PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Marginal Basins through Geological Time [and Discussion]

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Phil. Trans. R. Soc. Lond. A 1981 **301**, 217-232
doi: 10.1098/rsta.1981.0107

Phil. Trans. R. Soc. Lond. A **301**, 217–232 (1981)
Printed in Great Britain

Marginal basins through geological time

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Marginal basins are common features of present-day plate tectonics. Whereas some may represent trapped segments of normal ocean floor, many owe their origin to extensional seafloor spreading behind active volcanic arcs. They exhibit a variety of forms. Some are completely intraoceanic; others develop at continental margins, where back-arc spreading may lead to the detachment and dispersal of continental fragments. Marginal basins can be recognized in the early stages of formation; others have developed through more than one pulse of back-arc extension, and some have aborted shortly after formation. Closure of marginal basins may result in preservation of part of the basin floor as obducted ophiolite.

Although the reasons why seafloor spreading occurs behind volcanic arcs are still imperfectly understood, all suggested mechanisms invoke a strong link with subduction. Thus if subduction occurred in the past it is logical to expect that fossil marginal basins may be preserved in the geological record. However, allowing for the gradually evolving thermal and chemical nature of the Earth's mantle, ancient marginal basins need not necessarily duplicate every feature of modern ones. This contribution examines possible Phanerozoic, Proterozoic and Archaean marginal basin analogues in the light of the geological features shown by modern basins and attempts to assess their importance for crustal development.

Introduction

Marginal basins are an important feature of the present plate tectonic régime (Karig 1974; Uyeda & Kanamori 1979), and are closely associated with lithosphere subduction. As most of the Earth's present lithosphere subduction is taking place around the Pacific Ocean margin, so most actively forming or recently active marginal basins occur in the Pacific, particularly the western Pacific. The minor amount of subduction occurring in the western Atlantic, in the Caribbean and the Scotia Sea, has also given rise to marginal basins. Being located at destructive margins of course, marginal basins have a limited lifespan and are readily subject to closure. Nevertheless fossil marginal basins (or evidence for the existence of such basins) are increasingly being recognized in the Phanerozoic record (see, for example: Dickinson 1978; Dewey 1976), during which it is generally admitted that plate tectonic processes have operated.

Modern marginal basins occur in a variety of forms. They may occur between an active volcanic arc and a continent, between an active arc and a remnant arc, between two remnant arcs or between a remnant arc and a continent; some may represent trapped ocean floor, others appear to have formed by some form of back-arc spreading, and yet others perhaps formed as a result of ridge subduction and/or the existence of 'leaky' transform faults. They may be completely intraoceanic, or develop by rifting and severing of a continental margin to varying extents, or take the form of lithospheric thinning without actual rifting. Different types of basin may not only vary widely in their sedimentation history, but also exhibit diversity in the nature and composition of the volcanic products forming the floor of or flanking the basin.

Because marginal basins are such a common present-day phenomena, and are invariably linked to subduction, it seems reasonable to assume that they may have formed in ancient times, provided that some form of lithospheric subduction took place (irrespective of whether strictly plate tectonics or not). Here we summarize the main features of modern and Phanerozoic marginal basins and examine possible Proterozoic and Archaean analogues in an attempt to assess the role of marginal basins in crustal development through geological time. It is not assumed that all ancient marginal basins conform to a single pattern of development; the great variety of forms exhibited by modern basins indicates that this may be too simplistic. The conditions under which such basins formed may have changed subtly with time as the Earth's mantle has evolved. Nevertheless certain basic features of their volcanic, sedimentary and tectonic history are common to all basins and serve to aid their recognition in ancient terrains.

Modern and Phanerozoic marginal basins

Modern marginal basins are both intraoceanic and ensialic. Intraoceanic basins have developed entirely within oceanic regions, the best example being those associated with the Mariana–Bonin arc system in the western Pacific, the Lau and Fiji basins in the southwestern Pacific, and the East Scotia Sea in the South Atlantic. Ensialic basins have formed at continental margins and show a whole range of development from simple back-arc extensional faulting, through sedimentary basins with or without associated basic volcanicity, to basins where the subcontinental lithosphere has been ruptured and the continental-based volcanic arc become separated from the stable continental margin by a basin with essentially ocean-floor characteristics. Potentially marginal basins could, by continued spreading, develop into major ocean basins. In practice no such example is known. It appears that spreading in marginal basins is somehow linked to subduction and may not persist for more than about 15–20 Ma.

The spreading process in intraoceanic marginal basins is broadly similar to that at major ocean ridges. Good datable magnetic lineations were first described from the East Scotia Sea (Barker 1972) and lineations have subsequently been recognized in the Shikoku Basin (Watts & Weissel 1975; Watts *et al.* 1977*b*; Shih 1980), Parece Vela Basin (Mrozowski & Hayes 1979), Mariana Trough (Karig *et al.* 1978), South Fiji Basin (Watts *et al.* 1977*a*) and Lau Basin (Lawver & Hawkins 1978). However, whereas the faster spreading basins such as the East Scotia Sea have yielded correlatable magnetic anomaly patterns, many of the smaller, more slowly spreading basins, particularly the Lau Basin, North Fiji Basin and the Mariana Trough, have diffuse low-amplitude anomalies which are mostly uncorrelatable. Lawver & Hawkins (1978) have suggested that the spreading process in many marginal basins is less regular, with more transforms, abundant ridge jumps, and magmatism distributed over a wider zone. As a consequence the bathymetry is less regular and there are more sea mounts with differentiated lavas. Recent deep-sea drilling of western Pacific basins has shown off-axis basaltic sills to be common. This may in part be a result of the higher sedimentation rate in marginal basins, as has recently been demonstrated for the Gulf of California (Curray *et al.* 1979*a*), but may also be a consequence of magmatism being distributed over a wider zone, contributing to the high heat flow. Barker & Hill (1980) have argued that spreading in some marginal basins is asymmetric, with accretion being favoured on the arc side. Marginal basins tend to be deeper than normal ocean crust of equivalent age.

Available data on the geochemistry of marginal basin basalts from the Scotia Arc region and from the western Pacific have been reviewed by Tarney *et al.* (1981). Petrologically and geochemically basalts formed as a result of back-arc spreading have many similarities with those generated at the major mid-ocean-ridge systems. They include types with the low concentrations of incompatible trace elements characteristic of normal mid-ocean-ridge basalts (N-type m.o.r.b.) as well as those with higher contents of such trace elements similar to the basalts (E-type m.o.r.b.) that occur near Iceland, 45° N, or the Azores, on the Mid-Atlantic Ridge (Tarney *et al.* 1979, 1980). Both types of basalt occur in the West Philippine Basin, regarded as a 'trapped' basin behind the Kyushu–Palau Ridge, as well as in the Shikoku Basin, which was formed by back-arc spreading after splitting of the Kyushu–Palau Ridge. Other basalts in back-arc basins have moderately high concentrations of large-ion lithophile elements such as K, Rb, Ba, Sr and the light rare-earth elements (r.e.e.) but do not show the similar enrichment in high-field-strength (h.f.s.) elements, such as Nb and Ta, that characterizes E-type basalts from the Mid-Atlantic ridge. This type of geochemical signature of course is typical of island-arc and calc-alkaline basalts (Saunders *et al.* 1980*a*).

Tarney *et al.* (1980*a*) noted that basalts with this transitional 'arc'-like signature (i.e. high ratios of large-ion lithophile (l.i.l.) to h.f.s. elements) were erupted during the early stages of back-arc spreading, when the spreading was initiated through splitting of a *calc-alkaline* volcanic arc, but were less common when the spreading occurred through splitting of a (more primitive) arc built of island-arc tholeiite lavas. Thus the Shikoku and Parece Vela basins developed through splitting of the Kyushu–Palau tholeiitic arc and the erupted basalts have no 'arc' signatures; however, during the later stages of development of the Mariana Arc system, the Mariana Trough developed through splitting of the calc-alkaline West Mariana Ridge, and basalts in the Mariana Trough, though somewhat variable in chemistry, do include many with higher l.i.l. element contents, moderate light r.e.e. enrichment, but with low Nb and Ta concentrations (Tarney *et al.* 1980*a*). This feature is distinctive of ensialic marginal basins such as Bransfield Strait (Weaver *et al.* 1979) and the Mesozoic marginal basin in southern Chile (Saunders *et al.* 1979). In the latter, basalts showing light r.e.e. enrichment gave way, as spreading continued to basalts showing light r.e.e. depletion (Stern 1980). Although such features could, in specific instances, be attributed to some form of dynamic melting (Langmuir *et al.* 1977; Duncan & Green 1980), Tarney *et al.* (1981) consider that, taking all evidence into account, a more probable explanation is that the mantle beneath island arcs becomes progressively metasomatized with time, either because l.i.l. elements are distilled from the downgoing slab or because water released from the slab causes incipient melting in the overlying mantle wedge, leading to upward enrichment of l.i.l. enriched melts or fluids. Mantle permeated by such fluids and included in the diapiric activity that splits the arc hence contributes to the basalts erupted during the early stages of back-arc spreading. Some involvement of water from the downgoing slab is suggested by the fact that marginal-basin basalt glasses have higher water contents and higher H_2O/CO_2 ratios than their normal mid-ocean-ridge equivalents (Muenow *et al.* 1980).

Sedimentation in marginal basins is very variable, simply on account of the wide variety of form of marginal basins, but sedimentation rates are normally higher (often significantly so) than in the major ocean basins. Trapped basins of course may have a typical ocean-floor sequence of chalks, cherts, siliceous or calcareous oozes and clays; the central parts of marginal basins which have formed by back-arc spreading and are well separated from the active arc and the remnant arc or the adjacent continent may have a similar sequence of sediments. On the other hand those sections of marginal basins that developed early in the spreading cycle will have an abundant supply of volcanigenic debris (greywacke, sands or clays) from the adjacent arc, to an extent depending on such factors as the nature of the arc volcanism, whether the arc was active during the spreading episode (cf. Sharaskin *et al.* 1981) and even the prevailing wind direction. The sediment contribution from the continental side may also vary widely, depending on the relief of the continental hinterland, location of major rivers, climate, distance from the continent and whether or not there is an intervening remnant arc. Ensialic basins have the potential for the development of extraordinarily thick sedimentary sequences, depending on the degree of extension, i.e. whether back-arc diapirism produces complete back-arc rifting or just crustal thinning. The sediment supplied may come both from the active volcanic arc and from the continental side. Sediment distributions of this type have been recognized in the southern Chile marginal basin (Dalziel *et al.* 1974) and its extension into South Georgia (Dalziel *et al.* 1975; Suarez & Pettigrew 1976). Again the production of sedimentary debris will depend on factors such as the activity of the adjacent volcanic arc, wind direction, relief and drainage patterns on the continental side, and climate. But in most cases there is little doubt that there are few difficulties in filling an ensialic basin.

In summary, the sedimentary characteristics of marginal basins are potentially so variable that few criteria can be constructed to unambiguously identify past marginal basins. The main criterion with which most basins might concur is sheer volume of greywacke-type sediments.

The tectonic development of true back-arc marginal basins is marked by some form of mantle diapirism which to a greater or lesser degree 'splits' the volcanic arc (either oceanic or continental margin). In southern Chile the first expression of this was the outpouring of vast volumes of silicic volcanics and ignimbrites (the Series Tobifera), which Bruhn *et al.* (1978) have interpreted as the result of subcrustal melting. Similar silicic volcanics are known from the Gulf of California (Karig & Jensky 1972), the Basin and Range Province (Scholz *et al.* 1971), and the Taupo volcanic zone of New Zealand, all of which are regarded as back-arc basins in varying stages of evolution. Ensialic basins, then, tend to be zones where the volcanism is bimodal basic–acid instead of having the broader range of composition that characterizes the adjacent calc-alkaline volcanic arc. Because there is some evidence from western Pacific arcs and marginal basins (Sharaskin *et al.* 1981) that arc magmatism and marginal basin magmatism tend to alternate, it may be that the silicic–basic volcanism in ensialic basins may not be accompanied by contemporaneous andesitic arc volcanism, though this may occur later.

Marginal basins, being sited at zones of plate convergence, are rather vulnerable features, which are readily destroyed by arc–continent or continent–continent collision. Most of the present arcs and marginal basins in fact are rather young features, less than 40 Ma old. Small basins, such as that in southern Chile, are uplifted, but preserved almost authochthonously (Bruhn & Dalziel 1977) during basin closure. Young marginal basins are relatively buoyant features, because of their recent thermal history, and hence are more easily obducted than the normal old, cold and dense ocean crust being consumed at subduction zones. Many ophiolite complexes may thus represent marginal basin floor (Hawkins 1977; Dewey 1976; Saunders *et al.* 1979) and can be used, along with other igneous, sedimentary and tectonic criteria, to establish the former existence of marginal basins in the geological record.

Such evidence for the existence of past marginal basins is increasingly being recognized throughout the Phanerozoic. Thus marginal basins have been recognized in California dating back to the late Palaeozoic (Schweikert & Cowan 1975; Dickinson 1976; Churkin & Eberlein 1977); they are common in the Alpine belt (Boccaletti & Guazzone 1974; Dewey *et al.* 1973), and particularly in the Appalachian–Caledonian belt (Burke *et al.* 1976; Kidd 1977; Ruitenberg *et al.* 1977), and even in the Cambrian (Crawford & Keays 1978) and the late Proterozoic (Schmidt *et al.* 1979). Because the Pacific appears to have been in existence as a major ocean throughout much of the Phanerozoic, and because island arcs and marginal basins are, by definition, associated with subduction and oceanic margins, the geology of a significant proportion of the circum-Pacific margin, from California to southeast Asia, is dominated by the effects of arc accretion of various types (Dickinson 1978; Hamilton, this symposium). The arcs and portions of marginal basins accreted in this way constitute an important component in the growth of the continental crust in these areas, at least during the Phanerozoic. Marginal basins and arcs of many different types are probably represented in these belts, but, because the very act of arc accretion also means marginal basin destruction, it is often very difficult to decipher the original form and nature of the basins.

Before we consider whether marginal basins, in one form or another, extend back in time beyond the Phanerozoic, it is probably worth reviewing some of the mechanisms that have been proposed to explain the formation of marginal basins.

Mechanisms of marginal basin formation

There is as yet no consensus as to the mechanism controlling the formation of marginal basins, although several have been proposed, some as general mechanisms, others for specific basins.

Uyeda & Miyashiro (1974) suggested that subduction of the Kula Ridge below the Asian mainland in the early Tertiary may have induced diapirism which led to the opening of the Japan Sea. While the supply of ridges able to be subducted may be too limited for this to be a universal explanation, subduction of the East Pacific Rise under Baja California 20 Ma ago may have led to the opening of the Gulf of California, which has further developed as a sort of 'leaky' transform with extension taking place along short ridges separated by long transform faults (Uyeda & Kanamori 1979). The Andaman Sea may be another example (Curray *et al.* 1979*b*). The S Chilean marginal basin (Dalziel *et al.* 1974), with its discontinuous mafic lenses, may also have developed as a leaky transform in the late Jurassic (Dalziel 1981).

A more generalized model for marginal basin opening has been developed by Toksöz & Bird (1977) and Toksöz & Hsui (1978). They suggest that the downgoing slab induces convective flow in the overlying mantle wedge, which eventually brings hot mantle material to the base of the lithosphere behind island arcs, heating and weakening the lithosphere and permitting it to rupture in the tensional environment. Their numerical models predict that back-arc spreading should take place within 5–10 Ma after the initiation of subduction. Such a model would broadly fit the development of the Parece Vela and Shikoku basins (which began opening 30–35 Ma ago), if one assumes that subduction began at the Kyushu–Palau Ridge 40 Ma ago coincident with the change in direction of motion of the Pacific plate at this time (Jurdy 1979). A feature of these models is that rifting of oceanic lithosphere is easier than that of continental lithosphere, which concurs with the abundance of present-day intraoceanic marginal basins as opposed to ensialic basins.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Mantle diapirism behind the volcanic arc has been proposed to explain the back-arc extension and high heat flow (Karig 1971). The cause of the diapirism has been attributed to frictional heating by the downgoing slab or to partial melting of the mantle above the slab triggered by fluids derived by dehydration of the slab. The fact that back-arc spreading in the Mariana arc system seems to be initiated through splitting of the volcanic arc and that basalts in the Mariana Trough (and in several other marginal basins) have transitional calc-alkaline–ocean-floor characteristics (Tarney *et al.* 1981; Wood *et al.* 1981; Saunders *et al.* 1980*b*) would appear to support this model.

Oxburgh & Parmentier (1977, 1978) have suggested instead that refractory Mg-rich depleted harzburgite in the subducting slab is inherently less dense and potentially more buoyant than that of surrounding more Fe-rich undepleted mantle. They suggest that this may segregate from the dense eclogite basaltic capping at depth and rise diapirically, providing a mechanism for the initiation of back-arc spreading. Such diapiric rise of lighter refractory mantle has been suggested as a means of building up a stable subcontinental tectosphere (Oxburgh & Parmentier 1978). Thermal calculations by Weaver & Tarney (1979) indicated that conditions for the uprise and melting of refractory diapirs might be more favourable in the Archaean. Nevertheless hydrous melting of more refractory harzburgite diapirs is potentially possible during the formation of modern and Phanerozoic basins. High-magnesian dykes and lavas having low contents of Ti and other incompatible trace elements, and with petrographic and major-element compositions similar to, or transitional towards, boninite or komatiite, have been described from several marginal basin ophiolites in Newfoundland (Gale 1973; Kidd 1977; Upadhyay 1978), Cyprus (Smewing & Potts 1976; Simonian & Gass 1978), Greece (Smith *et al.* 1975), Victoria, Australia (Crawford & Keays 1978; A. J. Crawford, personal communication), S Chile (Elthon 1979), Gorgona Island (Gansser *et al.* 1979) and elsewhere (Cameron *et al.* 1979). Many of these rocks have compositions that are more compatible with melting of more refractory sources than normal mantle and lend some support to the refractory diapir model. However such lavas have yet to be drilled or dredged from modern active basins, though they do occur in the fore-arc region of the Mariana–Bonin arc systems (Dietrich *et al.* 1978; Tarney *et al.* 1981).

Other authors have appealed to plate kinematics, age and geometry as important factors controlling marginal basin formation. Molnar & Atwater (1978) ascribed the lack of marginal basins in the eastern Pacific, compared with their frequent occurrence in the west Pacific, to the difference in age of the subducting Pacific ocean floor. That in the west Pacific is much older, cooler and more dense and sinks back steeply into the mantle with a strong vertical component, which may result in the trench moving oceanwards and permitting back-arc spreading. That in the eastern Pacific is younger, warmer and less dense and dips at a much shallower angle beneath the South American margin. Uyeda & Kanamori (1979) elaborated upon the differences between 'Mariana-type' and 'Chilean-type' interplate boundaries in terms of seismic activity, topography, gravity, volcanic activity and crustal movement. They suggested that there was a much tighter coupling between the upper and lower plates in the Chilean-type boundary, resulting in part from the westward motion of the overriding American plate; conversely the upper and lower plates are decoupled in the Mariana-type boundary. Thus, if the downgoing slab is anchored to the mantle and the landward plate is retreating, the potential for back-arc spreading exists. Chase (1978) has similarly argued that back-arc extension is strongly dependent on global plate kinematics.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

The opening of several marginal basins in the west and southwest Pacific (Shikoku, Parece Vela, S Fiji) may be linked to the change in direction of Pacific plate motion from northward to westward some 45 Ma ago (Jurdy 1979). The possibility arises that a new subduction zone may have been initiated at a major north trending transform at what is now the Kyushu–Palau Ridge about 42 Ma B.P. (late Eocene), which rapidly built up through the eruption of voluminous island-arc tholeiite in a few million years. Splitting of the arc followed after a period of 10 or 12 Ma to develop the Parece Vela Basin (30–16 Ma) and a new active arc, the West Mariana Ridge. The Mariana Trough (*ca.* 6 Ma B.P. – present) and the present active Mariana arc are both linked to a more recent phase of back-arc spreading; the initial frontal arc of the sundered Kyushu–Palau Ridge is represented by the Mariana fore-arc. It would appear from this sequence of events that back-arc spreading and marginal basin formation is a natural consequence of the initia tion of a new intraoceanic subduction zone, although not all marginal basins seem to have developed in this manner.

Table 1. Proterozoic greenstone belts

age/Ma	belt, location	reference
2100–1900	Birrimian–Tarkwaian groups, W Africa	Sillitoe (1979)
	Penokean succession, northern U.S.A.	van Schmus (1978); Cambray (1978)
1900	Richtersveld Province, SW Africa	Blignault (1974)
2000–1800	Tewings group, NE Australia	Wilson (1978)
1800–1700	greenstone belts, S Finland	Eskola (1963)
	leptite succession, Sweden	Löfgren (1979)
	Lynn Lake, central Canada	Zwanzig, H. V. *et al.* (1979)
	Flin Flon (Amisk), central Canada	Stauffer *et al.* (1976)
	Jerome (Yavapai), southwest U.S.A.	Norman (1977)
	Arizona, southwest U.S.A.	Anderson (1977)
	trans-Amazonian belts, Guiana	Choudhuri (personal communication)
1700–1600	Dalma, India	Gupta *et al.* (1980)
1300	Hastings area, eastern Canada	Moore (1977)
1350–900	Sinclair group, SW Africa	Watters (1976)
1000	Ife-Hesha (Kibaran belt), SW Nigeria	Klemm *et al.* (1979)

In summary, it appears that there are good arguments in support of several of the proposed mechanisms of marginal basin formation. Not all may be necessarily mutually exclusive. It is clear, however, that, except where there is a strong degree of coupling between the overriding and subducting plates, as in the Chilean-type boundary, there is a high probability of forming a marginal basin. Thus, if mantle convection was the primary mechanism for dissipating the Earth's internal heat in the Precambrian, there is a strong possibility that marginal basins may have formed in response to subduction. The extent to which such basins have been preserved may vary. Ensialic basins clearly stand a greater chance of survival.

Proterozoic greenstone belts

Greenstone belts typically contain an assemblage of mafic to felsic lavas and pyroclastics together with conglomerates, greywackes, sandstones, quartzites and shales. They have suffered only low-pressure greenschist-grade metamorphism and low-to-moderate deformation and are frequently bordered and intruded by granitic plutons. Greenstone belts are widely recognized in the Archaean and have traditionally been solely regarded as Archaean structures. However,

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

such belts are also found in the Proterozoic. They are diverse in type, but in this respect merely illustrate the variety and complexity of volcano-sedimentary basins formed at any one period in Earth history.

In the following account we outline some of the Proterozoic greenstone belts that have been described as such or interpreted in the literature as marginal basins, and whose features correspond broadly with modern marginal basins. These are listed in table 1. These early- and mid-Proterozoic belts provide a vital link between Archaean greenstone belts and the greenstone development in the late Proterozoic that can increasingly be related to Pan-African and Phanerozoic island arcs and back-arc basins.

The Jerome greenstone belt, Arizona. In the Jerome area, the Yavapai Series of mafic–felsic volcanics and tuffs has yielded zircon ages between 1770 ± 10 and 1820 ± 10 Ma (Anderson *et al.* 1971) and has the characteristics of an Archaean greenstone belt (Norman 1977). Passing upwards the sequence consists of pillow-bearing basalts, andesite and rhyolitic flows, tuffs, breccias, slates, jasper–magnetite beds and cherts. The sequence is intruded by quartz porphyry, quartz diorite and gabbro. Important Cu–Pb–Zn–Au–Ag mineralization is associated with these late intrusions (Anderson & Creasey 1958; Norman 1977).

Marquette supergroup, N Michigan. Van Schmus (1976) suggested that the Huronian supergroup (2200–2300 Ma B.P.) and the Marquette supergroup and Animikie group (1900–2000 Ma B.P.) of the Great Lakes region formed part of an early Proterozoic cratonic margin orogenic belt (Penokean orogeny, 1850–1900 Ma B.P.) in which a back-arc basin was the main depositional environment. More specifically the sedimentary environment of the Marquette supergroup, shallow epicontinental sea followed by subsiding restricted basins and finally thick gravity flows, was compared by Cambray (1978) with the pattern that accompanies rifting and growth of more recent marginal basins. He pointed out that the complete sedimentary and tectonic history is very similar to that found in a modern continental margin sequence associated with a developing ocean basin that was later involved in a collision-type orogeny.

Amisk greenstone belt, Manitoba. Although originally regarded as Archaean, more recent Rb–Sr age determinations on the volcanics, schists and plutons from this greenstone belt have yielded consistent ages of 1700 Ma with low initial $^{87}Sr/^{86}Sr$ ratios (Stauffer *et al.* 1975; Bell *et al.* 1975). The belt lies at the southern margin of the Churchill Province and consists of an upward sequence of pillowed and massive basalt flows, interlayered tholeiitic and calc-alkalic volcanics, calc-alkaline lavas and tuffs overlain by pelites and greywacke turbidites, all of which are overlain by clastic sediments of the Missi group. The sequence has been deformed and metamorphosed and is intruded by calc-alkaline gabbroic to granitic plutons. There are also earlier granitoid gneiss domes and hypabyssal intrusions possibly contemporaneous with the volcanic sequence. The belt is very similar to an Archaean greenstone belt, but developed almost 1000 Ma later than the main Superior Province greenstone belts.

Trans-Amazonian greenstone belt, Guiana. In the northern Guiana Shield, early Proterozoic greenstone belts extending for about 1500 km (A. Choudhuri, personal communication). The main lithologies are tholeiitic pillow lavas and spilites with oceanic affinity, basaltic and andesitic lavas, volcaniclastic greywackes, schists and quartzites. In central Guiana there are also water-lain tuffs (ferruginous cherts and graded greywackes) and in NW Guiana there are sediments enriched in Fe and Mn. The greenstones are intruded by quartz diorites, granodiorites and potassic granites that have Rb/Sr ages of 1810 ± 40 Ma. Choudhuri suggests that the combination of basic to acid volcanism, submarine extrusion and sedimentation is strongly indicative of a back-arc basin environment.

Sinclair group, SW Africa. This 1350–900 Ma volcanic–sedimentary–plutonic sequence occupies a marginal cratonic position parallel to the Damaran orogenic belt (Watters 1976). The sequence consists of basic, intermediate and felsic lavas together with immature clastic arkoses, conglomerates, grits, sandstones and shales, and is intruded by calc-alkaline plutons, which include gabbros, diorites, monzonites, syenites and granites. The belt has many features typical of Archaean greenstone belts, yet its similarities with modern continental basins suggest that it formed in a back-arc régime on the edge of a stable platform in connection with the development of the early Pan-African Damaran orogeny. The Sinclair group represents one possible Pan-African marginal basin. In view of the increasing evidence for the operation of the Wilson cycle in the formation of many Pan-African belts (Windley 1981; Gass 1977; Schmidt *et al.* 1979) it seems likely that many other late Proterozoic marginal basins will yet be defined.

Archaean greenstone belts

The thick, synclinal, volcano-sedimentary sequences that characterize Archaean low-grade terrains have frequently been regarded as the product of some unique geotectonic–thermal régime that was active during the early half of Earth history. On the other hand, Tarney *et al.* (1976) drew attention to the broad similarities between the 'rocas verdes' marginal basin ophiolite complex in southern Chile (Dalziel *et al.* 1974) and a typical Archaean greenstone belt, although pointing out differences in detail. Such analogies of course bear upon the problem of whether plate tectonic concepts can be applied to the Archaean and whether the similarities may be more fortuitous than real. Thick sedimentary successions with basin-like form do occur in the Phanerozoic but the volcanic component is rarely so prominent.

The essential issues centre around what caused the basins to form, why the volcanic component is so prominent and distinctive, why the belts (at least the younger ones) are multiple and what the relationship is between greenstone belts and Archaean high-grade terrains.

In most aspects the basin formation and the volcanicity can be ascribed to some form of mantle diapirism. The uprise of an active mantle diapir can cause crustal thinning and stretching (and in extreme cases actual rifting), but once this has been accomplished thermal decay of the diapir will lead to progressive subsidence of the basin over a period of perhaps 60 Ma, similar to the normal time–depth curve of oceanic crust or the subsidence of North Sea sedimentary basins (MacKenzie 1978). This would permit a thick series of *shallow water* sediments to accumulate in the basin; it seems to be required by the stratigraphy of most greenstone belts. The dominance of volcanics in the lower part of greenstone belt successions is therefore limited to the active phase of diapirism rather than the decay phase.

The cause of mantle diapirism is more problematical. Essentially diapirs rise because they are of lower density than the surrounding mantle. They can become less dense for two reasons (*a*) because they are hotter than the surrounding mantle, necessitating some means for generating excess temperature, or (*b*) because they are inherently (compositionally) of lower density, implying mantle heterogeneity, probably resulting from previous melting events. It is considerably easier to produce a density difference by change of composition than by increase of temperature; for instance extraction of basalt from pyrolite mantle, at depths where garnet and jadeiite are stable, leaves a refractory Mg-rich residue that is significantly less dense than

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

the pyrolite. Geochemical studies of the komatiitic and tholeiitic lavas of greenstone belts (see, for example: Hawkesworth & O'Nions 1977; Sun & Nesbitt 1978; Jahn *et al.* 1980; Nesbitt & Sun 1980) have suggested that they are probably derived from heterogeneous mantle sources. Whereas chondrite-normalized r.e.e. patterns for the tholeiitic lavas show light r.e.e. enrichment, those for the more Mg-rich lavas commonly show light r.e.e. depletion. Duncan & Green (1980) proposed that dynamic melting (Langmuir *et al.* 1977) could account for both tholeiitic and komatiitic lavas. In a rising diapir, tholeiitic melts are removed first (and not completely) and as the diapir continues to rise it eventually undergoes further melting to produce more magnesian komatiitic magmas from the more Mg-rich residue from which the first-stage tholeiitic melts had been generated. The difficulty with this mechanism is that in most greenstone belts it is the komatiitic lavas that are erupted first, i.e. the reverse of the sequence expected with dynamic melting.

Weaver & Tarney (1979) suggested that inherent density might be a more pertinent cause of diapirism because the fact that many komatiitic lavas show light r.e.e. depletion implies that their mantle source would also be low in other incompatible elements, the heat-producing U, Th, K and Rb included, and therefore unable to produce the temperature and density difference necessary for diapirism. They modelled the thermal budget of ascending Mg-rich diapirs and demonstrated that the requisite temperatures and degrees of melting to produce Mg-rich komatiitic magmas could easily be obtained. The thermal energy provided by such diapirs could also be dissipated in melting the surrounding undepleted mantle to yield tholeiitic magmas. This does then provide a means for generating lavas in the right time sequence, and also permits the repeated lava sequences of some greenstone belts to be produced by multiple diapirs. The main problem with such a mechanism lies in providing the Mg-rich mantle source essential for diapirism. Weaver & Tarney, following Oxburgh & Parmentier (1977), proposed that this might originate from the downgoing slab in a subduction zone. This of course implies a tectonic situation analogous to that of a marginal basin. Because such a mechanism depends more on the inherent properties of the diapir than on the kinematics of relative and absolute plate motions that govern the extension in modern back-arc basins, there is no necessity for actual rifting to take place. Most greenstone belts seem to have developed penecontemporaneously with the growth of sialic crust (McCulloch & Wasserburg 1978), and presumably before a significant subcontinental lithosphere (or tectosphere (T. H. Jordan 1977 and this symposium)) had developed. Hence it would have been easier for diapirs to penetrate this immature lithosphere. This may have happened at one or more locations in the back-arc region and need not necessarily have been constrained to one particular position as seems to be the case with modern marginal basins.

Most greenstone belts are invaded by tonalitic to granitic plutons, the compositions of which are broadly similar to those of continental margin Cordilleran belts except that they have r.e.e. patterns exhibiting heavy r.e.e. depletion (see, for example, Arth & Hanson 1975). Tonalitic to granitic plutons invade or are associated with the *rocas verdes* marginal basin complex in southern Chile (Dalziel *et al.* 1974) and were emplaced over a period of *ca.* 100 Ma after its formation. Moreover the continental basement in southern Chile, which pre-dates the development of the *rocas verdes* complex, includes foliated diorites, tonalites and granodiorites and metasediments, and is not unlike the gneissic basement of Archaean high-grade terrains. The formation of the *rocas verdes* rocks, which occurred during basin closure and uplift (Bruhn & Dalziel 1977) is similar to that shown by many greenstone belts (Tarney *et al.* 1976). The rocks within the basin retain the low-grade metamorphic mineral assemblages developed as a result of the hydrothermal activity during the spreading phase (Saunders *et al.* 1979).

In summary, comparisons can be drawn between the features of a typical Archaean greenstone belt and that of a modern ensialic marginal basin. There are of course many differences too, but it seems to us that these are not insuperable provided that allowance is made for the very different thermal régime operating during the Archaean.

Mineralization in marginal basins

Marginal basins provide a favourable site for ore deposits because the highly faulted oceanic crust that exists during the early stages of back-arc spreading provided fault-bounded troughs where low *Eh* and restricted seawater circulation provides an ideal environment for mineral deposition (Pearce & Gale 1977). This may also occur in continental margin back-arc rifts. Moreover, mineralizing fluids generated during back-arc mantle diapirism, enhanced perhaps by those derived from dehydration of the subducted slab, find early access into such basins. If marginal basins have formed throughout Earth history we might expect to find certain similarities in the type of mineralization. Here we briefly review the type of mineralization exhibited by greenstone belts of different ages.

Gold. 'Epigenetic gold deposits formed during Palaeozoic, Mesozoic and Cainozoic times are essentially similar in most respects to those of Precambrian age' (Boyle 1979). Thus gold similarly occurs in quartz veins in greenstones in Archaean belts such as Yellowknife, Val d'Or and Noranda in Canada and in the Barberton and Rhodesian belts in southern Africa, in the early Proterozoic belts at Jerome in Arizona (Anderson & Creasey 1958) and in the Birrimian of Ghana (Ntiamoah-Agyakwa 1979), in Palaeozoic marginal-basin ophiolitic greenstones at Buchan in Newfoundland, and in the Mesozoic Mother Lode in California.

Manganese. Mn deposits occur in chert–greenstone sequences ranging from Tertiary to Archaean in age (Shatskiy 1964; Snyder 1978). The younger cherts typically contain radiolaria, and banded red chert or jasper is common; some carbonate sediment is sometimes present. The altered pillow basalts and spilites are in places associated with ultramafic rocks and keratophyres. Commonly the stratiform Mn is associated with iron-rich deposits, and in older rocks with jaspilites or banded iron formations. It is generally accepted that the MnFe accumulations are of volcanic origin and were deposited by submarine hot springs at exhalative vents. Similar deposits have recently been discovered at the East Pacific Rise (Francheteau *et al.* 1979). The Phanerozoic chert–greenstone sequences of course belong to ophiolite complexes, many of which are thought to have formed in marginal basins. The Precambrian chert–greenstone sequences in greenstone belts do, however, have broadly the same stratigraphic associations as their younger counterparts.

Antimony–tungsten–mercury. Strata-bound stibnite–scheelite–cinnabar deposits are prominent in volcanic greenstones in the Penninic zone in the Alpine–Mediterranean belt, where they formed in early Palaeozoic back-arc marginal basins (Höll 1977). Very similar strata-bound Sb–W–Hg deposits occur in basic metavolcanics in the Bulawayan greenstone belts of Rhodesia (Cunningham *et al.* 1973). These authors also suggest that these are comparable to strata-bound tungsten mineralization in the 1950 Ma B.P. Piriwiri Formation of the Lomagundi system in Rhodesia and in the 500 Ma Pan-African Damara Formation of Namibia.

Sulphide deposits. Sulphide mineralization is a common feature of ophiolite complexes, many

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

of which represent marginal basin floor, and of Archaean greenstone belts. The incipient stage in the development of a continental marginal basin is manifest by cauldron collapse during extensive pyroclastic eruption, and this stage is associated with Kuroko-type massive sulphide mineralization (Sillitoe 1980). The Kuroko ores of Japan are Miocene in age. Similar massive sulphide deposits related to doming, cauldron subsidence and rhyolites or silicic pyroclastics occur in the Bathurst–Newcastle district of New Brunswick and the Lachlan greenstone belt of SE Australia (both Palaeozoic), in the Flin Flon and Snow Lake greenstone belts of Manitoba (Early Proterozoic) and in Archaean greenstone belts at Noranda, Quebec, and Mons Cupri, western Australia.

Massive sulphide deposits may form within basaltic lava sequences during two further stages of marginal basin development (Pearce & Gale 1977). First, Cyprus-type deposits, as in the ophiolite complexes of Troodos, Oman, and Betts Cove, Newfoundland, appear to form in the early stages of opening of the ocean basin; secondly the Lokken type, as exemplified by the deposits at Lokken, Norway, and York Harbour, Newfoundland, form in more mature basins. These types are distinguished by Pearce & Gale (1977) on the basis of the trace-element chemistry of their host lavas.

It is becoming increasingly apparent that the difference between Archaean greenstone belt sulphide mineralization and that represented by the more modern Kuroko- and Cyprus-type massive sulphides are small, and are probably related to subtle differences in the chemistry of the recycling sea water, the water depth and proximity to the exhalative centre (Large 1977; Plimer 1978).

Discussion

Although it is impractical, in a short review, to give a detailed account of all potential marginal-basin features throughout Earth history, there is no doubt that the same basic pattern of volcano-sedimentary sequence, deformation, plutonism and even mineralization can be recognized over a period of perhaps 3500 Ma. Of course it should not be expected that each marginal basin over this period will conform to a single mode of development. The great variety of form shown by modern marginal basins would make it difficult to define such a norm: extensional features at convergent plate boundaries range from simple back-arc faulting through basin formation with lithospheric necking to complete lithospheric rupture with active back-arc spreading. Likewise individual orogenic belts such as the Grenville, the Caledonian, the Hercynian and the Alpine, Himalayan and the Andean, each have very distinctive characteristics, yet all have developed through plate tectonic processes. Given that the Earth has evolved with time, both thermally and tectonically, there is no reason why back-arc extension should follow a strictly uniformitarian pattern. Nevertheless the fact that back-arc extensional basins are such a common feature at the present day and throughout the Phanerozoic makes it difficult to discount the probability that similar basins may have formed in the past.

There is a problem of preservation. All marginal basins develop at destructive plate margins and have a limited lifespan. Intraoceanic arcs and basins are particularly vulnerable to subduction and portions of such basins may only rarely be preserved as ophiolite complexes. These are valuable in marking the position of original sutures during continent–continent or continent–arc collision, but it is often difficult to accurately define the position of oceanic sutures, even in young orogenic belts. Clearly basins with a higher chance of being preserved are ensialic basins that have developed through back-arc crustal thinning rather than actual rifting.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
THE ROYAL SOCIETY A
PHILOSOPHICAL TRANSACTIONS OF

The absence of ophiolite complexes in the Archaean and early Proterozoic is often cited as evidence against plate tectonic processes such as subduction/obduction operating during this period. However, if most Phanerozoic ophiolites largely represent marginal basin floor rather than normal ocean floor and if greenstone belts are the ancient equivalent of modern marginal basins, then this difficulty is resolved. Blueschist belts, another modern indicator of subduction, are hardly likely to be preserved under the higher geothermal gradients of the Archaean and early Proterozoic. Present-day subduction zones may be either the sites of accretion of continental margin sediments, or, in other places, the sites of sediment subduction and continental recycling. There is little control on the relative importance of these two processes before the Phanerozoic. The supposed absence of accretionary prisms in the Archaean may be more apparent than real. In southern Chile for instance (unpublished observations) a massive accretionary wedge of probable late Palaeozoic age is invaded by, and deformed with, the tonalites and diorites of the Cordilleran batholith. The relationship resembles that of the tonalitic gneisses and associated high grade metasediments which is so characteristic of Archaean high-grade terrains. Given the higher rate of crustal generation in the Archaean it is likely that any such sedimentary prisms would be completely incorporated into the batholiths. The absence of large accretionary prisms does not therefore negate Archaean subduction.

In modern plate tectonics, back-arc extension can be regarded as an essentially passive effect strongly dependent on global plate kinematics (Chase 1978). The conditions for back-arc extension occur when the resultant velocities of the downgoing plate and the overriding plate in the global reference frame have a component directed away from the trench. This most commonly arises when the subducting lithosphere is old, thick and dense and sinks with a strong vertical component, producing seaward migration of the trench axis (Molnar & Atwater 1978; Uyeda & Kanamori 1979). Failure of the overriding lithosphere occurs along a line of weakness, which is inevitably the volcanic arc (Karig 1974; Tarney *et al.* 1981). The resultant back-arc diapirism, volcanism and accretion can be regarded as a subsidiary effect.

In the Archaean, however, the same basic process may have operated differently owing to the higher heat production. The lower mantle viscosity would permit diapirs to ascend more easily; indeed the more extensive basaltic volcanism and crustal growth would have produced greater localized thermal and compositional differences in the mantle, which might initiate diapirs (Weaver & Tarney 1979). On the other hand, plates would be smaller and thinner and have exerted less rigid control than at present on the sites of lithosphere penetration and rupture, although these would still be broadly constrained by relative and absolute plate kinematics. The net effect then is that diapirism becomes a much more important term in the overall balance of forces controlling back-arc activity. A consequence is that volcanism, resulting from active mantle diapirism, is an important feature of most Archaean greenstone belts whereas the extension is manifest more as crust–lithosphere thinning rather than as actual rupture. Diapirs may have been able to penetrate the immature subcontinental tectosphere much more easily than at the present day; hence volcano-sedimentary basins resulting from this activity make an important contribution to Archaean crustal development.

We thank K. C. Condie for discussion and information on Proterozoic greenstone belts and A. Choudhuri for a preprint.

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Discussion

A. Kröner (*Department of Geosciences, University of Mainz, F.R.G.*). If greenstone belts were generated in back-arc basins as you suggested here I see a major problem since back-arc spreading as understood today implies subduction nearby. This, again, implies the presence of fore-arcs and sutures as well as systematical lateral facies variations in greenstone belt terrains, none of which has been observed so far.

Also, one of the typical characteristics of greenstone volcanic successions is the occurrence, in one sequence, of typical ridge-type volcanics (in the modern sense) *and* typical island-arc volcanics. In modern back-arc basins these contrasting rock types do not occur by frequent interlayering, nor is there a single reported case where andesitic and bimodal volcanics are overlain by primitive tholeiite or komatiite as is the case in some greenstone belts, e.g. Vermilion belt, northeastern Minnesota (Schulz 1980).

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J. Tarney & B. F. Windley. It is a mistake to attempt to compare and contrast greenstone belts with modern intraoceanic marginal basins which have developed by back-arc spreading. One of the points that we have tried to stress is that back-arc extension in modern times is manifest in a whole variety of forms, from simple back-arc extensional faulting, through ensialic basins with or without associated basic volcanicity, to the back-arc extension with seafloor spreading such as is typified by the west Pacific marginal basins. We are not suggesting that Archaean greenstone belts represent marginal basins of the west Pacific type that have opened and closed, but rather than they are the ancient analogues of continental margin basins such as that already described from southern Chile, but that have developed through crustal thinning rather than crustal rifting. In this case the problem of fore-arcs and sutures does not arise since the basins are autochthonous and developed in regions of extensive crustal generation.

We would have thought that a tectonic environment similar to a marginal basin is almost ideal for the development of interlayered 'ocean-basalt' and 'island-arc' volcanic sequences, except where continued spreading has separated the spreading axis from the volcanic arc. Indeed, as stressed by Tarney *et al.*, basalts with arc-like geochemical signature are commonly developed during the earlier stages of back-arc spreading, and Wood *et al.* have described such interlayered sequences from the Mariana Trough.

References

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