amount. Assuming that the ductility of the asthenospheric rocks exceeded that of the lithosphere, major irregularities of the interface would be expected to deflect the stress trajectories set up during bodily movements of each lithospheric slab and hence to accentuate contrasts in the tectonic responses of craton and mobile belt. At the crustal level these contrasts, in fact, commonly led to the development of deep shear zones or thrust belts at the margins of the cratons.

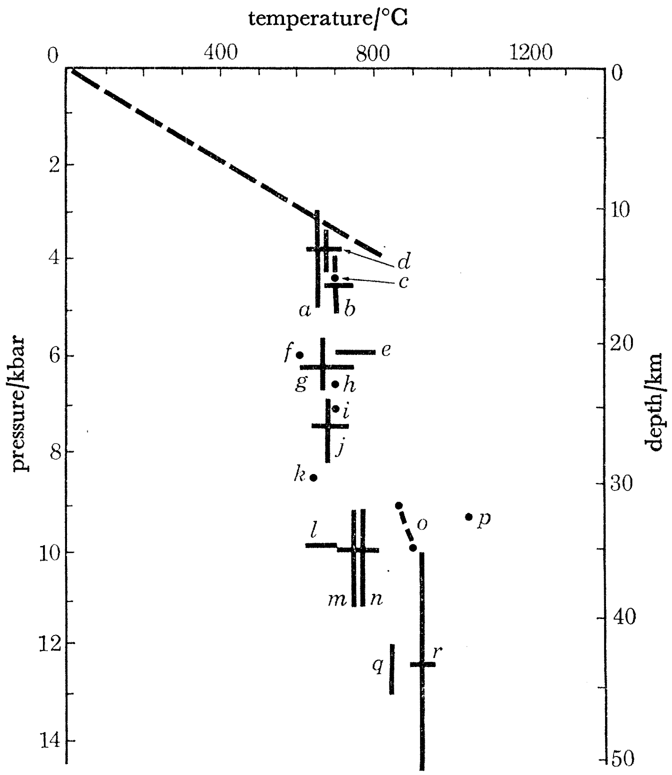

FIGURE 3. Published estimates of *PT* environments of formation of Proterozoic metamorphic assemblages (cf. figure 1). Most of the assemblages date from 2600–1000 Ma, a few are of uncertain age, greater than 600 Ma. *a*, Front Range (Gable & Sims 1970); *b*, *e*, Ceylon (Katz 1972); *c*, Grenville (Wynne-Edwards 1976); *d*, NW Scotland (Dickinson & Watson 1976); *f*, Snow Lake, Canada (Scott 1975); *g*, Opinicon Lake, Ontario (Currie 1971); *h*, *i*, *q*, eclogites of Norway, Poland (Råheim & Green 1975); *j*, dyke swarm NW Scotland (Dearnley, see Dickinson & Watson 1976); *k*, *n*, *r*, Lofoten-Vesteralen, Griffin *et al.*, in preparation); *l*, Cabo Ortegal, Spain (den Tex, Engels & Vogel 1972); *m*, Ivrea zone (Schmid & Wood 1976); *o*, *p*, Jotunheim (Griffin 1971; Battey & McRitchie 1975); broken line, Amazon belt (Leonardos & Fyfe 1974).

A second consequence of the cooling and thickening of lithospheric slabs was likely to be an increase in density of the mantle rocks which solidified during the process. Crough & Thompson (1976) calculate that, provided crustal thickness remained unchanged, a 40 km increase in lithospheric thickness would lead to a 1 km reduction of surface elevation due to the resulting isostatic adjustment. A. J. Baer has pointed out to me that such negative movements towards the end of the process of cratonic stabilization may help to account for the fact that many granite/greenstone belt provinces stabilized at about 2600 Ma seem to have suffered very little further elevation once the initial period of rapid erosion was complete (Watson 1976). Processes such as crustal shortening or underplating which thickened the crust and tended to increase its buoyancy might be partly counteracted by this mechanism.

The range of metamorphic crustal environments suggested by studies of mineral reactions and assemblages from Proterozoic mobile belts differs very little from that inferred for Archaean terrains and published estimates again fall within the known range of PT variation in the present crust (figure 3). Low-pressure metamorphic facies series are developed on a large scale in such classic terrains as the Michigan region described by James (1954) and the Svecofennide belt. Granitic and/or migmatitic rocks are abundant in these terrains and their relations suggest that the zones of metamorphism were established during or soon after periods of granite emplacement. Unlike their Archaean counterparts, several of the low pressure facies series (for example, that of the Grenville belt) include low pressure granulites and/or intrusive charnockitic suites, a feature which may be related to low partial pressures of water in the deeper parts of the series. Assemblages belonging to medium- or high-pressure facies series are represented perhaps more widely than in Archaean terrains and include a variety of schists, gneisses, granulites and eclogites: estimates of the conditions of formation of the high-grade members of these series coincide closely with those of the Archaean complexes already discussed. Their PT field is roughly limited by geotherms of 20 and 30 °C/km and, once again, shows a misfit with the field of the low pressure series which is bounded on the high-temperature side by a gradient of 60 °C/km inferred in the Amazon belt of Brazil by Leonardos & Fyfe (1974). From the field relations with granitic rocks mentioned above, I suggest that gradients in the low pressure terrains were distorted by the effects of heat transfer by rising granites in the manner discussed on page 435.

The relations of provinces of low-pressure and higher-pressure facies series in space and time have so far been little discussed. Katz (1974) regards a broad tract crossing Gondwanaland as being occupied by paired metamorphic belts developing from Archaean into Proterozoic times and relates the initiation of the tract to transform motions. Progressive adjustments in the metamorphic state of the Lewisian granulites and gneisses of northwest Scotland suggest that after being held continuously in conditions of medium- to high-pressure metamorphism for some hundreds of millions of years from 2800 Ma onward, the complex suffered simultaneous elevation and steepening of the geothermal gradient at or before 1800 Ma (Dickinson & Watson 1976; cf. O'Hara 1977); both processes may have been connected with the influx of heat borne by granites (Bowes & Hopgood 1973).

Evidence of the kind represented in figures 1 and 3 suggests that the metamorphic complexes of Archaean and Proterozoic age which can be examined today were developed within roughly the same range of geothermal gradients. A fundamental contrast between the two time-periods is provided by the absence of continental cratons from the Archaean system (p. 433). Whereas the whole of the Archaean continental crust might be expected to have heat flows and geothermal gradients similar to those characteristic of the present zones of high heat flow, a substantial part of the Proterozoic crust had the low heat flow and geothermal gradients of the order of 10 °C/km which characterize the cratons. It would no doubt be possible to test mathematically the adequacy of the uniformly moderate to high Archaean heat flow to dispose of the large output of radiogenic heat to be expected before 2600 Ma (see O'Nions, Evensen, Hamilton & Carter 1978). Any major deficiency of the mechanism would suggest that the Archaean crust – continental or oceanic – included regions of exceedingly high heat flow and geothermal gradients which cannot be matched today. Since no evidence of unique thermal conditions in the Archaean crust is on record, one might conclude that very high temperature zones were systematically destroyed in Archaean times and that the Archaean provinces which survive represent only the cooler parts of the crust of that era.

Taken as a whole, the metamorphic series so far discussed differ from late Proterozoic–

Phanerozoic complexes in the absence of representatives of the blueschist–eclogite facies series which seems to enter the geological record soon after 1000 Ma. Although it is customary to attribute this contrast to the effects of declining thermal gradients at destructive plate boundaries (Ernst 1972), the control may be indirect. As England & Richardson have pointed out (1977) the preservation of the blueschist–eclogite suite depends on tectonic processes leading to the rapid retrieval of rocks depressed to great depths and the incoming of the suite could coincide with a change in tectonic system. The initiation of the modern style of plate tectonics in which an essential feature is the return to depth of oceanic crustal slabs over-ridden at Benioff zones is linked by some writers, e.g. Ringwood and Green, with the onset of thermal conditions favouring for the first time the development of dense eclogite in the downgoing slabs.

A second contrast between Proterozoic tectonic systems of the period 2600–1000 Ma and the modern system concerns thermal régimes in the interior parts of large continental rafts. Under the present system, such rafts are regarded as essentially rigid and move as units with respect to the magnetic poles. The large Proterozoic continents, on the other hand, are thought by some workers to have suffered extensive internal distortion during their bodily migrations (Briden 1973; Sutton & Watson 1974). The internal distortion was concentrated in mobile belts of various kinds which are shown, by the regional occurrence of effects of metamorphism, penetrative deformation and granite emplacement, to have been zones of high heat flow. Ensialic mobile belts have no precise analogues in the present tectonic system and the origin of the thermal anomalies which they record remains to be established. Shackleton (1969) has expressed the view that most mobile belts were formed above rising (i.e. hot) mantle currents and Hurley (1973) has envisaged a situation in which metamorphism, vertical movements and granite emplacement took place in the heated crust above a rising current without rupture of the crust and consequently without sea floor spreading. Such processes, involving stretching of a not-entirely-rigid continental plate could account for the characters of some ensialic mobile felts (Watson 1976; Wynne-Edwards 1976), while others may represent broad ductile shear zones (cf. Sutton & Watson 1974; Katz 1976).

The apparent decrease in the importance of ensialic mobile belts from late Proterozoic to Palaeozoic times and the unique character of the anorogenic anorthosite-rapakivi intrusive suite of the period 1800–1900 Ma, which appears to record a thermal event not repeated in later times (Bridgwater & Windley 1973; Bridgwater, Sutton & Watterson 1974) may provide a record of the gradual decline of temperatures beneath the continental interiors and the consequent thickening and strengthening of the lithospheric plates.

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