Dominance of Horizontal Movements, Arc and Microcontinental Collisions During the Later Permobile Regime

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Abstract

In the latter part of the permobile regime ($\simeq 3{\cdot}5$–$2{\cdot}5$ b.y. ago), at a time when many of the oldest rocks now preserved were formed, most of the present continental mass had differentiated and the Earth's surface can be envisaged as rather less than one-third covered with continental material accreted by arc-amalgamation but only locally differentiated into granitic and granulitic fractions by Tibetan-type continent–continent collision processes. The continental masses were not significantly thinner than today's continents but were up to a hundred times smaller in area and hence much more numerous.

The remaining two-thirds of the Earth's surface was oceanic and the amount of water in the oceans about the same as today. Because heat-generation rates were perhaps three times higher than now, thermal gradients were rather steeper than at present and additional thermal energy was dissipated through greater ridge activity. Ridges either spread faster or the total length of ridge was longer than today or, more probably, both. The total length of subduction zone now roughly matches the total ridge length and this was presumably also the case 3 b.y. ago.

The lenticular style of the Superior Province is the characteristic signature of the rapid horizontal movements and consequent arc and microcontinental collisions of this time. Greenstone terrains such as the Superior Province preserve few obvious signs of the torsionally rigid behaviour at rupture and collision which became general during the Proterozoic. For this reason we characterize the deformation, although it involved sea-floor spreading, subduction and continental drift, as permobile deformation and not as plate tectonics.

Collisional processes affecting fossiliferous strata of Phanerozoic times generally disrupt the ophiolite, island arc, continental margin and exogeosynclinical successions involved and collision zones such as the Alps and Himalayas are recognized as areas of great tectonic complexity. Students of greenstone belts, who do their stratigraphy without fossil control, commonly report undisturbed successions. It may be that there are unrecognized tectonic boundaries within greenstone belts, recognition of which would help to resolve some of the supposed differences between these terrains and arc rocks involved in later collisions.

Introduction

Although, in the early history of the Earth, fractionation of anorthosite and dominance of structural evolution by impact and related eruption of basalt are likely to have characterized the surface development of the planet, no material produced in this phase appears to have survived in a recognizable state. By the end of the permobile regime (Burke and Dewey, 1973) conditions were much closer to those of today. We here suggest that these conditions imply a dominance of horizontal motion of a rigid lithosphere (boundary conduction layer in the case of the oceans) with, as now, rates and amounts of horizontal motion exceeding those of vertical motion by one hundred to one thousand times.

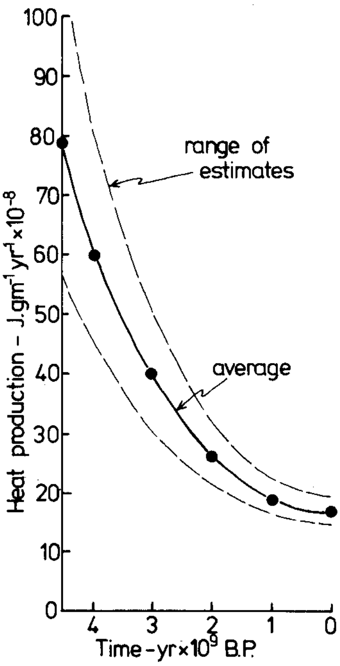

Fig. 1. Total heat production through geological time by U, Th, and K. (After Lee, 1967)

The Permobile Ocean: Faster Mantle Fractionation

Jakeš (1973), among others, has shown that the average composition of continental crust resembles that of island arcs and has suggested that continents have been made by a two-stage fractionation of mantle material, the first stage producing ocean floor igneous rocks at ridges and the second island arc igneous rocks above sinking inclined slabs of oceanic lithosphere. Continents form from arcs by collision following horizontal motion. In the first stage partial melting of mantle material to produce basalt and leave depleted pyrolite involves only vertical movement but once these rocks have been differentiated and emplaced horizontal motion comes into operation as the new lithosphere is carried away from the spreading ridge. Cooling and thickening of ageing lithosphere in this process is important in the convective dissipation of heat generated within the Earth. Because the thickness and elevation of cooling lithosphere change as the square root of the lithosphere's age (Parker and Oldenburg, 1973) more heat is dissipated by faster spreading ridges than by slow.

Fig. 1 shows that much more heat was being generated by radioactive decay in the Earth 2·5 b.y. ago than now. Many authors (for example, Goodwin, 1973) have emphasized the close resemblance of the rocks formed in the latter part of the permobile regime and preserved in greenstone belts to those formed at ridges and above sinking slabs today. It is evident (Jakeš, 1973) from this resemblance that the two-stage fractionation of mantle material at ridges and above descending slabs was then operating much as today. Because two or three times as much heat was being generated and because there is no evidence of the operation of any heat-dissipating process peculiar to ancient times, we infer that ridge processes were operating two or three times as fast as today. This greater intensity of operation could have been achieved either by greater average spreading rates or by greater total ridge length, or by both. Assuming that the volume of the Earth was about the same as now, there would have been either a greater length of subduction zone to consume the extra ocean floor or subduction would have worked faster. Since there is some evidence (see, for example, Forsythe and Uyeda, in press) that there is a general average for subduction rates at about 6 cm/yr (w.r.t. one of the slowly moving reference frames), a substantially greater length of subduction zone seems more likely.

The picture that emerges is that of an ocean in late permobile time with a great length of fast spreading ridge and a large number of long trenches, the two processes producing at a great rate ocean floor and arc rocks very like those of today. There seems little likelihood that oceanic thermal gradients were much greater in those times. Partial melting of pyrolite to make basalt presumably took place at temperatures and pressures similar to those of today and the greater amount of heat coming from within the Earth was mainly dissipated by the faster convective removal of basalt and depleted pyrolite at the faster spreading ridges. Ultramafic lavas appear to be rather more common in the Archaean than in later rocks and their occurrence perhaps indicates local melting at higher temperatures of the ancient pyrolite. Since even in the Archaean the proportion of these ultramafic lavas within basaltic piles is low, it seems unlikely that thermal gradients at ridges were much steeper than now. We suggest, as did Moores (1973), that Archaean oceanic crust is likely to have been somewhat thicker than present oceanic crust, both because the mantle was then likely to have been a little less depleted, and because a slightly higher percentage of partial melting occurred due to slightly higher thermal gradients. In contrast, the Archaean oceanic lithosphere (and continental lithosphere) is likely to have been rather thinner than at present.

Because all oceans soon 'self-destruct' (Atwater's term), it is impossible to make direct observation of extensive areas of old oceanic material and inferences about ancient ocean can only be made indirectly or from the dismembered slivers of oceanic and arc rocks to be found in greenstone belts. In contrast, continents are 'non-geodegradable' (Dietz's term) and once made, persist, although subject to episodic internal fractionation (Dewey and Burke, 1973) due to continent–continent collision. For this reason we do not consider here the idea that the Earth's surface was once covered with a greater area of continent.

Ancient Continents

Inspection of Fig. 1 shows that very roughly twice as much heat was generated by radioactive decay between 4·5 and 2·5 b.y. ago as has been generated since. If this heat was the sole source of formation of the continents through two-stage fractionation and continental accretion through arc collision, then about two-thirds of the continental mass would have been produced by 2·5 b.y. ago. However, the Earth was already inhomogeneous at planetary accretion and additional heat was generated in its early history from short-lived radioisotopes, and from gravitational potential energy during accretion and subsequent core formation. Major impacts must have provided additional energy, at least partly dissipated through mantle fractionation, over a relatively long time range, mainly between 4·5 and 3·9 b.y. ago if terrestrial history paralleled that of the Moon. Because these processes made available an unknown, but probably large, amount of additional heat to promote mantle fractionation, it seems likely that more than two-thirds of the present continental mass was produced under the permobile regime and that by its end ($\cong$2·5 b.y. ago) much, if not nearly all, of the present continental mass was in existence.

Other lines of evidence supporting this view include: (1) the increasingly widespread recognition of very old continental rocks (Hart and Goldich, 1975), (2) the arguments of Armstrong (1968) and Wise (1974) indicating that much of the present oceanic water volume was present by 2·5 b.y. ago and that continents stood then with their present freeboard and (3) evidence of recycling of strontium isotopes (Armstrong, 1968).

Radioactive isotopes concentrated in the continents at the end of the permobile phase were themselves generating approximately twice as much heat as now and for this reason continental thermal behaviour must have been somewhat different. However, there appears no strong evidence of persistence of extreme thermal gradients or very different tectonic behaviour within the ancient continental masses because there is little sign of widespread partial melting, other than that due to continent–continent collision.

The only geological evidence that clearly shows anomalous behaviour of continental lithosphere prior to about 2·5 b.y. ago is the extensive and very thick shelf sediment sequence of the Witwatersrand Triad (2·8–2·4 b.y.). This includes a huge thickness of silicic and mafic volcanics (Whiteside, 1970) also not like anything seen in younger shelf sequences. The Huronian (2·4–2·2 b.y.) is of more normal thickness and lithology over most of its extent, except in its southern-most part (Frarey and Roscoe, 1970). This southern area is bounded by faults (Card, 1970) and there are minor amounts of basalt confined to the base of the very thick (>10 km) sequence of shallow water sediments. It is possible that this part of the Huronian occupies an east–west trending aulacogen, opening eastward into the more highly deformed and metamorphosed turbidite-bearing sequence of the 'Southern Province,' which perhaps represents the continental slope/rise, outer shelf and underlying graben facies deposits of a rifted continental margin.

Partial melting of continental crust averaging a diorite–granodiorite or andesitic composition causes differentiation into a residual anhydrous anorthosite and granulite rich fraction and a hydrous granitic fraction (as was shown experimentally by Green (1969), and as is well displayed on a relatively small scale in such hot-spots where minor amounts of continental crust have melted; e.g. the White Mountains and the Jos plateau). Continental crust, in which neither the high level granitic nor the deep seated granulitic and anorthositic rocks are abundant and in which there is a marked preponderance of granodioritic (andesitic) bulk compositions, is unlikely to have been partially melted and hence is unlikely to have been subjected to exceptional thermal gradients. Although both granulitic and anorthositic rocks are known among the rocks formed in premobile times, they are restricted in their distribution. We do not know of any 'granitic' Archaean terrain, including those with granulites and anorthosites, whose properties (e.g. Bridgwater et al., 1974) do not resemble in most major aspects those of younger terrains of basement reactivation, which are clearly associated with major continental collisions. The preponderance of granodioritic compositions among the continental rocks formed in late premobile times (the greenstone–granodiorite terrains) we take as evidence that there has been relatively little partial melting and internal fractionation after arc accretion except in areas that have suffered post-premobile time collisions.

Although a great proportion of existing continent appears to have formed by the end of permobile times, there are no very large areas of old continent preserved intact. This is because there was, in permobile time, a great total length of plate boundary so that any piece of continent stood a large risk of being fragmented by development of new plate boundaries. Continued convergence after collision, the process that produces the Tibetan and Central Asian environments with accompanying *in situ* continental fractionation (Dewey and Burke, 1973; Molnar and Tapponier, in press), was relatively uncommon in permobile times perhaps for a similar reason. As there was a great total length of plate boundary, motion may have been more readily transferred to other boundaries after collision. Also, break-up by rifting and opening of oceans and Tibetan reactivation following continental collisions since 2·5 b.y. ago must have fragmented, and modified beyond recognition respectively, a significant proportion of continental crust originally formed in the Archaean.

Archean Tectonics—The Style of the Superior Province

General Considerations

A basaltic dyke swarm not younger than 3 b.y. and perhaps as old as 3·7 b.y. cuts the oldest rocks yet recognized on earth (McGregor, 1973). This fact clearly shows that the upper lithosphere of that time behaved *instantaneously* in a brittle manner under stresses similar to those responsible for younger dyke swarms of which most in the Proterozoic and Phanerozoic are directly connected to rifting and the opening of oceans.

A general and sustained surface temperature of less than 100°C since ~~0~~3·4 b.y. ago is necessitated by the presence of thick sections of that and younger ages containing pillow lavas. This observation, combined with reasonable estimates of heat generation (Fig. 1) and a likely distribution of the heat-producing elements also requires that there was a rigid lithosphere of significant thickness (~20–40 km) by 3·4 b.y. ago. Because mantle heat production was at least twice the present rate at the end of the permobile regime, strain rates due to first-order lithospheric tectonics cannot have been less than at present. Therefore such tectonics must have operated within the brittle regime from the top of the lithosphere to a depth perhaps slightly less than, but essentially the same as, that today. We therefore infer: (1) that a rigid lithosphere (boundary conduction layer in the case of the oceans) has been present since at least 3·7 b.y. ago; (2) that lithospheric relative movement has been by ocean-floor spreading, transform faulting, and subduction at least since this time, even though there was probably a significantly greater length of plate boundary, faster average relative plate motion, and many small 'plates' prior to 2·5 b.y. ago; and (3) that continental fragments behaved in a similar manner to those of later times, and that the plastic compressive deformations seen within them are due to orogenic events *localized* in time and space and due to subduction and/or collision-related processes, which may involve the *local* and *temporary* occurrence of greatly elevated thermal gradients and silicic magmas. Sequences dominantly composed of mafic rocks over 10 km thick all in low greenschist metamorphic facies (e.g. Glikson, 1972), whether composed of several thrust sheets or not, clearly demonstrate that equilibrium thermal gradients at least in these areas were not generally extreme in Archean times (not above 25°C/km).

Alkaline volcanic and/or plutonic rocks seem to be absent from rocks formed prior to ~~0~~2·5 b.y. ago (with one exception noted below). This absence of alkaline magmas indicates that in environments now occupied by alkalic magmas (hot spots with or without rifting/spreading, in both continents and oceans), the percentage of mantle partial melt was relatively high (>10%) and perhaps that this melting took place at shallower depths (less than 40 km) and may suggest that a thinner lithosphere was present prior to 2·5 b.y. ago.

No aulacogens (failed rift arms) have been recognized prior to about 2·2 b.y. ago (2·4 b.y. ago if the southernmost part of the Huronian is in an aulacogen). We suggest that the faster pace of plate motion and the thinner lithosphere prior to 2·5 b.y. ago resulted in very few failures of rifts to evolve into oceans, and therefore that there are no Archean aulacogens preserved in the small sample of essentially unmodified Archean crust remaining for inspection. Two types of Archean crustal terrain seem to be present. One is a 'granitic,' 'gneissic' or 'continental' terrain, and which has been most thoroughly studied in West Greenland (Bridgwater, McGregor and Myers, 1974). The other kind is the greenstone–granodiorite, or 'oceanic' terrain. We consider the Superior Province the best example of such a terrain because its area greatly exceeds that of other greenstone–granodiorite terrains and because it has been so thoroughly studied. It is evident that the tectonic environment responsible for greenstone–granodiorite terrains such as the Superior must be compatible with plate tectonics because the age of the Birrimian greenstone–granodiorite terrain of West Africa (~2·1–1·8 b.y.) is the same age and younger than the age of opening (~2·15 b.y.) of the ocean now represented by the Labrador Trough–Circum–Ungava–Nelson Front suture, which gives clear evidence of the operation of plate tectonics from this time onward (Burke and Dewey, 1972, 1973).

Island-arc Complex Assemblages and Tectonics Compared with Greenstone–Granodiorite Terrains

Several authors (e.g. Engel, 1968; Folinsbee et al., 1968; Anhaeusser et al., 1968; Jakeš and Gill, 1970; Burke and Dewey, 1972) have put forward arguments suggesting that the granodiorite–greenstone terrains represent island-arc rocks. Present-day island arcs and related surroundings contain five significantly different tectonic environments that all might be represented somewhere in the granodiorite–greenstone terrains if, as is likely, the analogy is valid. These environments are (Fig. 2A): (a) remnant arc; (b) marginal basin, including oceanic crust formed by back-arc spreading (Karig, 1971); (c) island arc; (d) arc/trench gap; and (e) trench mélange. Environments (a), (c) and (d) include oceanic crust on which the arc was built (Fig. 2B), and (e) includes pieces of oceanic crust and overlying pelagic sediment in thrust slices and as tectonic blocks in mélange. Upon arc–arc or arc–microcontinent collision the arc/trench gap and trench mélange are extremely likely, judging by their rarity in Phanerozoic orogenic belts, to be destroyed by being telescoped, sutured out, overthrust and eroded, perhaps because they are founded on essentially unmodified oceanic crust, which is readily subducted and/or obducted. Arcs involved in collisions tend to be uplifted, the calc–alkaline volcanic and volcaniclastic pile being eroded off and exposing the granodioritic plutonic terrain underneath. An example of this behaviour includes the Ladakh 'granite' (mostly granodiorite) in the Himalayas, which represents the roots of a volcanic arc that collided with the Asian continent in the Cretaceous. It is surrounded on both sides by narrow belts of ophiolitic rocks, including on the southern side 'flysch' derived from the arc and its erosion (Gansser, 1964). Another example is the Mesozoic batholithic belt of southern Chile, where the only evidence of the previously overlying volcanic rocks is now found in volcaniclastic sediments overlying the ophiolite complex floor of a Cretaceous marginal basin to the east (Dalziel et al., 1974). This marginal basin opened by spreading behind the arc and closed by collision of the arc with South America, deforming the oceanic crust and overlying sediments in upright folds and steep tectonic slide zones (Dalziel et al., 1974), and forming a greenstone belt (Tarney, Dalziel, and de Wit, this volume). Even without collisions, volcanic arcs tend to end up being represented by their plutonic roots, for example the Sierra Nevada batholith complex of California. These examples illustrate that it is more likely that the volcanics and volcaniclastics preserved in greenstone belts are, where they are narrow anastomosing belts separated from one another by large areas of granodioritic material (such as in the Kenora area of the Superior Province), more likely to represent marginal basins, and possibly arc/trench gaps and 'main' ocean sutures than they are to represent the island arc itself. However, where most of the outcrop area is greenstone with little granodiorite, as in the Abitibi area of the Superior Province, the arc volcanics, as might be predicted, seem to be well preserved (Goodwin and Ridler, 1970). It must be emphasized that significant quantities of marginal basin and arc/trench gap rocks are also likely to be present in this and similar areas.

Although the bulk of lava flows in any one marginal basin–arc–arc/trench gap complex may be largely confined to the arc, and oceanic crust of various ages, they are not wholly restricted to the arc. Arc/trench gaps at any one time are defined by the trench and by the volcanic front which is controlled by the depth to the descending slab (Fig. 2A). If, during a time interval, the dip of the descending slab increases or decreases, volcanism will occur in what was the arc/trench gap or marginal basin, respectively. Therefore, in an area representing an arc complex built over a significant time period the distinction between arc and arc/trench gap and between arc and marginal basin may not be clear-cut. Significant amounts of basaltic pillow lava are known within the mafic volcaniclastic sediment sequences overlying the oceanic crust of two originally very narrow (<50 km) 'fossil' marginal basins in western Newfoundland (Upadhyay et al., 1971; Kidd, 1974). These may represent a temporary shallowing of the

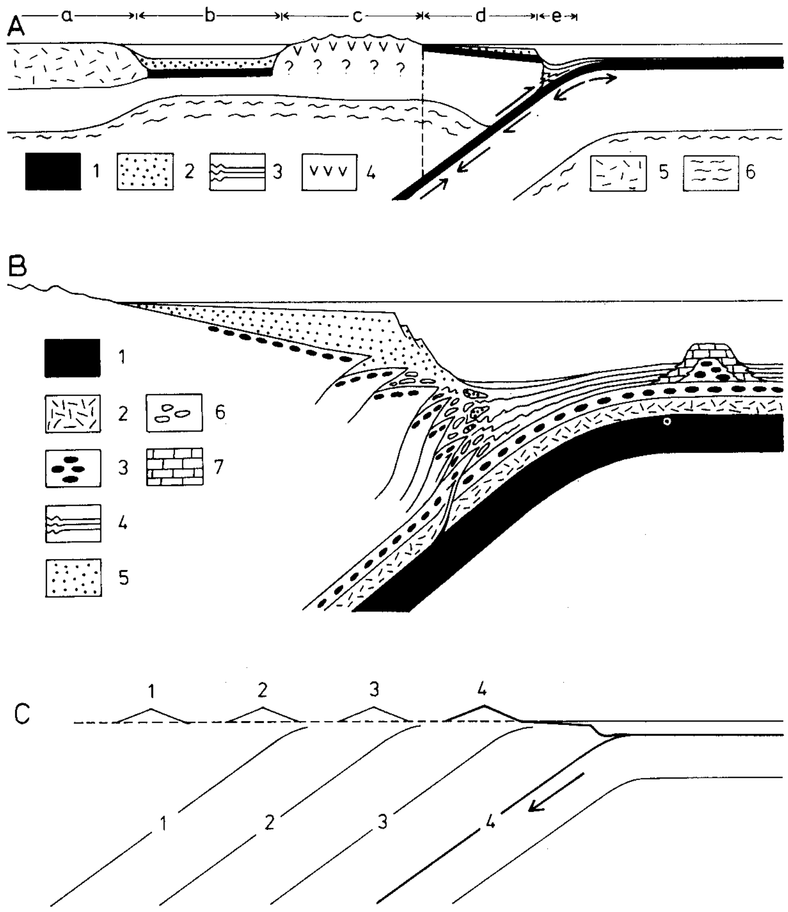

Fig. 2. A. Section across convergent plate margin showing (left to right): (a) remnant arc of microcontinent; (b) marginal basin; (c) volcanic arc; (d) arc/trench gap; (e) trench; together with oceanic lithosphere and subduction zone. 1—oceanic crust; 2—volcaniclastic sediments; 3—pelagic sediments; 4—arc volcanics; 5—remnant arc or microcontinental crust; 6—asthenosphere

B. Arc/trench gap-trench environment. 1—oceanic mantle (depleted harzburgite-dunite); 2—gabbro; 3—dykes and pillow lava; 4—pelagic sediments; 5—volcaniclastic sediments; 6—trench mélange; 7—carbonate platform and detritus on seamount

C. Proposed model for evolution of parts of convergent plate margins (e.g., Klamath Mountains, California). Previous sites of subduction 1, 2, 3 relate to previous arcs 1, 2, 3. Active arc 4 due to active subduction zone 4. Sites of subducted zones 1, 2, and 3 are not occupied by slabs after subduction moves to younger site

dip of a subducted slab, or perhaps some other process peculiar to marginal basins. It is important that the ophiolite complex pillow lavas (oceanic crust) of these two 'fossil' marginal basins, but not the lavas in the sediments above, appear to be basaltic komatiites (Strong, in prep.; Kidd, 1974).

If, in addition to these processes and the opening and closing of marginal basins, arc complexes are constructed with continuous or episodic oceanward migration of the subduction zone, such as has been suggested for the Klamath Mountains in California (Fig. 2C), horrendous complexities of tectonic and magmatic chronology are likely, quite apart from additional complexities superimposed during arc–arc or arc–microcontinent collisions.

As well as granodioritic plutonic complexes representing the roots of arcs, granodioritic to granitic plutonism during and immediately following collision is probable, due to partial melting of arc or microcontinental crust, or of the underlying mantle forming the rest of the lithosphere, by overthickening during collision (Dewey and Kidd, 1974). This process may have contributed a large proportion of the plutons in granodiorite–greenstone terrains, including most or all of the widely reported plutons that are syn-kinematic to some of the deformation sequence.

Two examples of alkaline volcanics and sills are reported from greenstone terrains. Both are in the Abitibi belt of the Superior Province, one near Kirkland Lake (Ridler, 1970), the other near Timmins (Goodwin, 1972). These may be island arc rocks, as suggested by Cooke and Moorhouse (1969) or, if there are more tectonic boundaries present than have presently been recognized, Archean seamounts washed into a trench complex (as is portrayed imminent in Fig. 2B).

Relatively small blocks mostly composed of quartzo-feldspathic rocks occur within the dominantly greenstone–granodiorite terrain of the Superior Province, the English River Gneiss Belt and the Quetico block being the best known examples (Goodwin, 1972). It is reported that the rocks are largely 'paragneiss', derived from quartzo-feldspathic sediments. We suggest that it is more likely that they will be found, after more detailed study avoiding use of the word paragneiss, to consist mostly of extremely deformed plutonic rocks, such as reported for the 'gneissic' terrain of West Greenland by Bridgwater et al. (1974). The boundaries of the two blocks mentioned are reported as zones of intense faulting and mylonitization (Goodwin, 1972). In some areas, particularly on the northern side of the English River Belt in Manitoba (Wilson et al., 1972), and to a lesser extent on the northern side of the Quetico block, tectonic slivers of serpentinite, and serpentinized ultramafic rock, are clearly associated with these zones, giving them a similar character to oceanic suture zones of younger orogenic belts. We suggest that these two blocks, and others like them, are microcontinents in terms of their present relationships to the granodiorite–greenstone terrains. Remnant arcs, and inactive arc complexes, can also be regarded for tectonic purposes as microcontinents with respect to active convergent plate boundaries and their active arcs. Such microcontinents may in some cases be identifiable by containing a mixture of plutonic rocks of two greatly different ages, one from its arc history, and a younger from its collision with an active arc that will only give ages near and including the younger age range. Such a relationship was suggested to be common in the Superior greenstone–granodiorite terrain by A. C. Lawson (1913) on the basis of clasts of granitic rock in greenstone belt conglomerates. Lithologic associations very similar to those in greenstone–granodiorite terrains, and deformed in a similar style, can be found locally within Phanerozoic orogenic belts. One example is the Round Pond area of south-central Newfoundland (Fig. 3A). This area, which includes a well-developed ophiolite complex, was deformed and intruded by silicic magmas during the Acadian continental collision, which was relatively weak in the salient of central Newfoundland (Dewey and Kidd, 1974). The 'triangular syncline' morphology, the scale, and the lithological assemblage in the Round Pond area are very similar to those in Archean greenstone–granodiorite terrains, illustrated for comparison in Fig. 3B by the Bulawayan greenstone-belt (Amm, 1940). The steep metamorphic gradients shown by the narrow dynamo-thermal aureoles of the silicic plutons in the Round Pond area are a particular point of resemblance to the Archean terrains. An even better example is found in the northwestern Sierra Nevada of California (Fig. 4); the resemblance of this area to Archean greenstone–granodiorite terrains is astounding. The ultramafic and gabbroic rocks in this area are almost all or perhaps entirely variably dismembered ophiolite complexes including, just to the

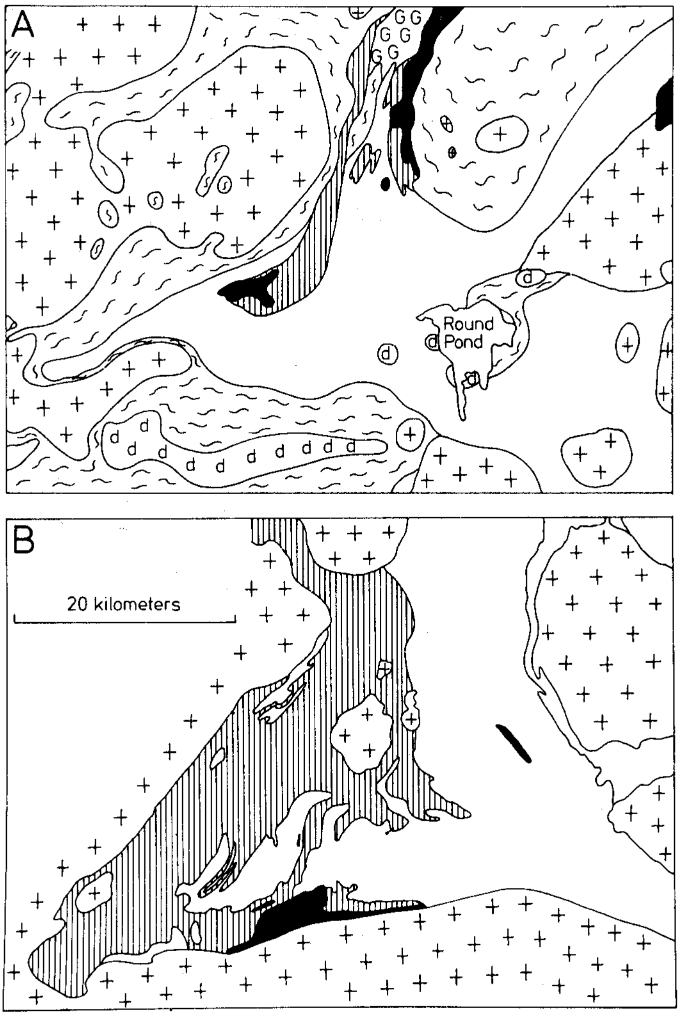

Fig. 3. Outline geological maps at the same scale to illustrate similarities in gross morphology, rock types and structural relationships between part of a Phanerozoic orogenic belt and part of an Archean greenstone belt. A. Round Pond area, central Newfoundland (Williams, 1970). B. Bulawayan greenstone belt, Rhodesia (Amm, 1940). Black ultramafic rocks; G—gabbro; white—sediments; vertical lines—mafic volcanics (mainly basalt and metabasalt); broken wavy lines—high grade metamorphic rocks (probably dynamo-thermal aureoles around 'granites'; crosses—'granites'; d—diorite

south of the area shown, one with well preserved sheeted dykes (R. G. W. Kidd, pers. comm.; Moores, 1975). Tectonic boundaries and slides are abundant (E. Nisbet, pers. comm., in prep.). The ophiolite complexes and the mafic volcanics and volcaniclastics are probably at least partly of marginal basin origin.

It is important to recognize that the apparently swirly and plastic deformation and intrusion features seen in the Phanerozoic analogues cited for the Archean greenstone–granodiorite terrains are the sum of a complex series of events. In each case these occurred during a significant time period in an area where the *instantaneous* situation at any time

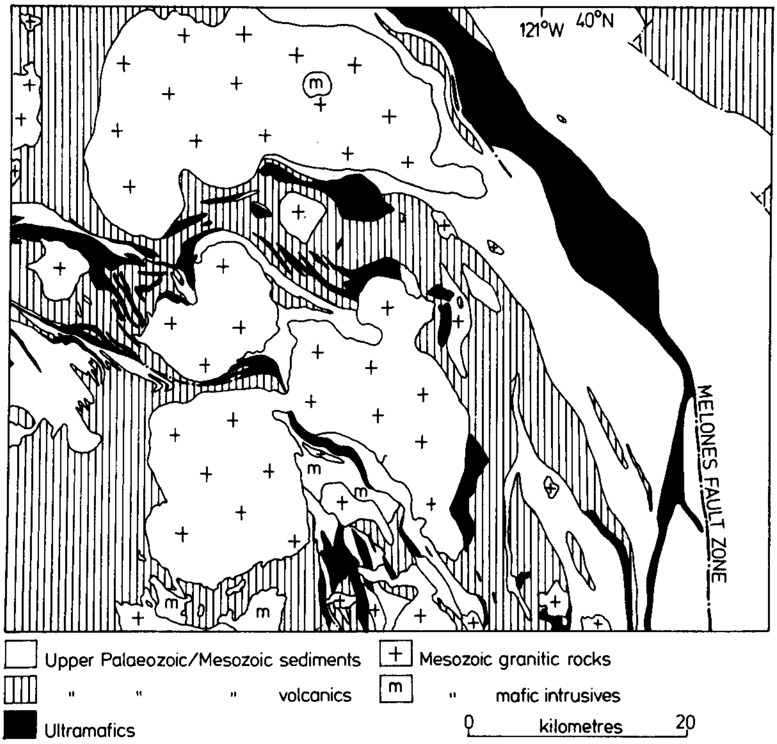

Fig. 4. Outline geological map of part of Northwestern Sierra Nevada, California (Burnett and Jenkins, 1962) to illustrate similarity to Archean greenstone belts

was either relative plate motion at sharply defined boundaries, or, during short-lived orogenic episodes, significant horizontal shortening with *consequent* upward motion of material to compensate. These episodes may be due either to collision across a marginal basin or a 'main' ocean, or to other (at present) poorly understood compressive orogenic episodes within and behind Andean and island arc complexes. We see no reason to suppose anything different for the Archean terrains.

An extensive gneissic terrain containing granulite-facies rocks forms the part of the Superior Province in the Ungava Peninsula. The radiometric ages (2·5–2·4 b.y.) and the nature of the rocks (Stevenson, 1968) show this to be an area of basement reactivation, which we interpret as due to convergence after a continental collision about 2·5 b.y. ago. This type of orogeny, where internal continental fractionation takes place (Burke and Dewey, 1970, 1973), seems to have been relatively uncommon in Archean time, as shown by the large areas of granodiorite–greenstone terrain that have survived. We suggest that it was the general incompleteness, or weakly developed nature of collisions in the Archean that has lead to widespread preservation of the greenstone belt material only locally preserved in younger orogenic belts.

The supposed 'vertical' tectonic style reported from greenstone belts seems to be based on the alleged control of deformation by the silicic plutons, the alleged simple synclinal structures of the belts and, perhaps, on the subvertical elongation lineation commonly observed. The latter is a necessary consequence of horizontal shortening across a steep cleavage, because vertical displacement of plastically deforming material is usually easier than lateral displacement. The supposed simple synclinal structure of greenstone belts is a myth (Ramsay, 1963), discussed below. The steep inclination of cleavage and the overall steep structure of greenstone belts is quite compatible with collisional tectonics, being observed in Phanerozoic orogenic belts where incomplete collision has occurred (Dewey and Kidd, 1974). It is also found in very complete collisions in tightly sutured zones such as the Indus suture (Gansser, 1964). Although the Indus suture, and others like it, represent oceans driven out by low to moderately dipping subduction and thrusting, the suture zone itself becomes oversteepened in the later stages of the collisional process. This effect is also demonstrable in the Baie Verte Lineament (Kidd, 1974) on a scale more compatible with an individual greenstone belt. Ramsay (1963) has shown that the silicic diapirs do not significantly control the deformation in the Barberton greenstone belt. It is difficult to see how a strong regional cleavage and associated horizontal shortening could be impressed on relatively hard, cold volcanics and sediments by soft, mushy granitoid diapirs. An externally applied horizontal compressive stress, such as is available during arc/arc or arc/microcontinent collision, is mechanically more reasonable, and is more compatible with the behaviour of the planet on which the major horizontal shortening clearly documented by Bridgwater et al. (1974) for the Archean of West Greenland occurred.

Ophiolite Complexes and Archean Oceanic Crust and Mantle

No examples of oceanic crust and mantle similar to fully developed ophiolite complexes have as yet been reported from greenstone belts. Our suggestion that Archean oceanic crust was thicker than present oceanic crust leads to the idea that possible examples of tectonic slices of Archean oceanic crust may not often reach down to the depleted harzburgite-dunite that is a diagnostic and characteristic of their Phanerozoic counterparts. We also suggest that the somewhat higher heatflow and perhaps less depleted mantle in Archean times resulted in a significantly greater thickness of essentially ultramafic cumulates and a proportionately lesser thickness of gabbro than is seen in Phanerozoic ophiolite complexes. As Moores (1973) pointed out the stratiform anorthosites of the Early Precambrian (e.g. Fiskenaesset and the Limpopo) may also represent the upper part of the cumulate gabbro of the thick Archean oceanic crust. Many potential examples of Archean oceanic crust (regarding ultramafic cumulates as crust for brevity) are stated to be sills. For example, the Bird River 'sill' in Manitoba consists of cumulate ultramafics and cumulate gabbro (Wilson et al., 1972). The relationships of the 'sill' to surrounding rocks are not well known, perhaps because it has been assumed to be a sill. It is stated (Wilson et al., 1972) that the contacts with volcanics or sediments are nowhere seen in outcrop, except where they are faulted. It seems entirely possible that they could be wholly tectonic, or perhaps, in some places, that there might be a gradation into diabase, sheeted dykes and pillow lava. In considering this, and many other examples of large supposed sills in the Archean greenstone belts that might have been chosen for discussion, we urge that the possibility of their being samples of oceanic crust (and perhaps mantle) be very carefully considered, if only because of the rich reward of information, particularly geochemical information, that would ensue from the positive identification as Archean oceanic crust. We do not intend to deny the existence of large genuine differentiated sills in greenstone belts, nor do we suggest that the Bird River Sill is definitely not a sill. We only wish to point out that there is reasonable doubt about the external relationships of many of these bodies that needs to be clarified by detailed mapping. Many Phanerozoic ophiolite complexes, now known to be variably dismembered and tectonically bounded slices of oceanic crust and mantle, were originally described as intrusive sills, despite the essential absence of significant metamorphic aureoles and the presence, usually overlooked, of narrow tectonic boundaries with large displacements. One example includes the Bay of Islands Complex (Smith, 1958). A more dismembered example, the Mings Bight and related Baie Verte Lineament Ophiolite Complexes (Bird et al., 1971; Dewey and Bird, 1971) were also originally described as concordant lenticular ultramafic and gabbro sills (Watson, 1943). Two common mistakes are to regard plutonic rocks as necessarily intrusive and in isolation from surrounding mafic volcanics. The fact that layering in them is coplanar with bedding in surrounding volcanics is immaterial; this is the usual case for ophiolite complexes of Phanerozoic orogenic belts in all stages of tectonic dismemberment. Many examples of supposed sills very similar to the cumulate ultramafic and gabbroic parts of ophiolite complexes are known in Archean greenstone belts, and we suggest that many of them are good candidates for samples of Archean oceanic crust, perhaps including, in some instances, depleted non-cumulate upper mantle as well.

These possible ocean floor fragments in greenstone belts with associated mafic volcanics and volcaniclastics closely resemble those dismembered ophiolites interpreted as being of marginal basin origin in Phanerozoic mountain belts (e.g. Bird et al., 1971; Dalziel et al., 1974). We suggested above that the contents of marginal basins are very likely to compose a large proportion of Archean greenstone belts. We suggest that there are two reasons why any Archean oceanic crust that may be identified will be more likely to represent marginal basin rather than main ocean crust. First, the permobile ocean with many scattered microcontinents and arcs appears likely to have had many more areas in which young ocean floor was close to arc and microcontinental sediment sources than in later times. Secondly, marginal basin ocean floor has a greater chance of being young, and hence thin and hot, near to a subduction-resistant arc or remnant arc (microcontinent) than does main ocean floor, and therefore has a greater chance of being obducted and hence preserved.

The Paradox of Greenstone Belt Stratigraphy

The environments represented in greenstone belts closely resemble those found in later orogenic belts where intense tectonism is the rule and although both intensive and extensive characters of the permobile earth indicate a dominance of horizontal motion, many publications on greenstone belt geology (e.g. Glikson, 1972, Fig. 4) report very thick (>10 km) tectonically unrepeated sections, or report such sections containing a few tectonic slides and then describe a stratigraphic sequence as if they were not present (e.g. Viljoen and Viljoen, 1970). We would like to suggest that tectonic boundaries, especially thrusts and tectonic slides concordant with bedding, are likely to be present, although as yet generally unidentified, throughout these sections. Map patterns in the Bulawayan greenstone belt (Amm, 1940) support the notion that such repetition is important. Indeed, the term 'schist belt' formerly used in Africa to describe what are now known as greenstone belts, even though it is not appropriate to the general metamorphic grade, does describe accurately, and speaks volumes for the general tectonic condition and state of deformation of the rocks contained within them. With the absence of fossil evidence the problem of sorting out stratigraphic sequences is acute, and we list briefly the following cautionary examples for those who insist that greenstone belt sequences are very thick and do not contain major tectonic repetitions and/or excisions. The examples are essentially instances where apparently homoclinal sequences have been shown from fossil evidence to consist of numerous parallel thrust sheets, or from the discovery of narrow high strain zones (mylonite or tectonic slide zones) throughout the sequence.

(1) East of the Green Mountains in Vermont an east-facing regular stratigraphic sequence is claimed from later Precambrian arkoses to Middle Ordovician volcanics and sediments (e.g. Cady, 1968). It is certain that the Taconic allochthon was expelled from this zone making the possibility of a continuous stratigraphic sequence negligible. The zone is riddled with detached isoclinal fold hinges and narrow mylonite (or slide) zones of high strain.

(2) In Newfoundland, the Bay of Islands ophiolite allochthon together with a series of underlying thrust sheets was long regarded as

a continuous upward facing stratigraphic sequence lying conformably above a Cambro–Ordovician platform carbonate sequence (Smith, 1958), as indeed was the Taconic allochthon of New York State. The discovery of fossils in the allochthons led to a re-evaluation of the sections showing them to be a stacked sequence of thrust sheets (Rodgers and Neale, 1963; Bird, 1969).

(3) The Scardroy and other Lewisian inliers in the Moines were long regarded as parts of the Moine 'stratigraphy' except by Kennedy (1955, especially the discussion following) until Clough's basal conglomerate in Glenelg was finally accepted as the unconformable base of the Moines, and Lewisian radiometric ages were determined.

(4) The stacking of parallel *decken* of Schrattenkalk in the Glaurus valley was recognized by careful lithologic and faunal work and the recognition of the significance of the thin *Lochseitenkalk* bands as calcareous 'mylonites' (Trümpy, 1969).

(5) As a final example, there is the awful lesson of the Col de Genet section in the French Alps, where, until foraminifera were collected, a continuous stratigraphic sequence was 'recognized' from Jurassic carbonates through a *schistes lustré* section with 'interstratified' ophiolites and flysch. Once fossil repetition was detected, very thin mylonite zones were found.

It is emphasized that addition (or removal) of strata several kilometres thick, or juxtaposition of sequences originally deposited a hundred or more kilometres apart, may take place along mylonite (slide) zones as little as a few centimetres thick and without causing any anomalous deformation outside them. Also, once one slide zone is recognized in any succession, it is almost certain that many are present.

Most of the examples cited above are from relatively flat-lying overthrust- and nappe-type situations. Narrow tectonic slides with large displacement are not confined to such terrains. For instance, a stack of nappes may be deformed into tight folds, and will then not only contain the earlier tectonic slides between nappes but also new ones coplanar with the axial surfaces of the folds. Simpler terrains affected by fairly upright large-scale tight folding, a single steep axial surface cleavage, and a moderately to steeply pitching elongation lineation are well known to contain and be bounded by large tectonic slides, for example the Croagh Patrick Syncline (Dewey, 1967) and the Baie Verte Lineament (Kidd, 1974; Dewey and Bird, 1971). It is interesting that both these examples are of rocks deposited in small Appalachian/Caledonian marginal basins, and deformed during weak continental collision (Caledonian, Acadian) by closure of the basins.

The greenstone belt that has been most adequately mapped and studied from a structural point of view is the Barberton Mountain Land. In a germinal, but conspicuously and unjustly ignored paper, Ramsay (1963) showed that the majority of rocks in this belt are strongly and complexly deformed in a manner no different from parts of Phanerozoic orogenic belts. He demonstrated that there are at least three distinct and regionally significant deformation episodes recognizable; that they all involved *horizontal* shortening as their principal strain component; and that the granodioritic plutons in his map area were intruded during the second deformational episode, are affected by it, and do not cause significant deformation themselves. His mapping also shows the presence of major tectonic slides formed during the second, cleavage-producing deformation. Inspection of the maps of Viljoen and Viljoen (1970, Fig. 7) shows that these second episode tectonic slides are abundant throughout the Barberton greenstone belt. Their maps also show (Viljoen and Viljoen, 1970, Fig. 2 and Fig. 7) that there is at least one major tectonic slide present that was formed during the first deformation (at the base of the Komati Formation) and their description of 'sheared and talcose' rocks just below the Middle Marker horizon may indicate the presence of another early tectonic slide. This terrain therefore resembles the hypothetical example given above where a stack of nappes is tightly folded with development of new tectonic slides. It also is stated (Viljoen and Viljoen, 1970) that the uppermost unit (Swartkoppie) of the Onvervacht sequence is everywhere separated from the other units by 'faults' (second episode tectonic slides), and yet it is shown with the other units in a continuous stratigraphic column. The Barberton Mountain Land seems to be regarded unofficially as the type example of a greenstone belt, and rightly so in view of the detailed mapping available and the relatively good exposure. As this belt has the structural properties listed above, it is highly unlikely that other greenstone belts are very different.

Wood (1966) stated that the dominant structure in Rhodesian greenstone belts is a steeply inclined cleavage across which up to 90% shortening has occurred. In the Superior Province a steep cleavage is the rule, and narrow zones of high strain, which are probably tectonic slides, are common (Goodwin, 1972). In particular, the Kirkland Lake–Larder Lake 'break' (Ridler, 1970; Goodwin, 1972) greatly resembles major tectonic slides in younger terrains. It contains, at the Kerr-Addison mine, a magnesite–quartz–fuchsite rock tectonically derived from ultramafic rock, and in this respect closely resembles major tectonic slides, found in some Phanerozoic terrains, that contain this rock type and tectonic slivers of ophiolite-derived harzburgite and dunite. Examples are found in Newfoundland (Kidd, 1974; Dewey and Bird, 1971) and California (including the area of Fig. 4). The magnesite–quartz–fuchsite rock was described as carbonate iron-formation by Ridler (1970), despite the negligible iron content, and illustrates the problems encountered in interpreting supposed stratigraphic sections when structural effects have not been considered. Major tectonic slide zones derived from mafic or felsic plutonics, volcanics or volcaniclastics may very easily be mistaken as merely strongly cleaved fine-grained mafic or felsic volcaniclastics if a worker is unfamiliar with the properties of tectonic slides. Archean greenstone belts in Western Australia are also complexly and strongly deformed and tectonic boundaries are common (P. F. Williams, pers. comm.). Supposed very thick, intact stratigraphic sequences described from there (e.g. Glikson, 1972) are therefore also unlikely to be real, especially in view of the generally poor outcrop.

We do not wish to suggest that partial stratigraphic sections cannot be obtained from greenstone belts. We do, however, suggest that major tectonic slides are common, and that stratigraphic sequences presently described from greenstone belts, especially those more than about 10 km thick, are likely to consist of several tectonically stacked sequences. If this is the case, correlation between tectonic slices is likely to be difficult or impossible in the enforced absence of fossils, unless very distinctive marker bands are present. We emphasize the fact that young untectonized volcanic stratigraphy is very heterogeneous and complex on both small and large scales, and that the few basic lithologies involved make even gross lithologic correlation very difficult merely in Neogene examples. We also suggest that it is very important to state where a thickness measurement was made, because significant changes occur during deformation, with shortening of 50 to 90% across cleavage and thickening of two to five times in fold hinges being commonly reported from greenstone belts (Wood, 1966).

Once the stratigraphy of greenstone belts is recognized as the product of tectonic stacking, many of the anomalous features associated with the belts cease to be anomalous. Thus apparent vast thicknesses become readily understandable and repeated cycles from ultramafic to mafic cease to be hard to explain. We suggest that the long-recognized close resemblance of the rocks of greenstone belts to those of oceans, marginal basins and arcs is real, and that they do mark the places where oceanic crust has been removed, just as do the suture zones of younger orogenic belts.

Conclusions

The characteristics of ancient ocean and continent outlined here have been inferred by assuming that rocks similar to those forming today were made by similar processes, and by allowing for the effects of faster heat generation in the past. Because much of the heat generated in the Earth today is dissipated in making oceanic lithosphere at spreading ridges, in ageing it on ocean floors, in partly melting descending slabs of oceanic lithosphere below island arcs, and in emplacing the igneous products of this melting, we have inferred that these processes operated more effectively during the permobile regime prior to 2·5 b.y. ago. The picture outlined by considering these intensive properties is of an Earth at the end of the permobile regime covered much as today with about one-third continent and two-thirds ocean, and the volume of water produced from the mantle concurrently with lithosphere being also similar to the present (Wise, 1974). We suggest that the length of plate boundary was, however, much greater, and this explains many of the differences. The extensive characters of exposed Archean terrains are consistent with this picture and permit its refinement.

We suggest that models based on erroneous views of the structure of greenstone–granodiorite terrains are unlikely to be realistic. Strong and Stevens (1974) state that the greenstone belts have been deformed by vertically acting forces, which is clearly incorrect (Ramsay, 1963), and they imply a random pattern of Archean magmatism, a point which has yet to be demonstrated. Their model is based on the intersection of higher Archean geotherms with the peridotite solidus. The present asthenosphere is thought to be a zone of partial melting resulting from the same process (Wyllie, 1971) and the effect of a higher Archean geotherm would be to increase its thickness and to reduce slightly the thickness of the lithosphere, thereby probably facilitating lithospheric relative motion. Present lithospheric relative motion leads to ordered magmatic sequences with calc–alkaline magmatic products in arcs cutting and overlying tholeiites of the oceanic crust. Archean sequences resemble those of the present day and imply that magmatism in the Archean was also not random. Possible convective instability in a thicker asthenosphere with a somewhat thinner lithosphere would presumably give rise to 'hot spots' like those of the present day, except that their composition would probably be tholeiitic rather than alkalic.

We therefore infer that the rocks in the granodiorite–greenstone terrains were made at spreading ridges (and perhaps 'hot spots') in oceans and marginal basins, and in and around arcs above descending slabs of oceanic lithosphere. We emphasize the potential importance of marginal basins and oceanic crust in the greenstone assemblages. We propose that the deformation and some of the plutonic activity occurred due to horizontal shortening during collision of arcs and microcontinents, and we suggest that the greenstone-belt sequences are very much more tectonized than has generally been recognized.

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Preface

The early history of the Earth was so long and complicated that current students usually work only on one or two aspects of the subject. But the research subjects include field geology, geochemistry, structural geology and tectonics, the evolution of the atmosphere and the oceans, palaeontology, geochronology and metallogeny, so it is not surprising if specialists in one branch commonly have little knowledge of another; and yet many of these fields overlap and knowledge of one should influence and contribute to that of another. The oldest rocks known at present have an age of about 3.8 b.y., but most geologists know little of that pre-geological period before 3.8 b.y. when the core, mantle and protocrust formed. The aim of the NATO Advanced Study Institute, held from 5–11th April, 1975, at Leicester, was to bring together specialists from many fields to produce their latest findings and ideas and to discuss the present state of knowledge on early Earth history in the period 4.5–2.5 b.y. ago. Much of the success of the meeting was due to the fact that 2 hours of every day was devoted to discussion of specific problems. This volume contains the text of papers presented at the meeting with the object of providing an inter-disciplinary approach for the student, teacher and researcher to the problem of how the Earth evolved in its early stages. With such a large subject it is obviously impossible to be comprehensive, but it is hoped that this volume goes some way in providing an integrated compilation. The papers vary from long reviews of major subjects to short reports of recent research. Some are quantitative, some speculative and some highly controversial, but this variation reflects the state of current research in the subject. No one pretends that research into the early history of the Earth has reached an advanced stage of development—far from it; there are not enough constraints to enable one to choose between various alternatives and models in almost every field. I hope that this volume imparts to the student the controversial nature of our knowledge of the Archaean and pre-Archaean.

The idea of having a NATO Advanced Study Institute on the early history of the Earth came from Professor J. V. Smith in 1971 when I was working with him on Archaean rocks in Chicago and when he was involved in the early organization of the 1972 Feldspar NATO ASI at Manchester. I am grateful to him and to Professor J. Sutton (Imperial College, London) for many discussions since then on the organization of the meeting, and to Professor P. C. Sylvester-Bradley of the Department of Geology, Leicester University, who made many useful suggestions and gave much encouragement.

The Institute was attended by 141 scientists from 11 NATO and 67 other countries. The organizers are grateful for a generous grant from the Scientific Affairs Division of NATO which covered the accommodation and partial travel costs of participants from NATO countries, partial support for the field excursions in Scotland, and the organizational expenses of the meeting. The National Science Foundation of the U.S.A. kindly provided travel grants for three research students, and the International Union of Geological Sciences financed two visitors from non-NATO countries. UNESCO gave a grant to enable five delegates from non-NATO countries to attend the first full committee meeting of the International Geological Correlation Project on Archaean Geochemistry, convened by Dr A. Glikson and held during the Conference.

Thanks are due to many people who helped to make the meeting a success, especially: Drs J. Peal and K. Davies, Wardens of Villiers and Gilbert Murray Halls of Residence, for providing excellent facilities; the technical and secretarial staff of the Geology Department of Leicester University for their services; Miss J. Baker and Mr M. Clarke for general assistance throughout the meeting; and Professor J. Watson who led two field excursions, to the Scourian of NW Scotland and of the Outer Hebrides, before and after the meeting. In particular, I am grateful to Judith, my wife, for indispensable administrative and secretarial assistance pre-, syn- and post- the meeting; she also organized the field excursions.

June, 1975 BRIAN F. WINDLEY

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