Pre–1 Ga (pre-Rodinian) ophiolites: Their tectonic and environmental implications

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ABSTRACT

A global survey of sutures predating formation of the supercontinent Rodinia, i.e., older than ca. 0.9–1.0 Ga, has yielded evidence for at least 35 described and possible ophiolite complexes[1]. Results indicate that ophiolitic complexes older than 1.0 Ga mostly lack mantle tectonites below the magmatic rocks. Geochemically, these pre-Rodinian ophiolites display a mix of supra-subduction zone (SSZ), mid-oceanic-ridge, and oceanic-island compositions, with higher proportions of komatiite in complexes older than 1.6 Ga. Magmatic ages of identified and inferred complexes tend to cluster at times of 1.0–1.5 Ga, 1.8–2.3 Ga, ca. 2.5–2.7 Ga, and ca. 3.4 Ga. The data mostly support the hypothesis that magmatic oceanic crust was thicker prior to 1.0 Ga and thinned abruptly at 1.0 Ga. Mesozoic oceanic plateaus, e.g., the Caribbean, may provide an analogue to pre–1 Ga oceanic crust. A modified ophiolite columnar section to accommodate pre–1 Ga complexes includes a ''1972 Penrose Conference–type'' section in which the igneous rocks are overlain by a thick accumulation of interlayered sedimentary and volcanic rocks, as well as high-level intrusive complexes and plutons. Thinning of oceanic crust, the end of the anorthosite-anorogenic granite ''event' at 1000 Ma, and discontinuous hotspot activity through time suggest discontinuous magmatic evolutionary processes in Earth history, similar to other planets. Differences in thickness and ''subductability'' of oceanic crust may provide clues to the Archean-Proterozoic tectonic transition. The stratigraphic and sea-level effects of the inferred 1 Ga thinning of oceanic crust include widespread continental emergence, development of seasons, and increased atmospheric oxygen leading to the Phanerozoic.

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GSA Bulletin; January 2002; v. 114; no. 1; p. 80–95; 10 figures; 2 tables; Data Repository item 2002040.

Keywords: Archean, oceanic crust, ophiolite, Precambrian, Proterozoic, tectonics

INTRODUCTION

For the past century, ophiolites have been recognized as major components of orogenic belts (e.g., Steinmann, 1906, 1927; Hess, 1939, 1955). The recognition that ophiolites of the Tethyan belt and the Newfoundland Appalachians probably were preserved fragments of oceanic crust and mantle formed at spreading centers (e.g., Moores and Vine, 1971; Dewey and Bird, 1971; Coleman, 1971; Gass, 1968) contributed to the extension of the ''new global tectonics'' (Isacks et al., 1968) to the geologic record (e.g., Moores and MacGregor, 1972; Moores, 1970; Dickinson, 1971). Recognition in the 1970s that no oceanic crust older than ca. 200 Ma was preserved in the oceans and that ophiolites alone were the only possible source of information about pre–200 Ma spreading processes led Moores and Jackson (1974) to make a first attempt to correlate spreading rates of oceanic crust with ophiolitic internal structure. Consideration of the geologic environment led to a separation of ophiolites into ''Tethyan'' and ''Cordilleran'' types, based upon the presence or absence, respectively, of tectonically underlying continental mantle (Moores, 1982).

The Ophiolite Sequence

Since the plate-tectonic revolution, much attention has been focused upon the ideal ophiolite sequence, also called the ''Penrose Conference type''(Anonymous, 1972), which consists of the following (from bottom to top): tectonite ultramafic rocks that are presumably preserved remnants of the suboceanic lithosphere, mafic-ultramafic plutonic rocks, a sheeted-dike complex, extrusive pillow lava and/or massive lava flows, and overlying deep-sea and/or pelagic or, in some cases, volcaniclastic turbidite sedimentary rocks. Though this sequence has been thought to represent an analogue for the oceanic crust formed at spreading centers, there are problems with this analogue. First, the average thickness of the oceanic crust is ~7 km (White et al., 1992; Cloos, 1993); most ophiolite magmatic sections range from ~3 to 5 km thick. Second, many complexes do not display the complete sequence. These observations have led to much misunderstanding and hampered full use of ophiolites as a tool in historical geology. Recent recognition of incomplete magmatic sequences in the oceanic crust studied during the Ocean Drilling Program (ODP) and other investigations has led to a partial resolution of this dilemma. Fast-spreading centers or magma-rich centers, such as the East Pacific Rise or the Reykjanes Ridge generally display a complete ''Penrose Conference–like'' sequence (Karson, 1998, 2001; Macdonald, 1998), whereas slow-spreading, centers with variable magma supply, such as the Mid-Atlantic Ridge or the Southwest Indian Ridge may be incomplete, with the magmatic section actually absent in large areas (e.g., Lagabrielle and Cannat, 1990; Lagabrielle et al., 1998; Karson, 1998). Ophiolites possibly formed at magma-rich fast-spreading centers include the Semail complex (Oman), whereas sequences formed at slow-spreading centers include the Troodos (Cyprus), the Kizil Dag (Turkey), and the Apennine ophiolites (Italy) (see Fig. 1, A and B, respectively). Intermediate-spreading-rate complexes may include the early Paleozoic bodies in the Norwegian Caledonides (Dilek et al., 1998). Ophiolites inferred to represent transform-fault sequences commonly have mafic sections thin or practically absent, with in-folding of tectonites and cumulates and dikes cutting the ultramafic section (e.g., Dilek et al., 1991), whereas such dikes are rare or absent in typical ''spreading-center'' ophiolites. Many Cordilleran ophiolites characteristically display a more complex internal arrangement, with hypabyssal and extrusive

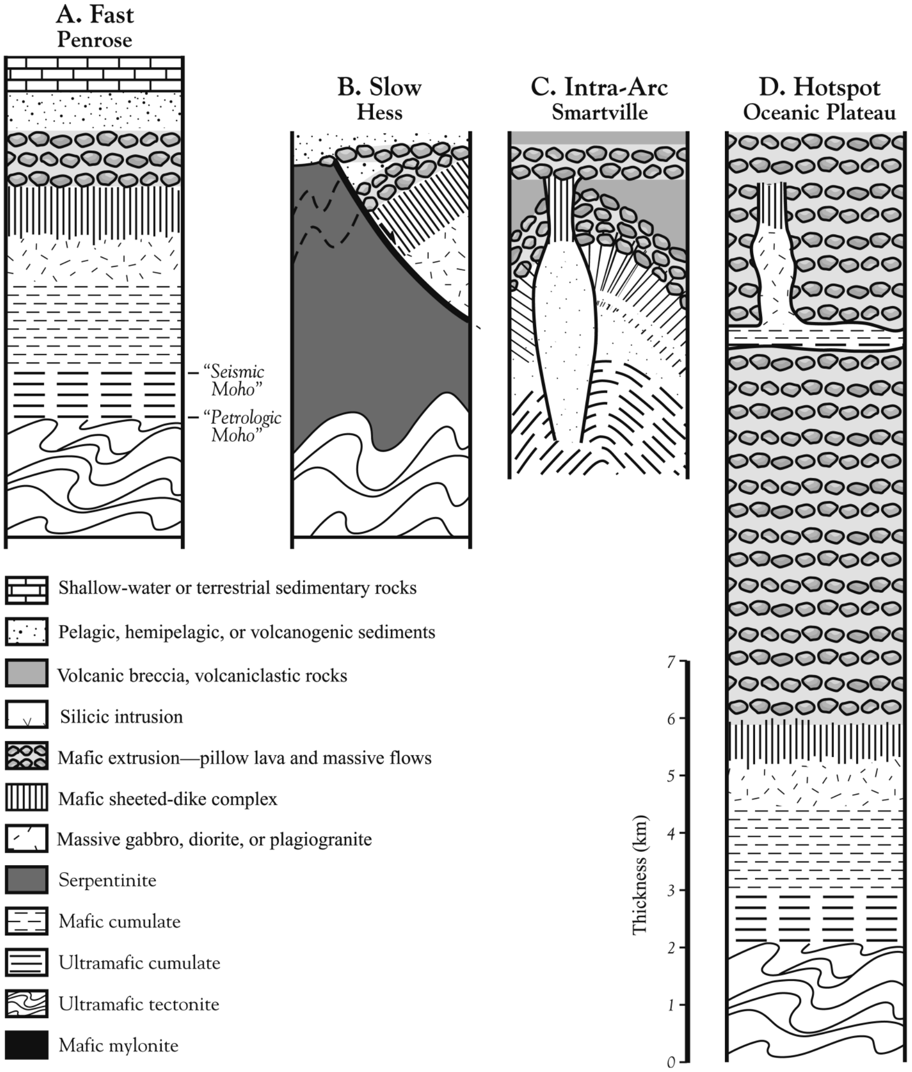

Figure 1. Schematic columnar sections of representative oceanic crust and corresponding ophiolite types. (A) Complete ophiolite sequence according to the ''Penrose Conference definition'' (Anonymous, 1972), characteristic of a magma-rich (generally fast-spreading) ridge. (B) Faulted, incomplete sequence characteristic of magma-starved spreading where tectonic processes dominate. Designated a ''Hess-type'' ophiolite from Hess's (1962) characterization of oceanic crust as serpentinized peridotite. (C) Complex composite section of oceanic island-arc sequences developed on oceanic crust. Designated ''Smartville-type'' for the Smartville Complex in the northwest Sierra Nevada, California (e.g., Dilek et al., 1991). (D) Possible hotspot (oceanic-plateau) section of oceanic crust. See text for discussion.

rocks intruded or deposited into or on an older and more deformed sequence of ocean crust, suggesting a polygenetic evolutionary history of these island-arc–related deposits (e.g., Dilek et al., 1991; Saleeby, 1992; Moores, 1998; Moores et al., 2001; see Fig. 1C herein). Hotspot (oceanic-plateau) sections might display a thick sequence of extrusive flows—possibly containing high-level differentiated intrusive sills—that was emplaced upon preexisting oceanic crust (Fig. 1D). Other evidence taken together suggests that in places, at least some of the lower oceanic crust worldwide is serpentinized mantle tectonite, as pointed out as early as 1971 by Vine and Hess.

Emplacement Mechanism

Debate over the question of emplacement of oceanic crust and mantle over continents has led to the realization that ophiolites were emplaced by obduction, i.e., subduction aborted because of the arrival of continental margins or island arcs at the trench (e.g., Coleman, 1971; Dewey, 1976; Moores, 1970, 1982). Cloos (1993) argued that normal-thickness oceanic crust should subduct without any difficulty. Most ophiolite complexes were emplaced within a few tens of million years of their formation (e.g., Hacker et al., 1996; Casey and Dewey, 1984; Moores, 1982; Nicolas, 1989). An important exception is the Troodos complex, Cyprus, which is undergoing emplacement now, some 90 m.y. after its formation.

Chemical Signatures

The chemistry of ophiolitic rocks has been also a subject of much debate and dispute (e.g., Miyashiro, 1973; Pearce and Cann, 1973; Robinson and Malpas, 1990) and led to the so-called ''ophiolite conundrum'' (e.g., Moores et al., 1984). Many ophiolites display compositions more reflective of modern arc and backarc environments than of most mid-ocean ridges. Recently, however, several mid-ocean ridge sites have been reported (Ross and Elthon, 1993, 1997; Leeman et al., 1990; Kamenetsky et al., 2001; Moores et al., 2001) that contain the chemical fingerprints of subduction-related sources or inferred continental lithosphere in mid-oceanic-ridge lavas. In addition, some modern arc complexes contain lavas reminiscent of mid-oceanic-ridge basalts (MORBs) and oceanic-island basalts (OIBs) (e.g., Leeman et al., 1990). Thus, mantle heterogeneity seems to be an important factor in both modern spreading centers and island arcs.

Many ophiolites with so-called subduction-related or ''supra-subduction zone'' (SSZ) sources lack any evidence of the existence of a well-developed arc edifice. Isotopic evidence for long-lived mantle heterogeneities and application of a new model of mantle structure led Moores et al. (2001) to propose a concept of ''historical contingency'' applied to the mantle source of a particular ophiolite's magmas. As defined, this concept posits that the composition of the mantle source is a function of the prior history of subduction and continental assembly and dispersal, rather than simply a reflection of the modern tectonic environment.

SIGNIFICANCE OF OPHIOLITES

In the light of the factors just cited, it is appropriate to restate briefly the current status of understanding of ophiolite complexes, because of widespread confusion about this issue:

1. Ophiolites represent oceanic crust and mantle formed at spreading centers and emplaced onto continental or island-arc margins. Possible settings for such centers in the mod-

TABLE 1. SUMMARY OF PRE–1000 MA (PRE-RODINIAN) OPHIOLITES

Complex and abbreviation	Magmatic age (Ma)	Ophiolitic lithologies	Sedimentary rocks	Comments	References
Baltica					
Jormua, Finland (JO)	1954 ± 2	Serpentinized peridotite, gabbro, plagiogranite, sheeted dikes, extrusives	Mafic hyaloclastics, carbonate, turbidites.	Similar to modern slow-spreading ridge and/or transform fault. ≈4 km thick	Kontinen (1987), Peltonen et al. (1996)
Laurentia					
Archean					
Slave province (SL)					
Cameron-Beaulieu	2574 ± 200 (Rb-Sr)	Ultramafic cumulates, gabbro, sheeted diabase, mafic-silicic volcanic rocks (including pillow lava)	Volcaniclastic rocks	Fault contact with other units.	Kusky (1990)
Yellowknife	ca. 2700	Gabbro, sheeted mafic dikes, pillow lavas and interflow sedimentary rocks	Tuffs, chert, iron formation	Basal contact unclear, probably tectonic. ≈12 km thick	Helmstaedt et al. (1986)
Pont Lake	ca. 2660–2667	Ultramafic cumulates, gabbro, anorthosite, sheeted dikes, pillow lavas, high-level sills	Iron formation, turbidites unconformable on top	Basal contact highly sheared, all rocks in series of thrust sheets. ≈7 km thick	Kusky (1991)
Wyoming					
Wind River (WR)	≥2630 ± 20	Ultramafic cumulates, gabbro, metadiabase dikes, pillow lavas	Mica schist and banded iron formation	Complexly deformed, dismembered	Harper (1985, 1986); Wilks and Harper (1997)
Proterozoic					
Trans-Hudson and correlatives					
Purtuniq (PU)	1998 ± 2	Ultramafic-mafic cumulates, gabbro, sheeted dikes, pillowed and/or massive extrusive rocks	Graphitic pelites and turbiditic sandstone	2 suites—lower MORB, upper OIB >5 km thick	Scott et al. (1991, 1992)
Flin Flon (FF)	1870–1920	Metavolcanic rocks	Basaltic turbidites	Metamorphosed mélange (?) MORB, OIB compositions. ≈3 km thick	Stern et al. (1995), Lucas et al. (1996)
New Québec (NQ)	1870–1880	Tholeiitic basalts and mafic-ultramafic sills	Basaltic turbidites	5–6 km thick. Interlayered lavas, sedimentary rocks, and layered intrusions	Skulski et al. (1993), Rohon et al. (1993)
Fox River (FT)	1882 ± 1.5	Mafic-ultramafic sills, volcanic rocks with interlayered sedimentary rocks	Quartz turbidites (?), dolomite, iron formation	300 km long, 15–20 km wide homocline. ~12–15 km thick	Baragar and Scoates (1981), Heaman et al. (1986)
Thompson nickel belt (FT)	2320 ± 20	Interlayered metasedimentary, metavolcanic, and ultramafic rocks	Quartz sandstones, pelites, iron formation, chert, graywacke	Metamorphosed and infolded with Archean migmatites	Baragar and Scoates (1981), Scoates (1981); Cumming et al. (1982)
Mazatzal-Yavapai block					
Payson ophiolite (PA)	1730	Mafic-silicic plutons, dike swarms, sheeted dikes, mafic-silicic submarine volcanic rocks	Dacite breccias with jasper lenses, thick turbidites	Chemically approximately like modern island-arc and backarc basins. 5–6 km thick	Dann (1991), Dann and Bowring (1997)
Prescott complex (PR)	ca. 1820	Serpentinized amphibole peridotite, gabbro, hypabyssal-extrusive mafic rocks	Turbidite sedimentary rocks	Multiply folded, amphibolite-facies metamorphism	Anderson and Blacet (1972), Anderson (1989), Moores (1976), Anderson et al. (1971)
Grenville orogenic belt					
Queensborough complex (QU)	≥1250	Cumulate peridotites, pyroxenite, gabbro, mafic dikes, pillow lavas	Pyroclastic rocks, silicic clastic metasedimentary rocks	Multiply deformed, amphibolite-facies metamorphism, mafic rocks ≈5 km thick	Smith and Harris (1996)
Coal Creek domain (CC)	1275–1326	Metamorphosed and sheared cumulates, gabbro, tonalite, amphibolite, mafic schist	?	Mafic rocks—low to medium-K tholeiites, similar to modern island arcs. ~5–10 km thick	Mosher (1998), Roback (1996)
Africa					
Archean					
Kalahari craton					
Pietersburg complex (PI)	3400(?)	Metamorphosed mafic-ultramafic cumulates, pillowed-massive tholeiites	Iron formation, clastic sedimentary rocks	At least 5–10 km thick	de Wit et al. (1992)
Jamestown ophiolite (JA)	≈3500	Deformed serpentinized dunite, gabbro, komatiite	Chert-shale sequence	Part of Barberton greenstone belt	de Wit et al. (1987), Sylvester et al. (1997)
Proterozoic					
Congo-Mozambique suture belt					
Kibaran complexes (KI)	>1400	Metamorphosed mafic-ultramafic cumulates, gabbro, amphibolitic sills	Metamorphosed turbidites, iron formation, quartzites, and calcareous rocks	Suture belt 350 km long, 10–35 km wide. Mafic-ultramafic rocks, mostly fault bounded	Burke et al. (1977), Deblond (1994), Deblond and Tack (1999), Rumvegeri (1991), Tack and Deblond (1990)
Kalahari-Congo					
Chewore ophiolite (CH)	1393 ± 22	Metamorphosed cumulates, gabbro, plagiogranite, sheeted dikes, basalt flows and pillow lavas	Metagraywackes and metapelites	100-km-long belt, geochemically similar to modern backarc basins. ≈1.5 km thick	Johnson and Oliver (2000), Oliver et al. (1998)

TABLE 1. (*Continued*)

Complex and abbreviation	Magmatic age (Ma)	Ophiolitic lithologies	Sedimentary rocks	Comments	References
Lurio Belt (LB)	≥1000	Metamorphosed ultramafic-mafic rocks, anorthosite.Multiply deformed, granulite-facies metamorphism	Laminated manganese-rich cherts	Multiply deformed, granulite facies metamorphism	Pinna et al. (1993), Sacchi et al.(1998, 1984)
Congo craton					
Mayombe (MA)	1000 ±	Metamorphosed and tectonized gabbro, dolerite, amphibolite.		≈5 km exposure on Congo River.	Vellutini et al. (1983).
West African Craton					
Kubongo (KU)	2100	Gabbro, sheeted dikes and sills, pillowed mafic-komatiitic lavas	Laminated manganese-rich cherts	Part of west African Birrimian greenstones. $\epsilon_{Nd} = +1.2$ to $+4.3$, Possible MORB-OIB mix. ≈1.5 km thick	Abouchami et al. (1990) Carlson (1993)
India					
Phulad ophiolite (PH)	1012 ± 78	Serpentinite, mafic-ultramafic amphibolite, intrusive and extrusive rocks		Part of Middle Proterozoic Delhi Supergroup, in Aravalli orogenic belt	Volpe and Macdougall (1990)
South America					
Pie de Palo (PP)	1118 ± 6	Dismembered mafic-ultramafic cumulates, sheeted dikes (?), and extrusive rocks	Metagraywacke	In Pampean Ranges, west Argentina	Vujovich and Kay (1998)
Niquelandia (NI)	1583 ± 25	Deformed ultramafic-mafic cumulates, anorthosite, extrusive rocks	Volcaniclastic rocks (?)	In central Brazil, in Goias region. Faulted sequence ≈15 km thick	Araujo and Nilson (1984), Danni et al. (1982), Danni and Kuijmjian (1984) Ferreira Filho et al. (1992, 1994)
North China					
Dongwanzi ophiolite (DO)	2505 ± 2.2	Harzburgite tectonite lenses, ultramafic-mafic cumulates, isotropic gabbro, sheeted dikes, pillow lavas	Interflow chert, banded iron formation and metapelite. Possibly overlain by thick-banded iron formation	Hebei province, 250 km northeast of Beijing. Mafic-ultramafic dikes cut the tectonite-cumulate boundaries. Resembles fracture-zone ophiolites in Phanerozoic	Kusky et al. (2001), Karson, (2001)
South China					
Anhui-Jiangxi complexes (AJ)	970	Anhui—intact metamorphic harzburgite, cumulate gabbro, spilite and keratophyre. Jiangxi—mélange with mafic-ultramafic units	Flysch	Sporadic blueschist. Major and trace elements similar to those of modern island arcs	Chen et al. (1991), Li et al. (1994, 1995, 1997)
Sibao ophiolite (SM)	1000–1100	Ophiolitic lithologies in mélange	Sheared flysch		Li (1993), Li et al. (1995)
Australia					
Archean					
Kalgoorlie (KA)	2600–3400	Sheetlike serpentinized dunite and dolerite, mafic-komatiitic pillow lavas	Volcaniclastic rocks	Yilgarn craton, near Kanowna Lake, Western Australia	Fripp and Jones (1997), Sylvester et al. (1997)
Norseman (NO)	2600–3400	Sheetlike mafic-ultramafic rocks, flaser-textured ultramafic rocks		Yilgarn craton, in and around Lake Cowan, Western Australia	Fripp and Jones (1997), Sylvester et al. (1997)
Proterozoic					
Narracoota-Trillbar complex (NA)	ca. 2000	Ultramafic-mafic intrusive rocks, sheeted dikes, pillow lava	Turbidites, chert, and minor carbonate	Capricorn orogen, between Yilgarn and Pilbara cratons. MORB-boninitic. 4–6-km-thick multiply deformed thrust sheets	Myers (1990c, 1989), Occhipinti and Myers (1999), Occhipinti et al. (1998), Pirajno and Occhipinti (2000, 1998), Hynes and Gee (1986)
Tickalara complex (TI)	1700–1900	Metamorphosed ultramafic-mafic cumulates, intrusive rocks and extrusive rocks	Chert, banded iron formations, turbidites, calcareous sedimentary rocks	Halls Creek orogen, southeast Kimberley block. Chemically similar to E-MORB and island-arc compositions	Sheppard et al. (1999), Hamlyn (1980), Griffin (1990)
Fraser complex (FR)	1300	Metamorphosed ultramafic, mafic, and silicic plutonic rocks	Quartzite, banded iron formation, pelitic rocks	Albany-Fraser orogen, southwest Australia; granulite-facies metamorphism; 30-km-thick thrust stack	Condie and Myers (1999), Myers (1995a, 1995b, 1990b, 1985)
Giles complex (GI)	1078 ± 3	Mafic-ultramafic plutonic rocks, possibly overlain by mafic-silicic metavolcanic rocks	Dolomite, conglomerate overlying volcanic rocks	Musgrave block, Arunta orogen, central Australia	Daniels (1967), Glikson et al. (1996). Goode and Moore (1975), Horwitz and Daniels (1967), Myers (1990a)
Antarctica					
Krylen intrusions (KR)	≤1700	Mafic-silicic intrusives and extrusives	?	In Dronning Maud Land	Bredell (1982)

Note: See Figure 2 caption regarding abbreviations.

ern oceans where oceanic crust is produced include mid-ocean ridges, pull-apart intra-arc basins, forearcs of incipient island arcs, and backarc basins.

2. Many ophiolites preserve oceanic spreading-center structure. Thus, they provide insight into the tectonics of modern spreading centers and the ''plumbing'' of thermal hot-spring deposits.

3. For oceans older than ca. 180 Ma, ophiolites represent the only exposed evidence of spreading history. Any idea of the processes that were active at pre–180 Ma spreading centers must come from ophiolites.

4. Well-preserved (intact, complete) ophiolites are emplaced (i.e., obducted; Coleman, 1971; Dewey, 1976) along mantle-rooted thrust faults that generally form by conversion of normal or strike-slip (transform) faults (originally parallel or perpendicular, respectively, to the spreading center) to mantle-rooted thrust faults during changes in plate motion (Dewey, 1976; Moores, 1982).

5. These mantle-rooted thrust faults are subduction zones, even if immature ones. Thus, ophiolite emplacement involves collision, i.e., aborted subduction, of an island arc or continental margin with an intraoceanic subduction zone. Such a collision predates and is not necessarily related to any subsequent continent-continent collision.

6. These ophiolite-emplacing thrusts faults thus represent fossil subduction zones; the better-studied sedimentary décollement fold-and-thrust belts present beneath ophiolite thrusts complexes are synthetic to subduction. They are secondary and subsidiary to the ophiolite thrusts.

7. Because they represent fossil subduction zones, ophiolite thrusts intrinsically cannot be balanced. Their displacements generally are large but unknown. Thus, ophiolite thrusts are of prime importance in the understanding of any orogen, a fact that is often overlooked, and their significance is commonly misunderstood and underrated.

8. Ophiolites have three important ages: (A) their spreading or magmatic age, (B) the age of their displacement or the age of the conversion of their associated ridge-parallel or ridge-perpendicular fault to an incipient subduction zone (thrust fault), and (C) their age of emplacement. The first age is given by the magmatic age of the rocks in question, although that age in complex sequences can vary by up to 30 m.y. or more. The stratigraphic age of rocks on top of or tectonically below also gives a minimum spreading age. The conversion age, as just defined, is difficult to determine; the age of metamorphism of the sole metamorphic slices may be the best indicator. The age of emplacement (i.e., obduction) is given by the age of deformation of the emplacing event. Dating of regional metamorphism of ophiolites and surrounding rocks gives a minimum emplacement age. For most complexes, only the magmatic age is available.

9. In the light of the emerging evidence for present mantle heterogeneity and its endurance through very long periods of time, the regional tectonic setting of an ophiolite complex cannot be determined on the basis of geochemical data alone. The regional geologic setting and the associated sedimentary rocks above and below an ophiolite complex also are crucially important ingredients for a positive interpretation. These factors need to be kept in mind when considering any statement of the tectonic environment of a given ophiolite based primarily upon geochemical data.

PRECAMBRIAN OPHIOLITES

The nature of Precambrian oceanic crust has been another subject of debate; see, e.g., Moores (1973, 1986, 1993), Burke et al. (1976), Sleep and Windley (1982), and Helmstaedt and Scott (1992). These reviews have suggested that pre–1 Ga magmatic oceanic crust was thicker than its post—1 Ga (Neoproterozoic and Phanerozoic) counterpart. Thermal considerations suggest that Archean (pre–2.5 Ga) oceanic crust was as much as 25–50 km thick (e.g., Sleep and Windley, 1982; Hoffman and Ranalli, 1988; Windley, 1995). Post–2.5 Ga to pre–1 Ga oceanic crust theoretically would have been intermediate in thickness between that of pre–2.5 Ga and post–1 Ga crust, thereby prompting the question of the nature of the transition from 25–50-km-thick Archean crust to 5–8-km-thick Neoproterozoic–Phanerozoic crust. The observation that mantle tectonites are generally lacking from most ophiolites older than 1 Ga led Moores (1986, 1993) to suggest that the magmatic oceanic crust was two to three times as thick so that the tectonite mantle was not preserved during emplacement (see also Burke, 1988). In addition, Moores (1986) proposed that the transition from thicker oceanic crust to the present 7 km thickness occurred rapidly at ca. 1 Ga.

In the 1990s, a great deal of new information has been published on the patterns of Precambrian continent assembly and fragmentation. It now seems clear that the continents were assembled in a supercontinent, named Rodinia, at ca. 1 Ga (e.g., Hoffman, 1991; Dalziel et al., 2000; Powell et al., 1993; Weil et al., 1998). This continent fragmented in Neoproterozoic time, resulting in the assembly of Gondwana and the separation of Baltica, Laurentia, and other continents that were separate entities until their reassembly in Paleozoic time, culminating in the formation of the well-known Pangea at the end of the Paleozoic.

Neoproterozoic ophiolites closely resemble their Phanerozoic counterparts (see summaries in Moores, 1986, 1993). The assembly of Rodinia thus corresponds in time approximately with the inferred sharp decrease in the thickness of oceanic crust. If this hypothesis is correct, the corollary is that most of the oceans that existed prior to the assembly of Rodinia would have had thicker crust. Thus, it is convenient for discussion purposes to call these pre–1 Ga complexes ''pre-Rodinian ophiolites.''

Identification of Ophiolites in Deformed Belts

One of the difficulties in the study of ophiolite complexes in deformed regions is that they are generally emplaced early, and they undergo considerable further deformation and alteration during subsequent orogenic developments. In post–1 Ga orogens, there are many complexes that have been well preserved, and the similarity between these complexes and exemplary Paleozoic–Mesozoic complexes, such as Troodos (Cyprus), Semail (Oman), and Bay of Islands (Newfoundland), is close enough to enable a fairly confident interpretation, even in metamorphosed regions. In pre–1 Ga orogens, however, there is the added complexity of the uncertainty of the nature of pre–1 Ga ophiolites, as already outlined, in addition to the widespread deformation and metamorphism of exposed rocks.

The question arises as to how to identify an ophiolite in a deformed belt. Features that support an interpretation of the existence of an ophiolite complex include the following: (1) part or all of an intact ophiolitic sequence; (2) faulted fragments that could represent disaggregated parts of an ophiolite; (3) sheeted dikes, especially when associated with other parts of a sequence; (4) serpentinite overlain depositionally or intruded by hypabyssal or extrusive rocks and/or overlain by marine sediments; (5) a fault zone or shear zone with lensoid fragments of possible ophiolitic affinity; (6) metamorphosed mafic and/or ultramafic rocks of possible ophiolitic parentage; (7) geophysical indications of high-density or highly magnetic rocks suggesting the presence within the continental crust of unserpentinized or partly serpentinized peridotite; and (8) overlying sedimentary rocks that indicate deep-water environments.

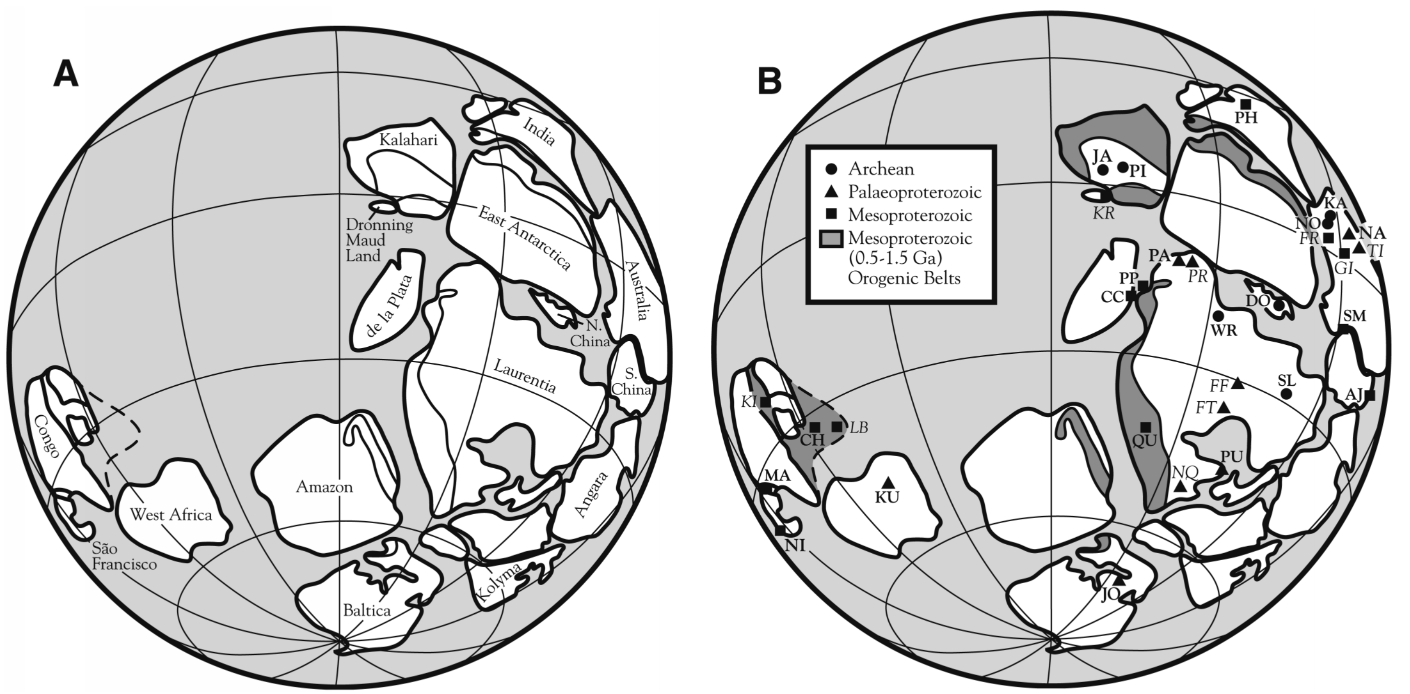

Figure 2. (A) Provisional map of the supercontinent Rodinia at 1 Ga. (B) Pre-Rodinian ophiolites and possible ophiolites. Nonitalic letters—ophiolites described in literature. Italic letters—inferred ophiolites. Squares—Mesoproterozoic ophiolites. Triangles—Paleoproterozoic ophiolites. Circles—Archean ophiolites. Map slightly modified from Pisarevsky and Powell (2000) through the use of the PLATES program of the University of Texas at Austin. Abbreviations and references as in Table 1.

It is worth keeping in mind that in the exemplary Troodos ophiolite (Cyprus), the contacts between the sheeted dikes and the overlying and underlying units are transitional. Near the upper contact, pillow-lava screens are present within the dikes, whereas near the lower contact, plutonic rocks are present as screens and dike rocks are present as xenoliths in plutonic rocks. Elsewhere in the same complex, the dike–plutonic rock contact is an oceanic fault (e.g., Moores and Vine; 1971; Moores et al., 1990).

In many cases, the exposures and the rocks are ambiguous. In such cases it is appropriate to ask, What else could the rocks represent? Possibilities include ophiolites, fossil island arcs, oceanic plateaus, stratiform mafic and ultramafic intrusions, intrusive and extrusive rocks associated with initial stages of rifting, mélange zones—either of convergent or transform margin origin, or subcontinental mantle slices. Each of these possibilities contains distinctive features that may be useful in differentiating one possibility from the other. Nevertheless, caution is essential when interpreting such ambiguous field exposures. The very ambiguity of these rocks underscores the need for the ‘‘continued vital role of geological field work in formulating Precambrian tectonic models’’ (Helmstaedt and Scott, 1992, p. 57).

Pre-Rodinian (Pre–1 Ga) Ophiolites

This section discusses the available information on previously recognized and heretofore unrecognized, but possible, pre-Rodinian ophiolites. Table 1 summarizes the main features of these complexes (more detailed discussions are available in the Repository item). Figure 2 shows the locations of these complexes plotted on a tentative Rodinia configuration at 1000 Ma (Pisarevsky and Powell, 2000). Figure 3 presents the identified and inferred complexes together with summaries of their approximate ages. Figure 4 portrays columnar sections of six representative pre–1 Ga ophiolitic sequences.

From the literature survey summarized in Table 1, 35 ophiolitic or possible ophiolitic complexes have been identified around the world. These complexes were emplaced during or prior to the assembly of Rodinia at ca. 1000 Ma. Examination of the list of designated and possible ophiolites outlined in Tables 1–2 and Figures 2–5 yields several observations:

1. Using the Rodinia fit of Pisarevsky (Fig. 2) causes some areal patterns emerge. Complexes ranging from 1.6 to 1 Ga (Mesoproterozoic) not surprisingly line up fairly systematically along 1–1.4 Ga orogenic belts in three main clusters, one extending from east Africa through to the São Francisco craton, another extending along the North American Grenville belt and connecting with the Rio de la Plata craton, and another extending from the South China block through south-central Australia to northern India. Complexes ranging from 2.3 to 1.7 Ga form a belt extending from Baltica through Canada into the Yavapai-Mazatzal area of the southwest US. This trend may be misleading, however, and would need to be plotted on a reconstruction for 1.7 Ga before drawing any conclusions. Ophiolitic rocks apparently are present in Archean terranes, as well. A few examples have been included in this study, and there may well be many other possible ophiolites elsewhere in Archean regions (e.g., Windley, 1995; Sylvester et al., 1997; Kusky and Vearncombe, 1997; Burke and Şengör, 1986).

2. Pre-Rodinian Proterozoic ophiolites apparently are roughly clustered in time (Fig. 3). Most Mesoproterozoic ophiolites are 1000–1400 Ma (with the obvious exception of Niquelandia), Paleoproterozoic complexes cluster in the 1650–2300 Ma interval, and Archean complexes cluster at ca. 2700 and 3400 Ma. It is interesting to speculate whether these age groupings are related to times of major oceanic subduction and consequent ophiolite emplacement events predating continent collisions. Because the reported ages are es-

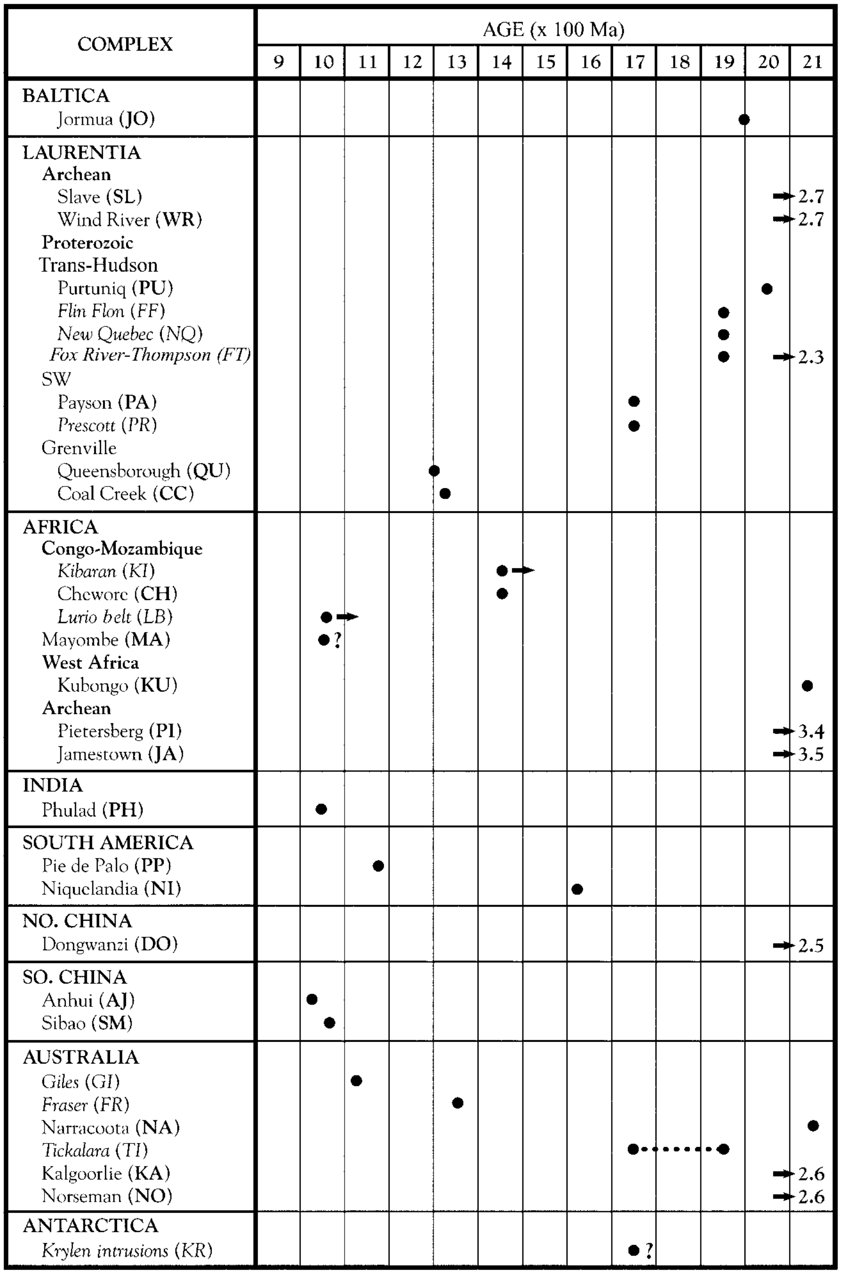

Figure 3. Ages of pre-Rodinian ophiolites and possible ophiolites organized by location. Abbreviations and references as in Table 1; italics and nonitalics as in Figure 2.

sentially all magmatic ages and because of the structural complexity and lack of detailed information, it is not yet possible to make any firm inferences on the geometry of plate interactions that led to the preservation of the pre–1 Ga ophiolite complexes, as is so often done for more recent ones.

3. The observations of pre-Rodinian ophiolites support the inferences of Moores (1993, 1986, 1973), Helmstaedt and Scott (1992), Burke et al. (1976), and Burke (1988) of thicker oceanic crust in the pre–1 Ga Earth. Mantle tectonites below cumulates are generally lacking in ophiolites older than 1000 Ma. Only the Jormua, Norseman, Jamestown, and Dongwanzi localities have rocks described as mantle tectonites. In Jormua, Jamestown, and Norseman, the rocks are highly serpentinized, some described textures are cumulate-like, and these rocks might also interpreted as cumulate ultramafic rocks deformed by ‘‘hot faults’’ (cf. Harkins et al., 1980). Jormua also apparently has extrusive rocks lying directly on serpentinites, implying synvolcanic faulting. Dongwanzi convincingly displays harzburgite tectonites, but as lenses possibly infolded within cumulates and cut by dikes, and thus is reminiscent of ‘‘fracture-zone’’ ophiolites, rather than ‘‘normal’’ sequences. In addition, its magmatic section is considerably thicker than that of any complex of this survey, except Niquelandia.

4. Table 2 and Figure 5 summarize the magma types that are present in pre-Rodinian ophiolites surveyed in this paper. Except for the increase in komatiites in rocks older than 1800 Ma, most rock types are reminiscent of those forming in the modern oceans. This fact suggests that the mantle source regions for ophiolites as old as 2.3 Ga or even older bore considerable similarity to Phanerozoic mantle sources, and that tectonic and magmatic processes in the Paleoproterozoic were accordingly comparable to those of today.

5. There is an increase in OIB volcanism in pre–1.6 Ga (Paleoproterozoic) ophiolitic complexes relative to subsequent times. This fact and the inferred greater thickness of pre–1 Ga oceanic crust suggest a greater role for hotspot activity, or at least greater magmatic activity at or near a spreading center, if pre–1 Ga ophiolites were emplaced within a few tens of million years after formation as were most Phanerozoic complexes.

This increased magmatic activity may resolve a dilemma posed by Davies (1992), who argued that increased heat production in the past would lead to faster plate motion, so that prior to 1 Ga, plates would not have sufficient negative buoyancy to sink, thereby precluding the operation of plate tectonics. Greater amounts of melting and consequently more intense magmatic activity would also lead to greater heat loss, thereby mitigating the need for faster plate motion. Furthermore, plate age is a factor only in slab-pull driving forces; other forces would not be affected by plate age, and young plates subduct in the modern oceans (e.g., Juan de Fuca, Cocos plates).

6. The presence of ophiolitic rocks in pre–1 Ga orogens argues against the formation of these orogens by ‘‘ensialic’’ processes. Rather the ophiolites’ presence indicates that continental fragmentation and reassembly, island-arc formation, seafloor spreading, ophiolite obduction, and arc-continent or continent-continent collisions characterized Proterozoic times, although with differences as already outlined herein.

7. Archean terranes pose interesting questions. Table 1 shows examples of suggested ophiolites from Archean terranes in North America, China, and Australia, and there no doubt are many other possible complexes around the world. Sleep and Windley (1982) argued from thermal considerations that Archean oceanic crust should be 25–50 km thick.

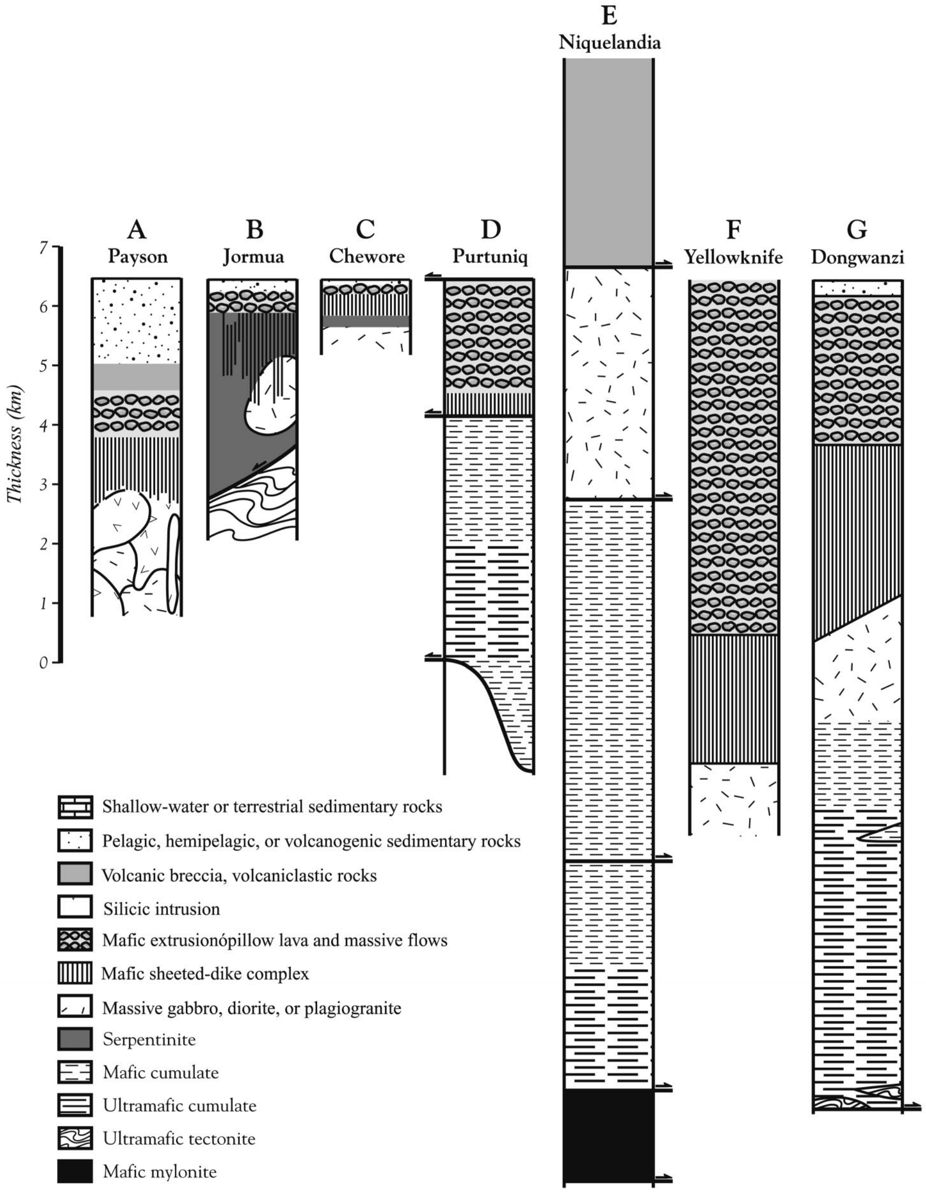

Figure 4. Cross sections of six pre-Rodinian ophiolites. (A) Payson ophiolite (after Dann, 1991). (B) Jormua complex (after Peltonen et al., 1996). (C) Chewore ophiolite (after Oliver et al., 1998). (D) Purtuniq ophiolite (after Scott et al., 1992). (E) Niquelandia complex (after Ferreira Filho et al., 1992). (F) Yellowknife complex (after Helmstaedt et al., 1986; Helmstaedt and Scott, 1992) (G) Dongwanzi complex (after Kusky et al., 2001).

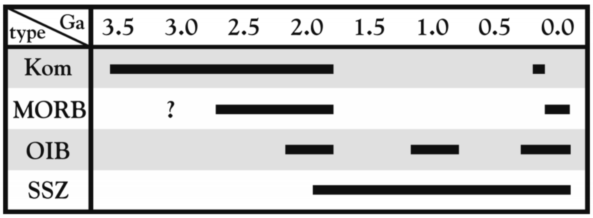

Figure 5. Summary of magma types for ophiolites and on-land volcanic provinces vs. time, modified in part after Moores et al. (2001). Abbreviations as in Figure 5.

Kusky and Vearncombe (1997) documented abundant low-angle faults in greenstone belts. Hoffman and Ranalli (1988) argued for a weak lower Archean oceanic crust, fostering the ‘‘delamination’’ of the crust during ‘‘oceanic flake’’ tectonics. Cloos (1993) suggested that oceanic crust thinner than ~17 km would subduct, but the arrival of thicker crust at a subduction zone would cause a plate reorganization, and the crust would be offscraped and would be added to the accretionary prism. These factors suggest a possible explanation of the Archean-Proterozoic transition, as outlined subsequently in this paper, although undoubtedly, Archean and Proterozoic oceanic crust was at least as complex as today’s (Karson, 2001).

TOWARD A REVISED OPHIOLITE DEFINITION

The observations so far presented suggest that some revision of the definition of an ophiolite is in order to account for the pre–1 Ga (pre-Rodinian) complexes discussed in this paper. This revision, summarized in Figure 6, suggests that a pre–1 Ga ophiolite is a pseudostratiform sequence that consists from bottom to top of the following:

The lowest part is composed of mantle tectonites, which mostly exhibit low-pressure compositions. However, high-pressure peridotites are present in some well-developed Phanerozoic ophiolites (e.g., the Yakuno ophiolite; Ishiwatari, 1985) and thus should not be a surprise in Proterozoic examples. So far, few if any mantle tectonites have been reported from pre–1 Ga terranes. The oldest reported harzburgite tectonite is in the 970 Ma Anhui-Jiangxi complexes, China (see Table 1).

The middle part of the ophiolitic sequence comprises mafic-ultramafic-silicic plutonic complexes; sheeted-dike complexes of ultramafic, mafic, or silicic composition; and extrusive ultramafic, mafic, or silicic volcanic rocks.

The uppermost rocks comprise sedimentary deposits of pelagic, chemical, or volcaniclastic origin, variably including chert, iron formation, carbonate, and quartzose turbidite (see Table 1). These are overlain by a thick succession of intercalated extrusive volcanic and sedimentary rocks, reminiscent of Phanerozoic ‘‘superplume’’ sites (e.g., Larson, 1991) or sedimented ridges (Morton et al., 1994).

In addition, a variable number of high-level, massive, stratiform plutons or limited-extent sheeted-dike complexes may be intruded within the ophiolitic igneous rocks. Some intrusions may even reach up into the sedimentary

TABLE 2. AGE AND/OR COMPOSITION RELATIONS OF MESO–PALEOPROTEROZOIC OPHIOLITES

Complex	Abbreviation	Magmatic age (Ga)	Tholeiite	MORB	OIB	SSZ-IA	Kom
Anhui/Jiangxi	AJ	0.972				•	
Phulad	PH	1.012		•		•	
Giles	*GI*	1.078				•	
Pie de Palo	PP	1.118				•	
Mayombe	MA	~1.000	•				
Lurio Belt	*LB*	≥1.000	•			•	
Coal Creek	CC	1.275				•	
Queensborough	QU	1.250				•	
Fraser	*FR*	1.300				•	
Chewore	CH	1.393				•	
Kibaran	*KI*	>1.400	•			•	
Niquelandia	NI	1.583	•				
Payson	PA	1.700				•	
Tickalara	*TI*	1.7–1.900				•	
New Quebec	*NQ*	1.880		•			
Thompson	*FT*	1.882	•				•
Flin Flon	*FF*	1.900		•	•	•	•
Purtuniq	PU	1.998		•	•		
Jormua	JO	1.954		•	•		
Narracoota	NA	2.000				•	
Kubongo	KU	~2.1		•	•		•
Fox River	*FT*	2.320	•				•
Dongwanzi	DO	2.504	?				
Pietersburg	PI	>2.7					•
Wind River	WR	2.63		•		•	
Slave	*SL*	2.6–2.7	•				•
Jamestown	JA	>3.3					•
Kalgoorlie	KA	>2.6					•
Norseman	NO	>2.6					•

Notes: Ophiolites, inferred ophiolites, and reported magma types are arranged in order of age. MORB—mid-oceanic-ridge basalt. OIB—oceanic-island basalt. SSZ-IA—supra-subduction zone or island arc. Kom—komatiite. Other abbreviations and references as in Table 1; italics and nonitalics as in Figure 2.

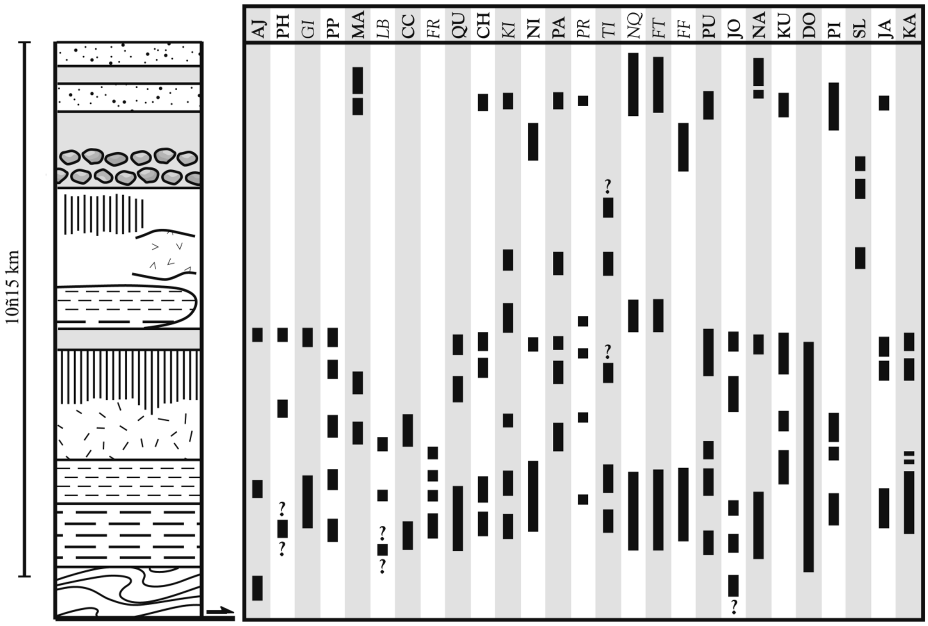

Figure 6. Possible revised ophiolite sequence for Paleoproterozoic–Mesoproterozoic time. Abbreviations and references as in Table 1; italics and nonitalics as in Figure 2.

and volcanic deposits at the top of the ophiolite.

Figure 6 shows an illustration of this hypothetical sequence, together with an estimate of the degree to which the complexes discussed in the previous section fit this expanded definition. As with pre–1 Ga complexes, ophiolites thus defined are present along sutures that represent disappeared oceans. They are interpreted to have formed at spreading centers in mid-ocean or backarc and intra-arc settings, and to have been emplaced in island arcs or on continental margins by the obduction, i.e., aborted subduction, of island-arc or continental-margin crust, as previously outlined herein. The tectonic setting is best indicated by the structure and stratigraphy of the ophiolite and associated rocks. The petrology and geochemistry of the ophiolitic rocks provide a comparison with modern tectonic environments and an indication of the mantle composition in the source region, but the composition is spatially heterogeneous and ''historically contingent'' and does not necessarily reflect a setting identical to modern ones.

Sheeted-dike complexes have been one of the key indicators of seafloor spreading recorded in ophiolites. As just outlined, however, many complexes lack sheeted dikes, and they seem to be lacking in slow-spreading centers. Thus, the presence of such a dike complex is not necessarily an argument against the presence of an ophiolite in pre-Rodinian settings. The formation of a sheeted-dike complex depends on the relative rates of magma supply versus extension. If the magma supply is less than extension, tectonic processes will dominate in creation of oceanic crust, and dikes will be rare to absent. Similarly if magma supply is greater than extension, the result will be a thick pile of extrusive and intrusive rocks; dikes may be present at various levels, but they will not be ubiquitous. If magma supply keeps pace with extension, dike complexes will be well developed.

POSSIBLE MODERN ANALOGUES: OCEANIC PLATEAUS AND THE CARIBBEAN

Oceanic plateaus possibly provide modern analogues to Paleoproterozoic–Mesoproterozoic oceanic crust. The Caribbean basin provides an excellent example. The geology of this basin, summarized by Burke (1988) and Kerr et al. (1997), exhibits many similarities to the hypothetical Paleoproterozoic–Mesoproterozoic oceanic crust already outlined. The oceanic crust of this basin, lying between North and South America, is thought to have been formed at a spreading center in the Pacific and subsequently overlain or intruded, after little time delay, by a thick volcanic-plutonic complex related to a hotspot, possibly the Galapagos. Remnants of this thick oceanic plate crop out in Central America, in the Greater Antilles, and in Venezuela and Colombia (Kerr et al., 1997) and may be present as far north as the thick ophiolitic basement of the Great Valley of California (Moores, 1998; Godfrey and Dilek, 2000). The Caribbean oceanic crust, summarized by Kerr et al. (1997), is thick and compositionally diverse;

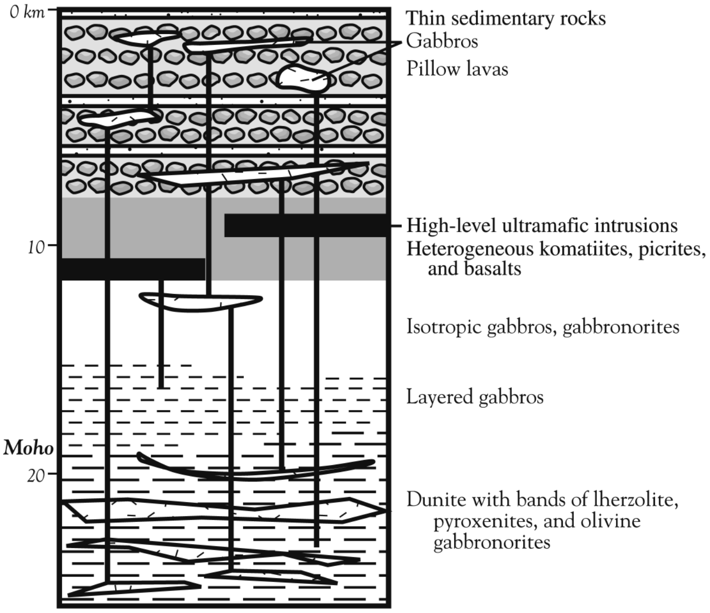

Figure 7. Hypothetical columnar section of Caribbean oceanic crust (modified after Fig. 11 of Kerr et al., 1997).

it consists chiefly of three main magmatic sequences: (1) tholeiitic basalt lavas sills with flat chondrite-normalized REE (rare earth element) patterns, similar to those of other Pacific oceanic plateaus, (2) a light-REE–depleted series of lavas and intrusions, including komatiitic lavas of Gorgona Island, Colombia, and (3) ''mildly alkalic (Kerr et al., 1997, p. 136) basalts. Minor felsitic lavas and intrusions are present in some places. Similar-aged island-arc volcanic and intrusive rocks are in fault contact with on-land exposures of the Caribbean oceanic-plateau rocks.

Figure 7 shows a composite section through the Caribbean crust, modified from Kerr et al. (1997). The section shows a complex series of intrusive and extrusive rocks of diverse compositions, with intercalated and overlying sediments, some 20 km thick. It is interesting to note that nowhere in the section do tectonite mantle rocks appear, a fact also noted by Burke (1988). It is also noteworthy how similar Figure 7 is to that derived separately from the survey of pre-Rodinian complexes (Fig. 6).

THICKNESS OF THE OCEANIC CRUST THROUGH TIME

The data reported from this survey generally support the hypothesis that pre–1 Ga oceanic crust was thicker than Phanerozoic crust. Figure 8 schematically depicts these and other relationships. The plot shows an ~7-km-thick oceanic crust extending from the present back to ca. 1 Ga. The inferred positions of the Anhui-Jiangxi and Pie de Palo complexes are shown. Also shown are the ~10 km maximum thickness of Phanerozoic ophiolite thrusts (Moores, 1986, 1993), the 17 km maximum thickness of subductable oceanic crust (Cloos, 1993), and the 17–20 km inferred thickness of Archean oceanic ''flakes'' (Hoffman and Ranalli, 1988). The limits on the steplike change in thickness of oceanic crust are set by the Anhui-Jiangxi and Pie de Palo complexes, as indicated in Figure 8. The thickness and change in thickness in oceanic crust prior to 1 Ga are clearly speculative. The constraints are the postulated 1 Ga change in thickness and the theoretical considerations indicating that Archean oceanic crust was at least 20 km thick (Sleep and Windley, 1982; Hoffman and Ranalli, 1988). Figure 8 shows a range of possibilities for 2.5–1.0 Ga oceanic crust. The upper boundary of the shaded region assumes a steplike increase in thickness from 7 to 15 km at 1.0 Ga and a linear increase through time to a thickness of 25 km at 2.5 Ga. The lower boundary of the shaded region assumes a steplike increase from 7 to 11 km, a constant ~11 km thickness of oceanic crust from 1.1 Ga to 2.5 Ga, and then an abrupt increase to a thickness of >25 km. These two curves may bound the range of possibilities for pre–1 Ga Proterozoic thicknesses of oceanic crust. I suspect that most Paleoproterozoic–Mesoproterozoic oceanic crust would lie within the lower part of the shaded region of Figure 8, but some, such as Jormua, may have been sufficiently magma starved to produce oceanic crust thin enough to preserve tectonites.

Neoproterozoic Oceanic Crustal Thinning

In two earlier papers (Moores, 1986, 1993) I suggested an abrupt thinning of oceanic crust at ca. 1000 Ma to explain the lack of preservation of mantle tectonites in ophiolites prior to that time. Although exceptions to this general rule may exist (as discussed in the foregoing; see Table 1), the general observation, summarized in Figure 6, is that mantle tectonites are not present in ophiolites older than 1000 Ma, nor are they preserved in Phanerozoic exposures of oceanic plateaus, as already discussed herein.

The abruptness of this hypothetical change is an important issue. One can get some idea from the ages of complexes on both sides of the ''divide'' between the ''with tectonite'' and ''without tectonite'' complexes. The oldest Proterozoic complexes apparently with tectonite are the Anhui-Jiangxi complexes at ca. 970 Ma. The youngest complex on the other side of the ''divide'' is the Pie de Palo, with an age of 1050–1118 Ma. These ages thus bracket the steplike change in oceanic crust as occurring between 970 and 1050 Ma.

Ophiolites and the Archean-Proterozoic Boundary

The relationships illustrated in Figure 8 also emphasize possible differences in Archean, Paleoproterozoic–Mesoproterozoic (2.5–1.0 Ga), and post–1 Ga tectonics. These differences are summarized in Figure 9, which shows a scenario of two subduction zones for Neoproterozoic–Phanerozoic (1–0 Ga; Fig. 9, A and B), Paleoproterozoic–Mesoproterozoic (2.5–1.0 Ga; Fig. 9, C and D), and Archean (>2.5 Ga; Fig. 9, E and F) times. For the Archean, Hoffman and Ranalli (1988) argued that 10–20-km-thick ''flakes'' of oceanic crust would be produced during subduction of a thick oceanic crust because of a weak zone at midcrustal levels. However, if the thickness of these flakes exceeded 17 km, they would not subduct, according to Cloos (1993). Subduction of oceanic crust thus would be characterized by accumulation of these flakes along the edges of the overriding plate. In this scenario, Archean accretionary prisms would be wide and contain multiple thrust slices of the upper oceanic crust (see Fig. 9E); in other words, these thrust slices are the fault-rich (Kusky and Vearncombe, 1997) ''greenstone belts'' of Archean terranes. Because the addition of each thrust slice would be a ''collision'' in the sense of Cloos (1993), i.e., deforming both

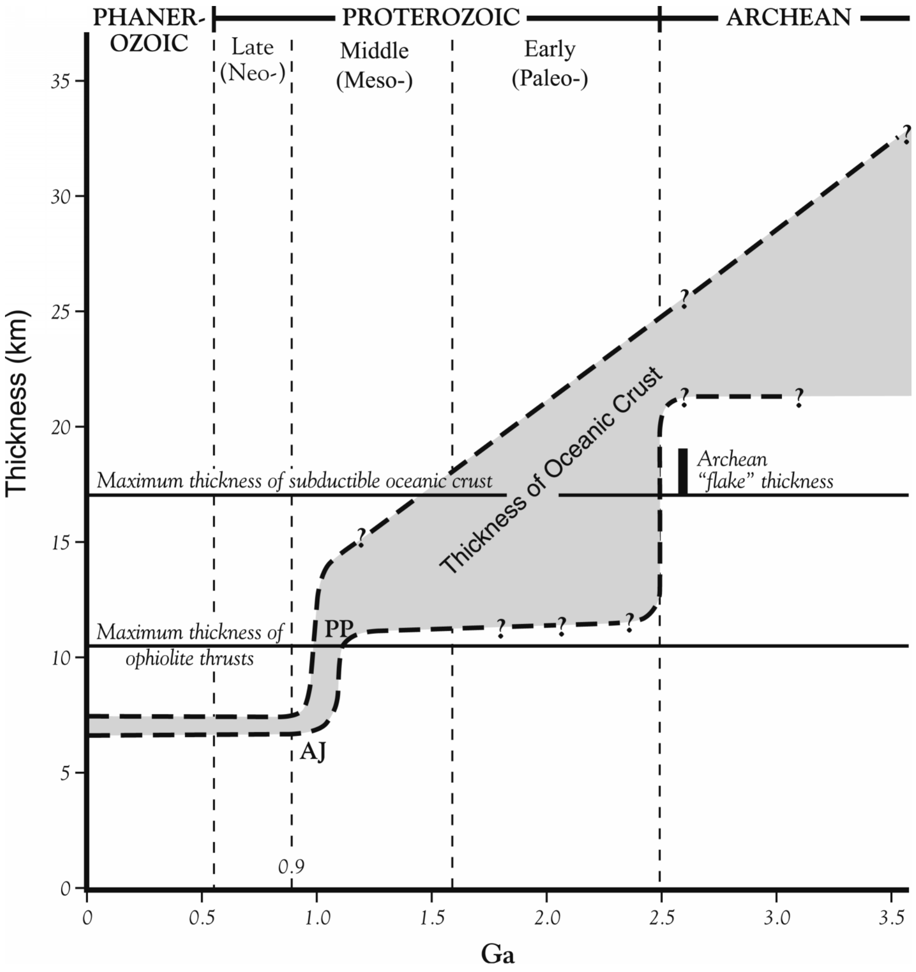

Figure 8. Hypothetical plot of inferred thickness of the oceanic crust vs. time. Modified after Moores (1993, Fig. 1, p. 6). AJ and PP symbols for Anhui-Jiangxi and Pie de Palo complexes, respectively. Maximum thickness of subductable oceanic crust after Cloos (1993). Thickness of Archean ''flakes'' after Hoffman and Ranalli (1988).

colliding blocks and forcing a change in relative plate motion, multiple deformations would be ubiquitous and might have led to preservation of small amounts of mantle tectonite, especially in relatively thinner crust, as in Dongwanzi, Jamestown, or Norseman. Emplacement or ''obduction'' of oceanic crust during the Archean by collisional processes might also result in detachment along mid-crustal levels, a possibility also noted by Harper (1985, 1986). For 1.5–2.5 Ga oceanic crust, the oceanic crust complexes would have been thin enough to be subducted normally (Fig. 9C, left subduction zone), but the ''obduction'' or collisional emplacement of ophiolites would generally have preserved the magmatic oceanic crust and not the tectonite mantle (Fig. 9D).

Thus, the Archean-Proterozoic boundary may have been a tectonic change occasioned by thinning of the oceanic crust to a level where the entire thickness could be subducted, resulting in narrower accretionary prisms, as illustrated in the difference between Figure 9E and Figure 9C. The subducted Archean mantle and lower crust may well be preserved beneath Archean cratons as the depleted, less dense mantle documented by O'Reilly et al. (2001).

ENVIRONMENTAL IMPLICATIONS

A widespread oceanic crust thicker than the 7 km thickness of modern oceanic crust has environmental implications, as well. This model implies that the contrast between the continental elevation and the ocean-floor elevation—i.e., the ''freeboard'' as used by Moores (1993)—would have been less, and the continents therefore would have been mostly submerged until the Neoproterozoic. Data on the stratigraphic record on various continents (Eriksson et al., 1999) and the existence of submarine volcanic rocks (mafic-ultramafic pillow lavas) erupted through continental crust (Arndt, 1999) supports this hypothesis.

Figure 10 summarizes several environmental, igneous, stratigraphic, tectonic, oceanic, and atmospheric events that may have happened in the Proterozoic (between ca. 2.5 and 0.5 Ga). This diagram, a modification and update of Figure 3 of Moores (1993), explores some of the consequences of the postulated thinning of oceanic crust at 1.0 Ga. As outlined previously (Moores, 1986, 1993), an oceanic crust two to three times as thick as at present would probably mean that the continent-ocean elevation contrast would decrease from the present 5.5 km to 2–3 km (3.8 km according to Cloos, 1993). Modern analogues of such an oceanic crust include the oceanic plateaus—such as the Alaska-sized Ontong-Java Plateau, lying at ~2–3 km depth, but rising to a depth of only 1700 m in its central part (e.g., Neal et al., 1997)—or as discussed earlier, the Caribbean Sea, the surface of which ranges from depths of <500 m to 5 km, but which averages 2–3 km in depth (e.g., Nierenberg et al., 1977).

It is important to keep in mind that the thickness of continental crust and the thickness of oceanic crust result from two mostly different processes. Oceanic crust formation is a combination of magmatic and tectonic processes; which process dominates is determined by the amount of magma that is formed during divergent or mid-plate events. The thickness of continental crust, on the other hand, is a mostly tectonic process, primarily controlled by accretion at subduction zones and arc-continent or continent-continent collisions, as well as by the strength of continental crustal rocks. Though operating, the contribution of magmatic processes to the thickening of continental crust is arguably subordinate to tectonic processes.

Simple isostatic balance requires that the equilibrium elevation difference or ''freeboard'' between average continents and seafloor would be expected to decrease if oceanic crust were thicker. Both an increase in spreading rate and a younger average age for oceanic lithosphere would have the same effect (Davies, 1992). Moores (1986, 1993) suggested that as much as 95% of Earth's surface may have been covered with the oceans, if the same relative volume of continental crust and ocean water as at present is assumed.

This amount of global flooding may seem alarming at first, particularly in view of the prominent presence of subaerial sediments in many Precambrian regions. It is important to keep in mind, however, that 5% of Earth's surface is approximately equivalent to the size of

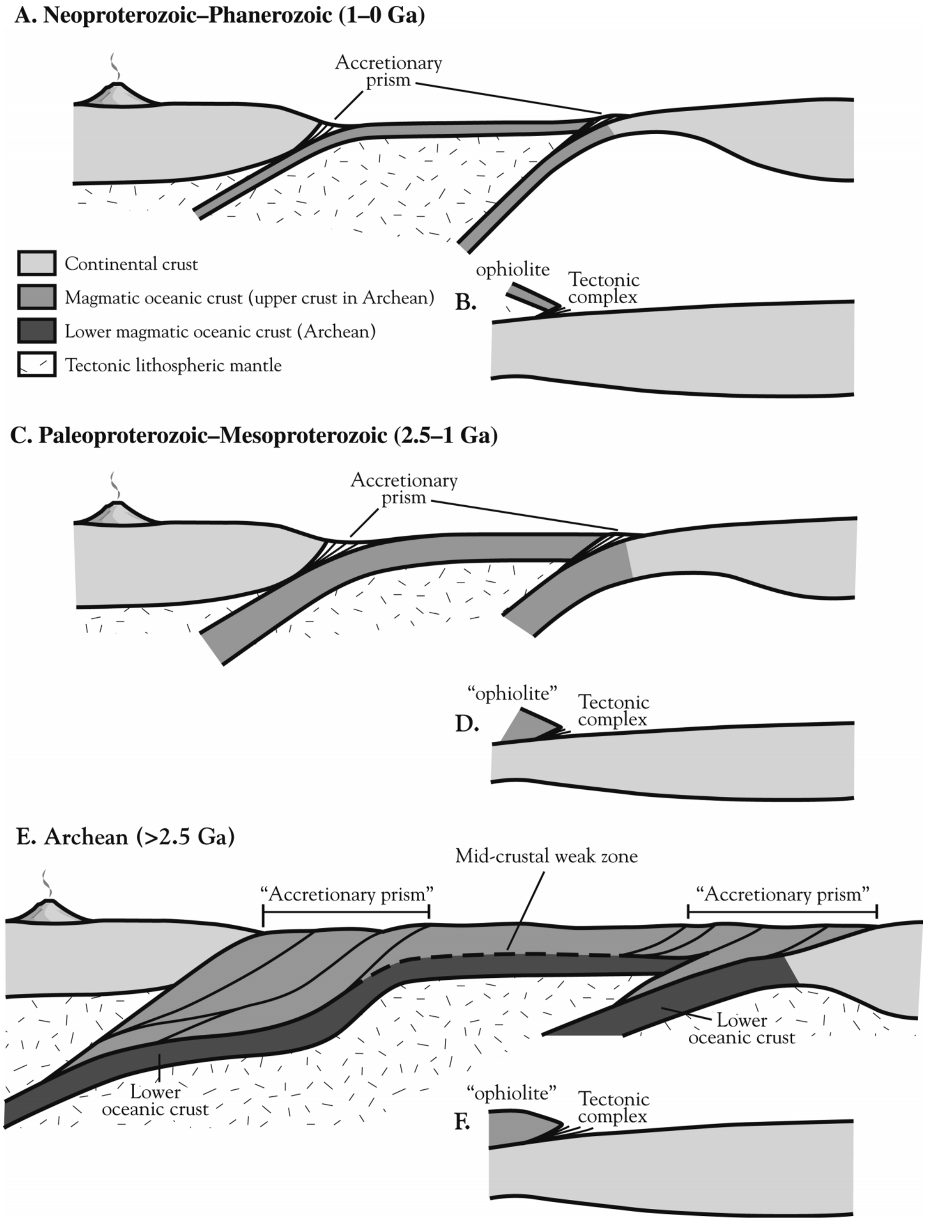

Figure 9. Schematic diagrams showing different oceanic crust and continental crust interactions for three intervals of time. (A) Neoproterozoic–Phanerozoic (1–0 Ga). Oceanic crust ≤7 km thick. Two subduction zones shown—left: conventional oceanic crust and mantle subducting beneath continental margin; right: immature subduction zone in process of collision with continental margin. (B) Scenario on right subduction zone in A after collision and ophiolite emplacement. Ophiolite contains both magmatic crust and underlying mantle tectonite. (C) Paleoproterozoic–Mesoproterozoic (2.5–1.0 Ga). Oceanic crust is ~15 km thick, less than the 17 km maximum subductable thickness of Cloos (1996). Both subduction zones similar to those in A except for thickness of oceanic crust. (D) Scenario on right subduction zone in C after collision and "ophiolite" emplacement. Preserved remnant is only part of magmatic oceanic crust; tectonite mantle is not preserved. (E) Archean (≥2.5 Ga). Oceanic crust ~25 km thick, too thick to be subducted intact, but containing midcrustal weak zone of Hoffman and Ranalli (1988). Both accretionary prisms now show evidence of wide areas of off-scraped upper oceanic crust (Hoffman and Ranalli, 1988). Only lower oceanic crust subducts with mantle. (F) Scenario on right subduction zone in E after collision and preservation of "ophiolite," derived only from upper oceanic crust.

the North American continent! Thus, there should have been be ample dry land on which to form continental deposits.

Possible perturbations of the continental "freeboard" are also shown schematically in Figure 10. These temporary changes may have resulted from major continent-continent collision events that led to the formation of supercontinents, such as Nuna (Hoffman, 1997) and Rodinia. The resultant orogens might have looked more like the mountainous islands of southeast Asia and the southwest Pacific—e.g., the highlands of New Guinea—than the Himalaya. Another event that would decrease continental flooding would be the occurrence of continental glaciations, as shown in Figure 10 (Knoll, 1991; Eriksson et al., 1999). Other developments and their timing shown in the diagram include iron formations (Barley and Groves, 1992; Knoll, 1991), anorogenic granites and anorthosites (Anderson, 1983), hotspots (Moores et al., 2000), dike swarms (Ernst et al., 1996), komatiites (this work), atmospheric oxygen (Rye and Holland, 1998), and oceanic circulation effects and evolutionary events (Brasier and Lindsay, 1998).

Although times of major glaciation and mountain formation, such as continental rifting or orogenic activity, would have been times of relative continental emergence, the hypothesis presented herein implies that the Neoproterozoic would have been the first time of major and permanent continental emergence. This time, called the "Origin of Seasons" in Figure 10, would have seen the first onset of widespread sustained continentality and resultant seasons (see Moores, 1993). It is interesting that this time corresponds with the rise of oxygen to modern levels (Rye and Holland, 1998). As discussed in an earlier paper (Moores, 1993), the increased seasonality that resulted from the decrease in thickness in the oceanic crust, increase in continental "freeboard," and increase in oceanic-basin volume would have resulted in greater erosion and volume of sedimentary deposition, as implied in Figure 10. Increased erosion of itself would be expected to decrease atmospheric oxygen (Brasier and Lindsay, 1998). But increased oceanic circulation, as also indicated by Brasier and Lindsay, would have resulted in greater biomass productivity and, together with the increased sedimentation, would have resulted in sequestering of carbon in sediments. Carbon isotope data support the increased burial of organic carbon at this time (Kaufman and Knoll, 1995). This biogenic fixation of carbon would have overwhelmed the uptake of oxygen from greater weathering, leading to the rise in atmospheric oxygen documented by Rye and Holland (1998) and mak-

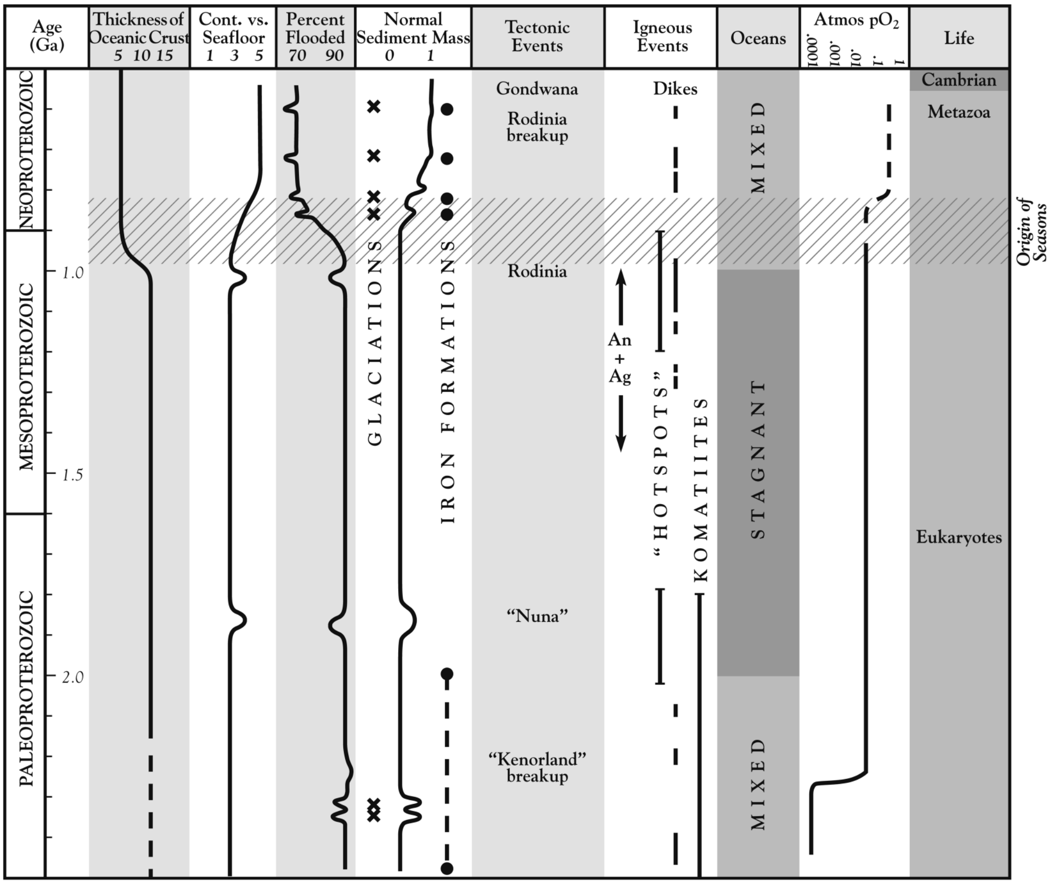

Figure 10. Diagram showing possible thickness of oceanic crust through time and environmental, tectonic, and igneous events. Modified after Moores (1993). Features depicted in columns: "Cont. vs. Seafloor"—continent-seafloor elevation contrast ("continental freeboard") in kilometers. Percent of Earth's surface flooded—estimates from Moores (1993, 1986). Norm. sed. mass—normalized sedimentary mass, i.e., the mass of terrigenous sedimentary rocks divided by the present mass, after Hargraves (1976) and in Moores (1993). Tectonic events include "Nuna" formation and "Kenorland" breakup after Hoffman (1997). Igneous events include An + Ag, i.e., anorthosite and anorogenic granite intrusions, after Anderson (1983, his Fig. 6); and dikes after Ernst et al. (1996). Oceans column modified after Brasier and Lindsay (1998, their Fig. 1). Atmospheric *p*O_2 after Rye and Holland (1998). Life—events after Brasier and Lindsay (1998, their Fig. 1). See text for discussion.

ing possible the development of animals, as outlined by Knoll (1991).

D.U. Wise (*in* Wise and Moores, 1994) criticized the hypothesis of greater continental "freeboard," arguing that (1) ophiolites are unreliable indicators of thickness of the oceanic crust; (2) the hypothesis "violates the master principle of tectonics," meaning the constancy of continental freeboard (as defined by Wise, this term indicates the average height of continents above sea level, rather than the continent-ocean elevation contrast as used here); and (3) the hypothesis requires a more static Earth. With regard to ophiolites, it is true that they are commonly affected by deformation, even multiple deformations. Many bodies exhibit, however, deformation features that are demonstrably seafloor features. Furthermore, it is not the preserved thickness of ophiolites that engenders the abrupt thinning model—it is the abrupt lack of preservation of mantle tectonite in complexes older than ca. 1000 Ma. Ophiolites remain a good indicator of spreading processes, and for times prior to 200 Ma, they represent the only record that exists.

With regard to Wise's constant "freeboard" hypothesis, as mentioned in the preceding paragraph, there is nothing magic about the constancy of the continent-ocean elevation contrast, because the thickness of the crust of these two realms arises from two different processes, mainly tectonic and mainly magmatic, respectively. Finally, the hypothesis presented in this paper implies just the opposite of a "static" Earth and proposes that tectonic and magmatic processes and products at times prior to 1000 Ma differed from those subsequently.

The current hypothesis also resolves a dilemma that concerned Rodgers et al. (1984), who pointed to widespread epicontinental marine basins in Proterozoic time in areas mostly emergent in the Phanerozoic. Rodgers et al. (1984) argued for a thickening of the continental crust by magmatic underplating to explain the stratigraphic observations, although they acknowledged that there was little direct evidence for such a process. The hypothesis of increase in ocean-basin volume by thinning of the oceanic crust and increasing the contrast between the continent and ocean-floor elevations or "freeboard" resolves this issue.

CONCLUSION

This analysis of pre-Rodinian ophiolites and their environmental consequences has argued for distinctly nonuniformitarian evolution of igneous, tectonic, and environmental processes during Earth history. In addition, Moores et al. (2001) have argued for nonuniformitarian processes in the chemical composition and tectonic significance of magmas formed at past spreading centers. What would have caused the abrupt decrease in thickness of the oceanic crust at 1000 Ma, i.e., a decrease of melting at oceanic spreading centers? At present, the answer is unknown, but three possibilities can be suggested: (1) exhaustion of more fertile mantle and recycling of already-depleted mantle into the source region of oceanic magmas; (2) some process leading to less heat loss from Earth's core (e.g., Larson and Olson, 1991); and (3) stabilization of the "hot abyssal layer" of the lower mantle (Kellogg et al., 1999). The reader might conceive of other possibilities, as well.

This invoking of nonuniformitarian processes may disquiet some readers. However, in their survey of planetary volcanism, Head and Coffin (1997, p. 432–433) stated, "The geologic processes dominating the geologic record over the last several hundred million years on the planets (i.e. the temporal equivalent of the Phanerozoic on the Earth) are not the same as those operating in the earlier history of Mars, Venus, Mercury, and the Moon. Thus, we should anticipate potentially major changes in the style of volcanic extrusion as a function of geologic time on Earth.'' The record of inferred thickness of the oceanic crust and the implied spreading processes from pre-Rodinian ophiolites suggests that Earth also records comparable differences. It is worth reemphasizing the ''continued vital role of geological field work in formulating Precambrian tectonic models'' (Helmstaedt and Scott, 1992, p. 57).

ACKNOWLEDGMENTS

This project was encouraged by the late Chris McA. Powell and was undertaken while on sabbatical leave from University of California, Davis, as a Gledden Senior Research Fellow at the Tectonics Special Research Centre of the University of Western Australia. I am much indebted to the Gledden Foundation for its support, to Rosemarie Powell for her generous and unfailing assistance, and to Catherine Wetherley and Janice Fong for drafting the figures. My understanding of ophiolites, particularly as related to the Precambrian, has been enriched by discussions with many people. I am particularly indebted to field trips to the Jormua complex, led by A. Kontinen, Vuollo Juoni, Eero Hanski et al. in 1998, to the Niquelandia complex in 1982, led by J. Danni, R.A. Fuck, and O. Leonardos, and to the Norseman and Kalgoorlie complexes with R. Fripp and M. Jones in 2000. I have benefited greatly from discussions with M. Barley, I.W.D. Dalziel, M. Dentith, M. de Wit, R. Fripp, C. Hall, S. Occhipinti, M.G. Jones, S.P. Johnson, Z.-X. Li, F. Pirajno, S. Pisarevsky, C. McA. Powell, R. Powell, D. Scott, I. Tyler, R.J. Twiss, and many others. Constructive reviews by M. Brandon, K. Burke, and N. Mortimer significantly improved the manuscript. This is publication 165 of the Tectonics Special Research Centre of the University of Western Australia.

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Manuscript Received by the Society September 22, 2000
Revised Manuscript Received May 16, 2001
Manuscript Accepted August 21, 2001

Printed in the USA