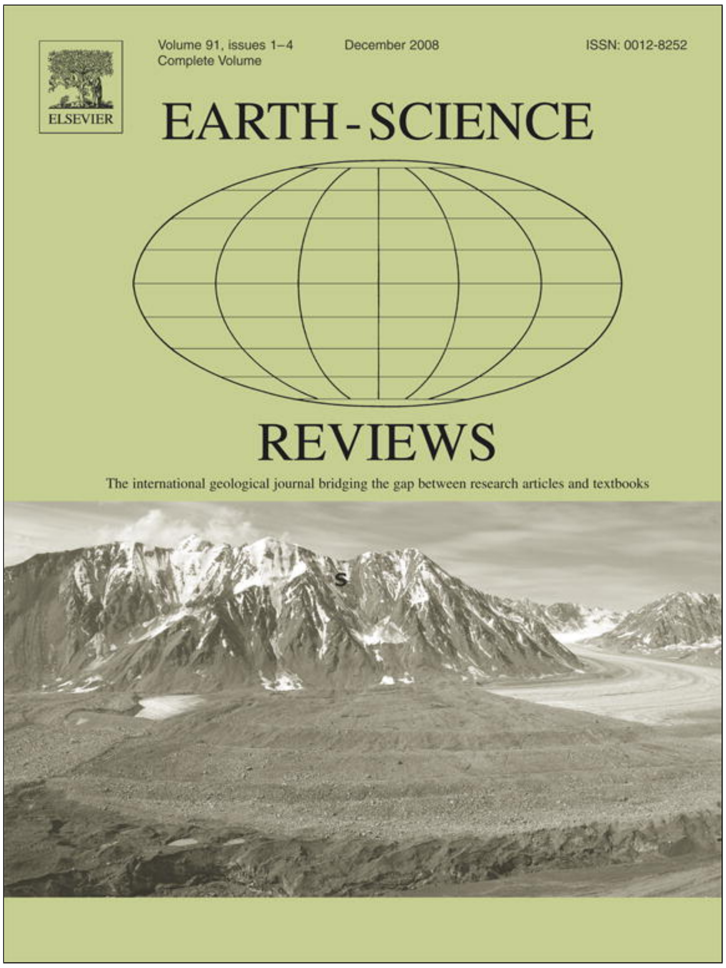
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Passive margins through earth history

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A R T I C L E I N F O

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A B S T R A C T

Passive margins have existed somewhere on Earth almost continually since 2740 Ma. They were abundant at 1900–1890, 610–520, and 150–0 Ma, scarce at ca. 2445–2300, 1600–1000, and 300–275 Ma, and absent before ca. 3000 Ma and at 1740–1600. The fluctuations in abundance of passive margins track the first-order fluctuations of the independently derived seawater $^{87}Sr/^{86}Sr$ secular curve, and the compilation thus appears to be robust. The 76 ancient passive margins for which lifespans could be measured have a mean lifespan of 181 m.y. The world-record holder, with a lifespan of 590 m.y., is the Mesoproterozoic eastern margin of the Siberian craton. Subdivided into natural age groups, mean lifespans are 186 m.y. for the Archean to Paleoproterozoic, 394 m.y. for the Mesoproterozoic, 180 m.y. for the Neoproterozoic, 137 m.y. for the Cambrian to Carboniferous, and 130 m.y. for the Permian to Neogene. The present-day passive margins, which are not yet finished with their lifespans, have a mean age of 104 m.y. and a maximum age of 180 m.y. On average, Precambrian margins thus had longer, not shorter, lifespans than Phanerozoic ones—and this remains the case even discounting all post-300 Ma margins, most of which have time left. Longer lifespans deeper in the past is at odds with the widely held notion that the tempo of plate tectonics was faster in the Precambrian than at present. It is entirely consistent, however, with recent modeling by Korenaga [Korenaga, J., 2004. Archean geodynamics and thermal evolution of Earth. Archean Geodynamics and Environments, AGU Geophysical Monograph Series 164, 7–32], which showed that plate tectonics was more sluggish in the Precambrian. The abundance of passive margins clearly tracks the assembly, tenure, and breakup of Pangea. Earlier parts of the hypothesized supercontinent cycle, however, are only partly consistent with the documented abundance of passive margins. The passive-margin record is not obviously consistent with the proposed breakup of Nuna (Columbia), the assembly of Rodinia, or the assembly or breakup of the putative Pannotia. An alternative model is put forth involving (a) formation of two or more supercratons during the late Paleoproterozoic, (b) a Mesoproterozoic interval dominated by lateral accretion of arcs rather than by continental breakup and dispersal, (c) wholesale collision to form Rodinia by the end of the Mesoproterozoic, and (d) staged breakup of Rodinia through much of the Neoproterozoic.

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1. Introduction

Passive margins are among the most common of the Earth's first-order tectonic features. The present-day passive margins have an aggregate length of 105,000 km, even longer than the spreading ridges (65,000 km) or the convergent plate boundaries (53,000 km). Since the Neoarchean, passive margins have been key players in Wilson Cycles (Burke et al., 1976). Sedimentary successions formed during the rift, drift, and collision stages of passive-margin evolution are major repositories of the stratigraphic record. These strata contain a substantial fraction of the world's hydrocarbon resources (Mann et al., 2003), carbonate-hosted lead–zinc deposits (Leach et al., 2001), and phosphorite deposits (Cook and McElhinny, 1979).

It is surprising, then, that the ancient passive margins have never been systematically studied as a group. When did the first one form? Has their distribution through Earth history been roughly constant or irregular? What is the average duration, or "lifespan", of a passive margin and how has this value changed over time, in response to the decline in Earth's radiogenic heat production? What is the longest-lived passive margin on record? The present study addresses these questions through a survey of the regional geologic literature. In assessing the history of each candidate margin, the goals were: (1) to document the tectonic evolution of the margin, (2) to establish the timing of the rift-drift transition, (3) to establish the timing of the passive margin to foreland-basin transition (except for those rare margins that met another fate), and (4) to identify critical targets for geochronological study. For context, the ages of all the present-day passive margins–those that have yet to complete their Wilson Cycles–were also compiled.

An early synthesis of passive margins was published, in abstract only, by Burke et al. (1984). They estimated the lifespans of 25 mostly Phanerozoic passive margins, and debunked what at the time was a popular idea: that passive margins commonly convert into Andean-type margins by growing old and failing in compression. In a study of supercontinent behavior, Condie (2002) tabulated approximate ages of 38 rifting events and 39 subsequent collisions over the past 1400 Ma. His rifting estimates, however, were based on conflated data from pairs of supposedly conjugate margins (*e.g.*, Laurentia–Kalahari). Accordingly, it is impossible to parse breakup ages (rift-drift transitions) for individual margins from Condie's (2002) data table. The focus of the present study was to document both the abundance of passive margins and their individual lifespans as a function of time. Preliminary accounts of the present study were presented by Bradley in 2005, when the count stood at 50 ancient margins, and in 2007, when the number had reached 63. The present paper covers 85 ancient margins. There remain many more to be mined from the literature, particularly in the Tethyan realm, but the 85 margins are sufficient to reveal first-order trends.

2. Methods

2.1. Definitions

The term *passive margin* is a synonym for the bulkier *Atlantic-type margin*, *trailing-edge margin*, *rifted margin*, or *divergent margin*. A passive margin is one formed by rifting followed by seafloor spreading, so that the resulting plate consists of both continental and oceanic lithosphere, welded across an igneous contact. The distinction between "lower-plate" and "upper-plate" (or volcanic versus non-volcanic) subtypes of passive margins (*e.g.*, Lister et al., 1991) has only been made for a few of the ancient margins in the present compilation. This should be pursued further, as it will help in refining the ages of some rift-drift transitions, as Stampfli et al. (1991) have done for several Tethyan margins. Portions of passive margins that evolved from transform segments of ridge-transform systems (*i.e., sheared passive continental margins* of Scrutton, 1982) are included in the present study. Passive margins formed by backarc extension are a special case. For this study, margins along the continental sides of backarc basins are included (*e.g.*, Chinese margin of the South China Sea), but the volcanic-arc margins of such basins are excluded (The latter would likely be recognized in the rock record as an arc rather than as a passive margin.). Also excluded are any margins that might be more accurately termed *inactive margins*—ones that became "passive" when subduction along a convergent margin stopped. An important distinction in the present compilation is between the modern margins, which face extant ocean basins, and the ancient margins, which occur in orogenic belts.

The entry of a passive margin into a subduction zone is referred to as an *arc–passive margin collision*. In such collisions, it is the forearc and not the magmatic arc that comes into contact with the passive margin; commonly the forearc is a recently formed ophiolite and its syncollisional thrust emplacement is what some workers refer to as *obduction*. For present purposes, it makes little difference whether a collision was of extensional or compressional character, or whether the overriding plate was an oceanic arc or a continental margin—what matters is when collision began. Severely tectonized margins that have been through multiple orogenies are hard to unravel, and this is a particular problem for identification and analysis of Archean and Proterozoic margins. Another challenge is in working out the history of two successive margins that lie one atop the other: the arc that collides to end the first cycle may be carried away at the start of the second cycle.

2.2. Criteria for recognition of ancient passive margins

All of the ancient passive margins have been trapped within orogenic belts and are no longer flanked by ocean floor. Nonetheless, many are still readily recognized by the following criteria: (1) they flank cratons or microcontinents; (2) they are underlain by continental basement, which may or may not have a sedimentary cover predating origination of the margin in question; (3) rift basins overlie the basement; (4) rift deposits and basement are together overlain by an immediately younger, seaward-thickening, seaward-deepening miogeoclinal prism; (5) shallow-water deposits of the miogeoclinal prism are (or more commonly, were) flanked by coeval deep-water facies inferred to have been deposited by stretched continental or oceanic lithosphere, and (6) ophiolites and/or arc sequences were later thrust onto platfom deposits, providing evidence that an ocean basin once existed next to the miogeocline. The deep-water facies and ophiolites also aid in the distinction between a passive margin that was involved in collision, and a deformed (inverted) intracratonic basin that was flanked on all sides by continent and never faced a true ocean.

The hallmark of a passive margin is a miogeoclinal prism—a sedimentary wedge that thickens seaward from a feather edge to 15 km or more. Most easily recognized are miogeoclinal prisms that formed at low latitudes and consequently are dominated by carbonate rocks: limestone, dolostone, and metamorphosed equivalents. In contrast, high-latitude miogeoclinal prisms are typified by compositionally mature shallow-marine sandstone, siltstone, and shale. A carbonate platform succession can still be recognized for what it was despite high-grade metamorphism whereas the depositional setting of metamorphosed siliciclastic rocks may not be so easy to unravel.

2.3. Fates of passive margins

Each of the ancient passive margins in the present synthesis had one of three fates: (1) collision, (2) re-rifting, or (3) direct conversion to a convergent margin (Fig. 1). Most passive margins since the Neoarchean met the first fate: they formed by rifting, endured for a time, and ultimately collided with an arc (Fig. 2). This paper is mainly concerned with this dominant subset of passive margins; the other fates are briefly discussed in Section 10.

2.4. Determining the start date and end date

For modern passive margins, the *start date* (age of initiation) is the age of the oldest magnetic anomaly next to the margin. For ancient passive margins, the equivalent time is approximated by the so-called rift-drift transition—an upward transition from local, fault-controlled rift-type facies (immature clastics, basalt, lacustrine deposits, evaporites) to regionally extensive platformal carbonates (at lower latitudes) or siliciclastics (at high latitudes) (Fig. 2a). In some cases, a breakup unconformity is taken as approximating the rift-drift transition. Tectonic subsidence curves typically show a sharp bend representing a transition from rapid, fault-controlled subsidence to slower, exponential, thermally controlled subsidence; the inflection point approximates the rift-drift transition. Dating the rift-drift transition is more difficult than dating the passive-margin to foreland-basin transition because (1) mafic rocks that typify rift settings are more difficult to date than felsic tuffs and syntectonic granitoids in collisional settings; (2) the onset of seafloor spreading corresponds to a shift in the locus of tectonism to a place where preservation is highly unlikely, (3) rifting before seafloor spreading often lasts many

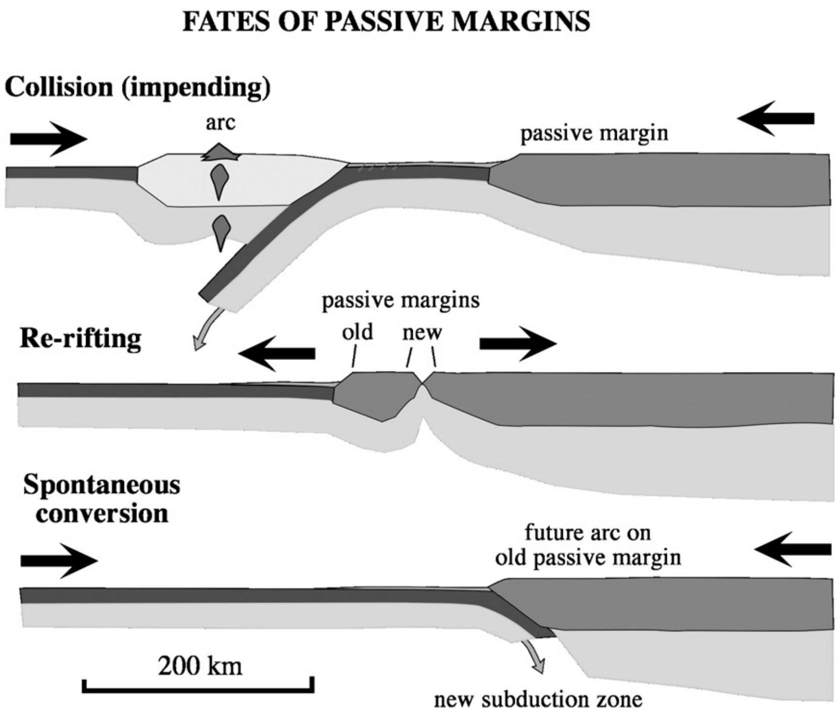


Fig. 1. The three common fates of passive margins. (a) Collision between a passive margin and an arc. For the aims of this study, the nature of this arc–whether intraoceanic or continental, and whether extensional or compressional–is not important. (b) Re-rifting of a preexisting passive margin by separation of a ribbon microcontinent. (c) Spontaneous conversion of a passive margin to a convergent margin by initiation of a subduction zone at or near the ocean–continent boundary.

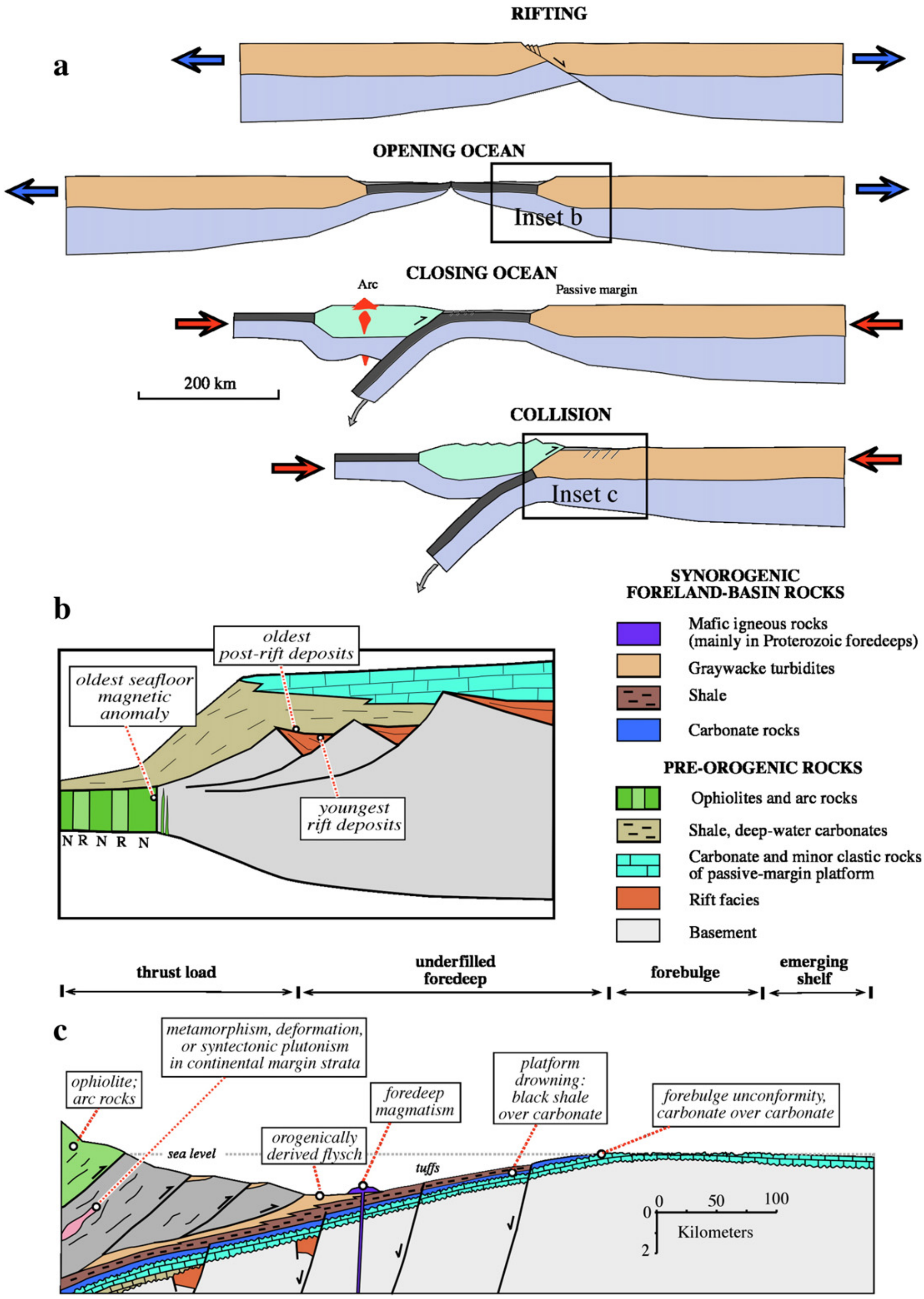


Fig. 2. (a) Model of passive-margin evolution showing stages of rifting, seafloor spreading, ocean closure, and collision. (b) Close-up showing criteria for picking the age of the rift-drift transition (“start date” in text). (c) Close-up showing criteria for picking the age of the passive-margin to foreland-basin transition (“end date” in text).

tens or even hundreds of millions of years, so dated rift deposits may not provide very precise age bracketing, and (4) most passive-margin successions lack igneous rocks entirely and thus are not *directly* datable by the most common targets of conventional isotope geochronology. Less direct age constraints on the age of the rift-drift transition are also provided by oceanic rocks in some orogens. When an arc eventually collides with a passive margin, the oldest arc rocks are commonly interpreted to date the onset of subduction. Barring complications, the rift-drift transition at the passive margin should be older still.

For passive margins that ended with collision, the *end date* is defined here as the time when the continent–ocean boundary is subducted, as is happening today just south of Taiwan (Malavieille et al., 2002). The end date can be gleaned from sedimentary rocks in the undeformed foreland, and from sedimentary, igneous, and metamorphic rocks within the orogen (Fig. 2c). The stratigraphic signature

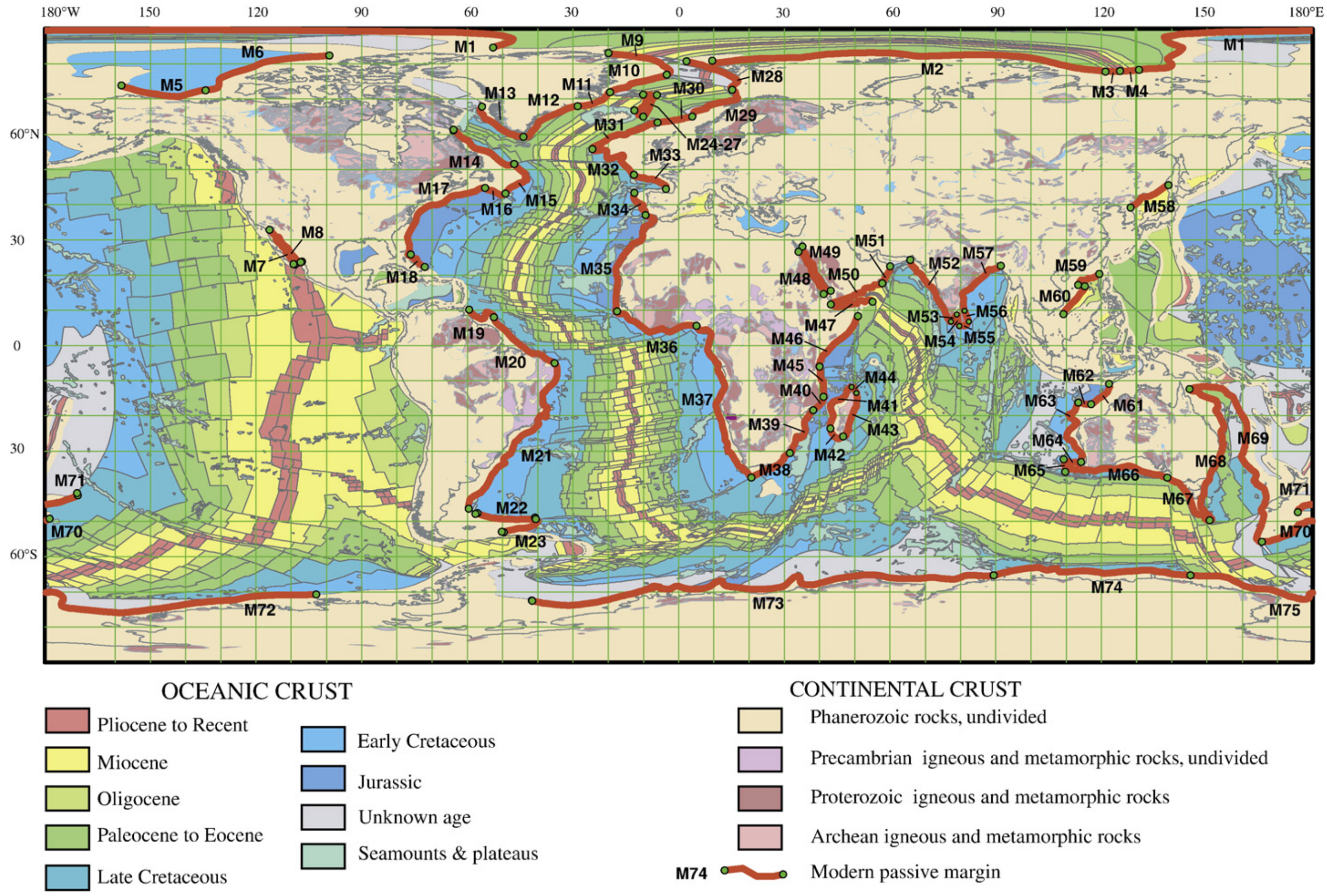


Fig. 3. World map showing modern passive margins. Base map from Commission de la Carte Géologique du Monde (2000). Green circles divide margins into age sectors.

of an arc–passive margin collisional foreland is remarkably systematic (*e.g.*, Rowley and Kidd, 1981; Bradley, 1989; Sinclair, 1997) and is described in Section 4.2 on the Appalachian margin. Briefly, the transition from passive margin to foreland basin is recorded by a platform drowning sequence and influx of outboard-derived siliciclastics; many dating opportunities are afforded by fossils and (or) ashfall tuffs. Syncollisional extension of the foreland–a consequence of flexural loading of the passive-margin plate (Bradley and Kidd, 1991)–is common, and in some orogens has been mistaken for rifting at the *start* of the Wilson Cycle. In some cases, the best age control comes from within the orogenic belt itself, as revealed by the age of initial metamorphism of continental-margin rocks or by the age of the oldest syntectonic plutons in the orogenic wedge.

2.5. Census methods, disclaimers, and error estimates

The present compilation includes summaries of 85 passive margins, based almost exclusively on the recent English-language literature. A number of margins were unearthed using the Georef bibliographic database, by means of search criteria such as "Amazonia" plus "passive margin", "Mesoproterozoic" plus "passive margin", and "India" plus "passive margin". Schematic plate reconstructions (*e.g.*, Sengör et al., 1988; Stampfli et al., 1991; Zhao et al., 2002, 2004; Scotese, 2004) provided leads to other margins.

The most severe errors are likely to come from tectonic misinterpretations. Despite the fact that arc–passive margin collisions are the most tractable type of orogeny, every such collision has its controversies and interpretive problems. Even in the thoroughly studied Wopmay and Appalachian orogens, rift and foreland-basin sequences were once mistaken for one another. In working out a margin's lifespan, this kind of mistake would trump all other potential errors. Even when the tectonostratigraphic framework has been correctly interpreted, age control may be poor. Ideally, the rift-drift and passive margin-foreland-basin transitions are tightly bracketed by dated rocks from immediately below and immediately above the crucial boundaries; but in reality, the rock record is not so forthcoming. Uncertainties simply cannot be quantified where one of the key age constraints is a guess. The only kinds of uncertainties that can be quantified are encountered at a more detailed level, such as analytical errors of geochronological ages, biostratigraphic uncertainty of fossil collections, and uncertainties in the calibration of the geologic time scale.[1]

Given the many potential problems, it is fruitless to guess at the numerical error associated with the start date, end date, and lifespan for most margins. Instead, each margin is assigned an overall quality rating on a scale from A to D (best to worst). A margin with a rating of "A" has a robust tectonic interpretation, and both the start date and end date are reasonably well constrained; the estimated lifespan is probably correct within 10 to 20 million years. A quality rating of "B" is given when one key ingredient–start date, end date, or tectonic interpretation–is weak. A passive margin given a rating of "C" is only minimally acceptable for this study: both the tectonic interpretation and the lifespan estimate are debatable. For margins with quality

[1] The recent Gradstein and Ogg (2004) time scale was used here to assign numerical ages to stratigraphically or paleontologically defined events in the Phanerozoic. Where previous workers used an alternative older time scale for this purpose, the numbers have been revised accordingly.

Table 1
Modern passive margins. For locations, see Fig. 3. See the Appendix for notes that explain some of the more difficult age picks. Adapted from a preliminary compilation by David Rowley

Number	Margin name	Ocean	Oldest age (Ma)	Youngest age (Ma)	Mean age (Ma)	Length (km)
M1	Lomonosov Ridge, facing Barents shelf (R edge map)	Arctic	60	58.4	59.2	1393
M1	Lomonosov Ridge, facing Barents shelf (L edge map)	Arctic (Eurasian Basin)	60	58.4	59.2	633
M2	Europe, N margin (Barents shelf)	Arctic	58.4	15.4	36.9	2328
M3	Eurasia, N coast (Lena delta W of Gakkel ridge)	Arctic	52	0	26.0	109
M4	Eurasia, N coast (Lena delta E of Gakkel ridge)	Arctic	52	0	26.0	129
M5	Alaska, N coast	Arctic (Canada Basin)	130.2	130.2	130.2	789
M6	Canada, N coast	Arctic (Canada Basin)	130.2	130.2	130.2	1449
M7	Baja, E coast	Gulf of California	17.5	1.7	9.6	993
M8	Mexico, W coast	Gulf of California	17.5	1.7	9.6	1097
M9	Greenland, E coast, NE corner	Arctic	59.8	13.5	36.65	758
M10	Greenland, E coast (north)	North Atlantic	56.8	55.9	56.4	737
M11	Greenland, E coast (central)	North Atlantic	38.6	30.6	34.6	565
M12	Greenland, E coast (south)	North Atlantic	63.8	58.8	61.3	1244
M13	Greenland, W coast	Labrador Sea	89.5	87.5	88.5	1126
M14	North America, Labrador coast	Labrador Sea	109	68	88.5	1540
M15	N America, E coast (Grand Banks)	Central Atlantic	129	127	128.0	951
M16	N America, E coast (S side Grand Banks)	Central Atlantic	150.9	134.7	149.5	493
M17	N America, E coast (United States)	Central Atlantic	171	170	170.5	2889
M18	Bahama plateau, N margin	Central Atlantic	170	146	158.0	637
M19	S America, N coast (Guyana)	South Atlantic	103.8	94	98.9	813
M20	S America, N coast (Amazon)	South Atlantic	109.7	102.1	105.9	2434
M21	S America, E coast (Argentina)	South Atlantic	134.8	109.6	122.2	5198
M22	Falkland plateau, N margin	South Atlantic	134.3	116.7	125.5	1295
M23	Falkland plateau, SE margin	South Atlantic	139.8	138.9	139.4	725
M24	Jan Mayen plateau, N margin	North Atlantic	57.3	34.7	46.0	156
M25	Jan Mayen plateau, E margin	North Atlantic	59.3	52.4	55.9	641
M26	Jan Mayen plateau, S margin	North Atlantic	52.4	35.4	43.9	133
M27	Jan Mayen plateau, W margin	North Atlantic	35.4	35.4	35.4	582
M28	Svalbard, S margin	North Atlantic	46.6	14.6	30.6	790
M29	Europe, W coast (Norwegian Sea)	North Atlantic	58.6	46.7	52.7	976
M30	Europe, margin W of Shetlands facing Jan Mayen	North Atlantic	59.3	52.4	55.9	528
M31	Europe, margin W of Ireland facing Iceland	North Atlantic	59.9	56.5	58.2	1336
M32	Europe, margin SW of Ireland	Central Atlantic	108.7	82.6	95.6	1236
M33	Europe, W coast, France	Bay of Biscay	108.7	102.2	105.4	806
M34	Iberia, W coast	Central Atlantic	129.7	119.5	124.6	792
M35	Africa, W coast (Mauritania)	Central Atlantic	171.4	167.2	169.3	3169
M36	Africa, W coast (Guinea)	Central Atlantic	109.9	109.5	109.7	2545
M37	Africa, W coast (Namibia)	South Atlantic	133.6	119.9	126.8	5067
M38	Africa, S coast (Agulhas)	Southern	143.6	126.6	135.1	1316
M39	Africa, E coast (Mozambique)	Indian	179.6	179.3	179.5	1472
M40	Africa, E coast (Mozambique E of M39)	Indian	174.5	158.5	166.5	546
M41	Madagascar, W coast	Mozambique Channel	169.9	169.5	169.7	1431
M42	Madagascar, SW end	Mozambique Channel			132.9	490
M43	Madagascar, E coast	Indian	103.5	99.1	101.3	1448
M44	Madagascar, NE corner	Indian			134.3	228
M45	Africa, E coast (Dar Es Salaam)	Indian	169.7	156.3	163.0	966
M46	Africa, E coast (Somalia)	Indian	169.8	168.9	169.3	1992
M47	Africa, N coast (Somalia)	Gulf of Aden	29.8	28.5	29.2	1289
M48	Egypt, E coast	Red Sea	5	5	5.0	1591
M49	Arabia, W coast	Red Sea	5	5	5.0	1615
M50	Arabia, S coast (Yemen)	Gulf of Aden	29.4	28.5	29.0	1716
M51	Arabia, E end (33–75 Ma)	Arabian Sea	33	75	54.0	585
M52	India, W coast	Arabian Sea	119.4	98.6	109.0	2295
M53	India, SE tip facing Sri Lanka	Indian, Gulf of Mannar	128.1	122.2	125.2	298
M54	Sri Lanka, W coast facing SE tip of India	Indian, Gulf of Mannar	128.1	120.8	124.5	331
M55	Sri Lanka, S coast	Indian	120.2	119.6	119.9	323
M56	Sri Lanka, NE coast	Indian, Bay of Bengal	127.8	119.8	123.8	341
M57	India, E coast N of Sri Lanka	Indian, Bay of Bengal	128.1	120.4	124.3	1812
M58	N Korea, E margin	Japan Sea	28	28	28	620
M59	China, SE margin, Hainan to Taiwan	South China Sea	31	31	31	390
M60	Macclesfield Bank E of Vietnam, SE side	South China Sea	27	19	23	550
M61	Australia, NW coast (Broome)	Indian	150.6	140.1	145.4	844
M62	Australia, NW coast (N margin Exmouth Plat.)	Indian	150.9	134.7	142.8	403
M63	Australia, W coast N of Perth	Indian	134.2	128.3	131.3	1897
M64	Naturaliste Plateau, W of Perth, N margin	Indian	126	132	129.0	397
M65	Naturaliste Plateau, W of Perth, W margin	Indian	130	130	130.0	199
M66	Australia, S coast, western part	Southern	89.2	82.7	86.0	2576
M67	Tasmania, W coast	Southern	64.9	33.8	49.4	1658
M68	Australia, E coast	Coral–Tasman	89.3	72.9	81.1	4128
M69	Lord Howe Rise, W margin	Coral–Tasman	84.5	72.9	78.7	5123
M70	Campbell Plateau, S margin (R edge of map)	Southern	87.4	87	87.2	1183
M70	Campbell Plateau, S margin (L edge of map)	Southern	87.4	87	87.2	139
M71	Chatham Rise, S margin (R edge map)	Southern	80	80	80.0	377

Table 1 (*continued*)

Number	Margin name	Ocean	Oldest age (Ma)	Youngest age (Ma)	Mean age (Ma)	Length (km)
M71	Chatham Rise, S margin (L edge map)	Southern	80	80	80.0	755
M72	Antarctica facing Pacific	Southern	87.4	80	85.0	2898
M73	Antarctica, S of Africa	Southern	179.6	120.4	134.6	5381
M74	Antarctica, S of Australia	Southern	89.2	82.7	86.0	2591
M75	Antarctica, S of Tasmania	Southern	64.9	33.8	49.4	1564

ratings of "B", the quoted lifespans have been rounded to the nearest 5 m.y., for margins with a rating of "C", lifespans are rounded to the nearest 10 m.y. A rating of "D" is given to supposed passive margins for which a lifespan cannot even be estimated. Some margins in this category are quite controversial but are included nonetheless, with caveats. In general, local experts would likely bestow higher quality ratings than I have—a lifetime of study of a given margin will obviously lead to a deeper appreciation of the age constraints than I have been able to glean during this broad sysnthesis. The write-up for each margin explains my reasons for choosing the start dates and end dates.

Not all of the eighty-five margins of the complete dataset are amenable to the same treatment. For most of the margins, a numerical age can be set for both the start date and end date, and these are assigned to Group 1 in Table 2 (n=76). Group 2 includes a few margins for which the end date is known, but not the start date (n=5). Group 3 the remaining margins for which neither a start date nor end date is known (n=4).

3. Modern passive margins

The locations and ages of the present-day passive margins are evident from world maps of bathymetry, seismicity, and magnetic anomalies. Fig. 3 shows the distribution of modern margins and Table 1 summarizes their lengths and ages. The cited age of each margin is the age of the oldest flanking magnetic anomaly. Margins that formed diachronously by rift propagation were subdivided into age sectors and a mean age is cited for each sector. Lengths of modern margins were taken as the great-circle distance between endpoints.

The present Earth has about 105,000 km of passive margins (Fig. 3). They range in age from ca. 5 m.y. (Red Sea) to ca. 180 m.y. (Mozambique sector of Africa's east coast) (Fig. 4 and Table 1). The mean age of all the modern passive margins, weighted by length, is about 104 m.y.

4. Ancient passive margins

Table 2 lists the start date, end date, lifespan, and quality ranking of the ancient passive margins. These data are plotted in Fig. 5 and the margins are located in Fig. 6. Four margins are discussed at length in the main body of the paper: a modern-day collision (northern Australia), a classic Phanerozoic example (the Cambrian–Ordovician Appalachian margin of Laurentia), the longest-lived example (Mesoproterozoic eastern margin of the Siberian craton), and the oldest example that has sound age constraints (Kaapvaal craton). The rest are described in variable detail in the Appendix.

4.1. Present-day collision between Australia and the Banda forearc

Northern Australia was entirely rimmed by passive margins from the Cretaceous to the Oligocene when the eastern sector collided with an arc in New Guinea (Appendix, margin A80). The western part of the northern margin (margin A79) survived as a passive margin until a few million years ago when it began colliding with the Banda forearc (Fig. 7) (*e.g.*, Karig et al., 1987). The trend of the yet-to-collide part of the passive margin is oblique to that of the trench, and the collision tip is thus propagating westward. Pre-collisional conditions along the passive margin, forearc, and arc can be seen west of the collision tip (Fig. 7).

Australia's northern and western margins experienced a protracted history involving departure of a series of blocks during the

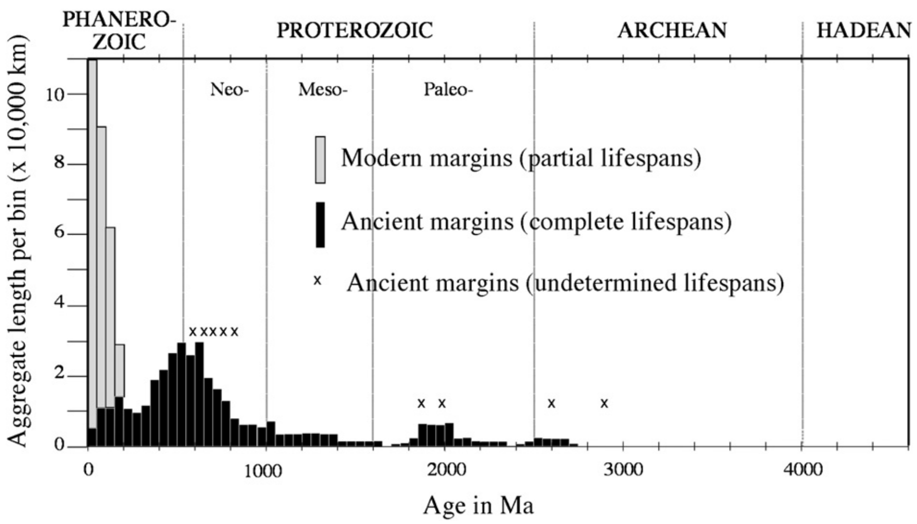


Fig. 4. Histograms showing the age distribution of ancient and modern passive margins using data in Tables 1 and 2. Bins are 50 m.y. in duration and each margin is weighted by length. Thus, a particular 1000-km-long passive margin that existed from 525 to 375 Ma would contribute 1000 km to the height of four consecutive bins.

Table 2
Summary data for ancient passive margins as described in text and Appendix. "Group" refers to the subset of data used for plots and calculations as discussed in Section 2.5. Asterisk indicates a rounded value. Locations are given in Fig. 6

Number	Group	Margin and orogen	Where	Start date (Ma)	End date (Ma)	Lifespan (m.y.)	Quality	High-P metamorphism?	Foredeep magmatism?	Length (km)
A1	1	Arctic Alaska microcontinent, S side, Brookian orogen	Alaska, Russia	350	170	180	A	Y	N	1230
A2	1	Farewell terrane	Alaska	545	435	110	C	N	N	460
A3	1	Laurentian craton, W side, northern sector, Antler orogen	Canada	710	385	325	B	N	Y	1560
A4	1	Laurentian craton, W side, southern sector, Antler orogen	USA, Canada	542	357	185	A	N	Y	1870
A5	1	Laurentian craton, N side, Innuitian margin, Ellesmerian orogen	Canada, Greenland	620	444	180*	C	N	N	1900
A6	1	Slave craton, W side, Wopmay orogen	Canada	2015	1883	132	A	N	Y	560
A7	1	Slave craton, E side, Kimerot platform, Thelon orogen	Canada	2090	1970	120	C	N	N	630
A8	1	Laurentian craton, N side, Borden Basin, Poseidon orogen	Canada	1255	1200	55	C	N	N	270
A9	1	Hearn craton, SE side	Canada	2070	1880	190	B	N	Y	820
A10	3	Steep Rock Lake platform, Superior Craton, Wabigoon Province	Canada	Unk (<3002)	Unk (<2870)	Unk	D	N	N	15
A11	1	Wyoming Craton, S side, Medicine Bow orogen	USA	2000	1780	220	B	N	N	310
A12	1	Superior craton, S side, Huronian margin, Penokean orogen	Canada, USA	2300	2065	235	C	N	NA	1170
A13	1	Superior craton, S side, Animike margin, Penokean orogen	Canada, USA	2065	1880	185	B	N	Y	1170
A14	1	Superior Craton, N side, Cape Smith and Trans-Hudson orogens	Canada	2000	1875	125	B	N	Y	370
A15	1	Superior craton, E side, New Quebec orogen e ("Labrador Trough")	Canada	2135	1890	245	B	N	Y	960
A16	2	Nain craton, W margin, Torngat orogen	Canada	Unk (>1876)	1859	Unk	D	N	N	290
A17	1	Nain Craton, N side, Makkovik orogen	Canada	2175	2010	165	C	N	N	110
A18	1	Laurentian craton, S side, Ouachita orogen	USA	520	310	210	A	N	N	1720
A19 (a, b)	1	Laurentian craton, E side, Appalachian margin, Taconic orogen (a) and NW Scotland (b)	USA, Canada	540	465	75	A	Y	Y	3320
A20 (a,b)	1	Laurentian craton, E side, E Greenland (a) and NE Svalbard (a)	Greenland, Svalbard	815	444	370*	C	Y	N	1740
A21	1	Baltic Craton, W side, Scandinavian Caledonide orogen	Norway	605	505	100	A	Y	N	1550
A22	1	Kola craton, S side, Kola suture belt	Russia	1970	1800	170	B	N	N	690
A23	1	Baltic craton, N side, Timanides	Russia	1000	560	440	C	N	N	1460
A24	1	Baltic craton, E side, Uralian orogen, Phase 1	Russia	1000	620	380	B	Y	N	1000
A25	1	Baltic Craton, E side, Uralian orogen, Phase 2	Russia	477	376	101	A	N	N	3130
A26	1	Baltic craton, S side, Variscan orogen	Ireland to Poland	407	347	60	A	Y	Y	1080
A27	1	Saxo-Thuringian block	Germany	444	344	100	B	N	N	400
A28	1	S margin of Europe, Alpine orogen	Switzerland	170	43	127	A	Y	N	790
A29	1	Pyreneean-Biscay margin of Iberia	Spain	115	70	45	A	N	N	390
A30	1	NW Iberia, Variscan orogen	Spain	475	385	90	B	N	Y	210
A31	1	Apulian microcontinent, Pindos ocean	Greece	230	60	170	A	Y	N	630
A32	1	Isparta angle, western margin	Turkey	227	60	167	A	Y	N	280
A33	1	Isparta angle, eastern margin	Turkey	227	53	174	A	Y	N	260
A34	1	Arabia, NE margin, Oman-Zagros orogen	Oman, Iran	272	87	185	A	Y	N	2300
A35	1	Alborz orogen	Iran	390	210	170	B	N	N	550
A36	2	Siberian craton, N side, Taymyr, Phase 1	Russia	>750	600	>150	D	N	N	1380
A37	1	Siberian craton, N side, Taymyr, Phase 2	Russia	530	325	205	B	N	N	1380
A38	1	Siberian craton, W side, Yenisei Ridge	Russia	1350	850	500	C	N	N	1910
A39	3	Gargan microcontinent, W side	Russia, Mongolia	Neoprot.	Neoprot	Unk	D	N	N	150
A40	1	Siberian craton, S side, Baikal	Russia	1000	600	400	C	N	N	1160
A41	1	Dzabkhan block	Mongolia	710	580	130	B	N	N	200
A42	3	Idermeg terrane	Mongolia	Neoprot.	Mid-Camb.	Unk	D	N	N	300
A43	1	Karakorum block	Pakistan	268	193	75	B	N	N	500
A44	1	Tarim microcontinent, N side, Tien Shan orogen	China	600	380	220	C	N	N	1320
A45	1	Tarim microcontinent, S side, Kunlun orogen	China	600	430	170	B	N	N	950
A46	1	Indian craton, N side, Himalayan orogen, Phase 1	India, Nepal	635	502	133	A	N	N	2460
A47	1	Indian craton, N side, Himalayan orogen, Phase 2	India, Nepal	271	52	219	A	Y	N	2460
A48	1	Aravelli-Delhi orogen	India	2000	1800	200	C	N	N	740
A49	1	South China Craton, NW side, Longmen Shan orogen, Phase 1 (ended by re-rifting)	China	600	300	300	C	N	NA	480
A50	1	South China Craton, NW side, Longmen Shan orogen, Phase 2	China	300	228	72	B	N	N	480
A51	1	North China craton, Central orogenic belt	China	2740	2530	210	C	N	N	900
A52	1	South China Craton, N side, Qinling–Dabie orogen, Phase 1	China	750	440	310	C	N	N	560
A53	1	South China Craton, N side, Qinling–Dabie orogen (2)	China	360	235	125	C	Y	N	560
A54	1	South China craton, SE side, Nanling orogen	China	635	445	190	B	N	N	1730
A55	1	China, E side, Taiwan orogen	Taiwan	28	5	23	A	Y	Y	500

Table 2 (*continued*)

Number	Group	Margin and orogen	Where	Start date (Ma)	End date (Ma)	Lifespan (m.y.)	Quality	High-P metamorphism?	Foredeep magmatism?	Length (km)
A56	1	Siberian craton, E side, Verkhoyansk orogen, Phase 1	Russia	1600	1010	590	C	N	Y	1700
A57	1	Siberia, E side, Verkhoyansk, Phase 2 (ended by re-rifting)	Russia	650	380	270	C	N	NA	1700
A58	1	Siberia, E side, Verkhoyansk, Phase 3	Russia	380	160	220	C	N	N	1700
A59	1	Guaniguanico terrane	Cuba	159	80	79	A	N	N	150
A60	1	S. American Craton, N side, Venezuela margin	Venezuela	159	34	125	A	Y	N	1080
A61	1	Amazon craton, SE side, Araras margin, Paraguay orogen	Brazil	640	580	60	B	N	N	470
A62	1	Cuyania terrane, Argentine Precordillera, E side	Argentina	530	473	57	A	N	N	640
A63	1	Sao Francisco craton, W side, Brasiliano orogen	Brazil	745	640	105	B	N	N	1080
A64	1	Sao Francisco craton, E side, Aracauai–Ribeira orogen	Brazil	900	590	310	B	N	N	750
A65	1	Sao Francisco craton, S side, Transamazonian orogen	Brazil	2500	2130	370	C	N	N	200
A66 (a,b,c)	1	Sierra de la Ventana (a), Cape (b), and Ellsworth Mtns. (c)	Argentina, S. Africa, Antacrtica	500	300	200	B	N	N	2170
A67	1	West African craton, N side, Anti-Atlas orogen	Morocco	800	580	220	C	N	N	560
A68	1	West African craton, W side, Mauritanide orogen	Mauritania	675	650	25	C	N	N	1692
A69	1	West African craton, E side, Dahomeyide orogen	Ghana, Mali	775	625	150	C	Y	N	2200
A70	2	LATEA craton, Hoggar, W side	Algeria, Mali	Unk	870	Unk	D	NA	NA	840
A71	2	LATEA craton, Hoggar, S side	Algeria, Niger	Unk	685	Unk	D	NA	NA	550
A72	1	East Africa Orogen, W side	Sudan	840	730	110	C	N	N	1460
A73	1	Congo Craton, W side, Kaoko Belt (N Coastal Branch) of Damara orogen	Namibia	780	580	200	A	N	N	2175
A74	1	Congo Craton, S side, Inland Branch, Damara orogen	Namibia	670	555	115	A	N	N	1800
A75	1	Kalahari Craton, N side, Inland Branch, Damara orogen	Namibia	670	550	120	A	N	N	540
A76	1	Kalahari Craton, W side, Gariep Belt, (S Coastal Branch) of Damara orogen	Namibia	735	535	200	A	N	N	690
A77	1	Kaapvaal Craton, W side	S. Africa, Botswana	2640	2470	170	C	N	N	810
A78	3	Zimbabwe craton, S side, Belingwe greenstone belt	Zimbabwe	ca. 2600	ca. 2600	Unk	D	N	N	60
A79	1	Australian craton, NW side, Timor orogen	Indonesia	151	4	147	A	Y	N	1530
A80	1	Australian craton, NE side, New Guinea orogen	New Guinea, Irian Jaya	180	26	154	A	N	Y	1380
A81	1	Northwest Australia craton, Halls Creek orogen	Australia	1880	1860	20	C	N	N	270
A82	2	Rudall Complex	Australia	>2000	1780	>220	D	N	N	330
A83	1	Pilbara Craton, S margin, Ophthalmian orogen	Australia	2685	2445	240	B	N	Y	580
A84	1	Gawler Craton, NE side, Kimban orogen	Australia	1900	1740	160	C	N	N	650
A85	1	Australia, E side, Tasman orogen	Australia	590	520	70	B	N	N	1670

Paleozoic and Mesozoic (Longley et al., 2002). Just southwest of the collision tip (Fig. 7), the age of the rift-drift transition is evident from the oldest seafloor adjacent to the continent (ca. 155 Ma; Longley et al., 2002). Longley et al. (2002, their Figs. 10 and 15) interpreted this sector of the passive margin to be related to the rifting away of the Argo microplate. East of the collision tip (Fig. 7), the already collided portion of the passive margin is interpreted to have a slightly younger rift-drift transition, related to separation of a part of the West Burma block at about the Kimmeridgian–Tithonian boundary (ca. 151 Ma) in the Late Jurassic (Longley et al., 2002, their Figs. 7 and 10). At the collision tip, the distance from Australia's shoreline to the ocean–continent boundary is about 500 km (Longley et al., 2002); this gives an indication of the distances that might be involved in reconstructing stratigraphy across an ancient passive margin and foredeep.

The present-day Java Trench can be tracked directly into Timor Trough, establishing the continuity between the pre-collisional subduction trench and the syncollisional foredeep. The former, floored by Jurassic seafloor, is 6–7 km deep, whereas the latter, floored by Australian continental crust, is only 2–3 km deep. At DSDP Site 262 in the axis of Timor Trough, modern trench sediments depositionally overlie Pliocene shallow-marine carbonates (Veevers et al., 1978), demonstrating that the Australian margin has subsided in only a few million years to abyssal depths. The southern margin of Timor Trough is cut by young normal faults. These have been interpreted as the product of collision-induced flexure of the downgoing plate and not the product of classic rifting related to plate divergence (Bradley and Kidd, 1991). A discontinuous forebulge is evident on the downgoing passive-margin plate, most notably at Kepulauan Aru where Neogene shelf deposits have been broadly upwarped along an arch that parallels the trench (Bradley and Leach 2003) (Fig. 7).

The northern wall of Timor Trough is an actively deforming submarine orogen (Karig et al., 1987). Although much of the collisional orogen lies below sea level, it is emergent on the island of Timor, where an ocean-floor sedimentary sequence that ranges from Neocomian to Miocene was thrust onto the Australian margin in mid-Pliocene time (ca. 4 Ma) (Carter et al., 1976, p. 184). An important feature of the present-day collision is that the magmatic axis of the Banda Arc is far from the collision zone, separated from the orogen in Timor by a deep forearc basin, the Savu Basin. Substantial orogenic shortening will need to happen if the magmatic arc itself is ever to come close to the passive margin.

Age controls for this relatively young margin are tighter than they are for most of the other ancient margins. The rift-drift transition was at ca. 151 Ma and the onset of collision was at about 4 Ma, yielding a lifespan of about 147 m.y. This is probably accurate to within about 10 m.y. and applies to a 1500-km-long section of passive margin.

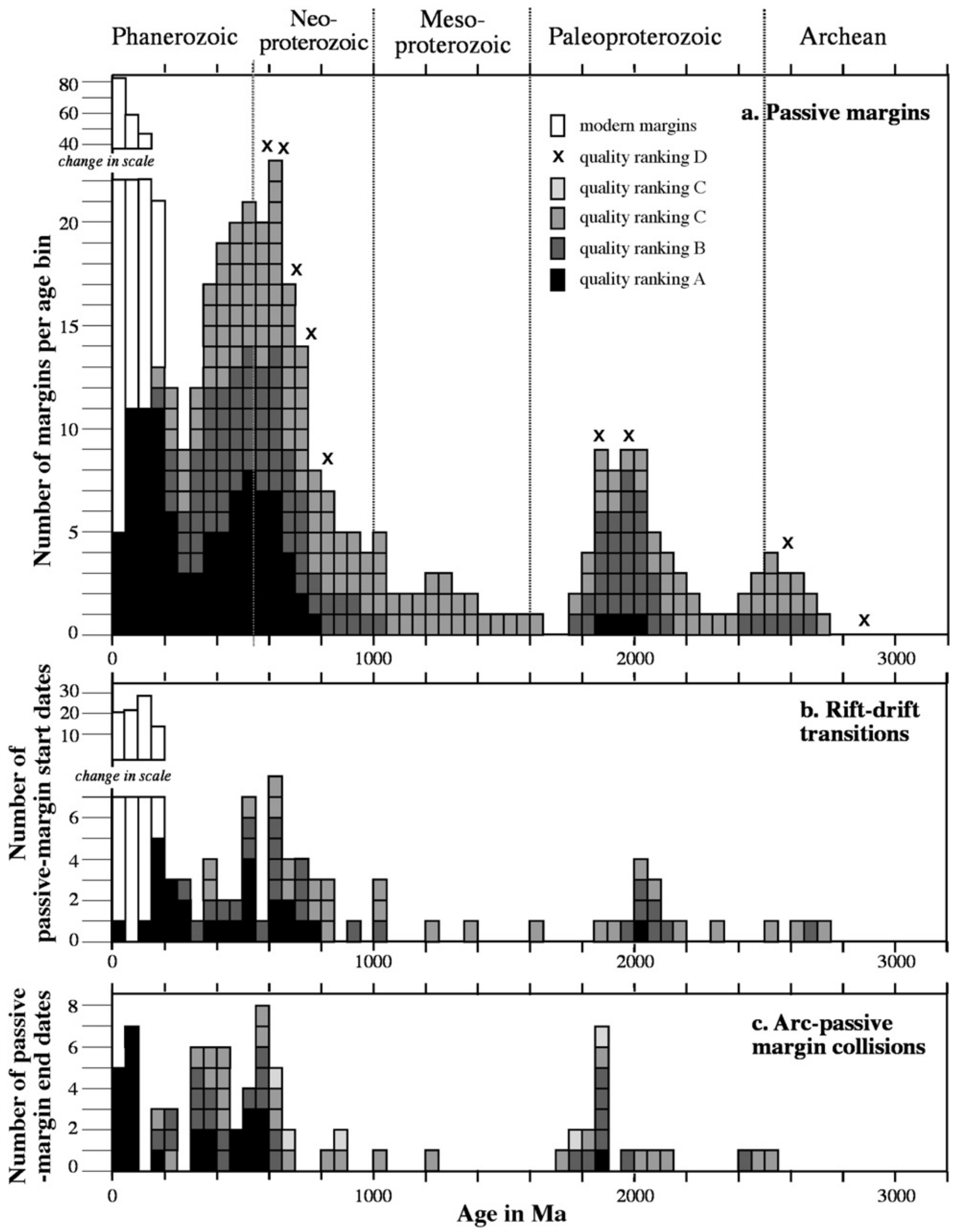


Fig. 5. Histograms showing age distributions of the ancient margins, from data in Table 2. Quality rankings are color coded. (a) Number of margins per 50-m.y. time bin. Each margin appears in anywhere from one to twelve consecutive histogram bins, depending on its lifespan. For example, a margin with a start date of 550 Ma and an end date of 475 Ma will show up in the 550–600, 500–550, and 450–500 Ma bins. Based on Group 1 data; $n=76$. Also plotted (x symbol) are the 9 margins having quality rankings of D. Except for the oldest one (Steep Rock Lake, margin A10), these amplify rather than modify the age distribution shown by the better-dated margins. (b) Times of rift-drift transition, based on Group 1 data (76 ancient margins) and 75 modern margins. (c) Times of arc–passive margin collision, based on Group 1 and Group 2 data; $n=81$.

4.2. Cambrian–Ordovician Appalachian margin of Laurentia

The Appalachian margin of eastern Laurentia (margin A19a) is one of the most thoroughly studied and well understood Phanerozoic passive margins. No margins in the unfossiliferous Proterozoic boast such extraordinary age control. The Appalachian margin extends a distance of about 3300 km from easternmost Canada to the southeastern United States. In the southeastern United States, Laurentia's Appalachian margin swings west and becomes the Ouachita segment; this segment has a different history from the Appalachians and is discussed separately (margin A18). On a fit of the Atlantic continents, a continuation of the Appalachian margin is recognized in northwestern Scotland (margin A19b); this segment is not discussed further but has a history similar to that of Newfoundland (van Staal et al., 1998; Cawood et al., 2007A). The portion of the margin within the United States was summarized in comprehensive reviews by Rankin et al. (1989), Read (1989), and Drake et al. (1989). The Canadian portion was likewise covered by Williams et al. (1995), Knight et al. (1995), Hiscott (1995), and Williams (1995). The following discussion focuses on two transects: (1) Newfoundland (Fig. 8b), where all the main phases of passive-margin evolution–rifting, thermal subsidence, and collision–are clearly legible in the rock record, and (2) northern New York (Fig. 8c), where the foredeep is especially well exposed on the flanks of a Neogene dome, the Adirondack Mountains.

Granulite-facies metamorphic rocks and plutonic rocks form the basement to the passive margin. These rocks belong to the 1.3 to 1.0-Ga Grenville Province (Rivers, 1997) and occur both as autochthonous

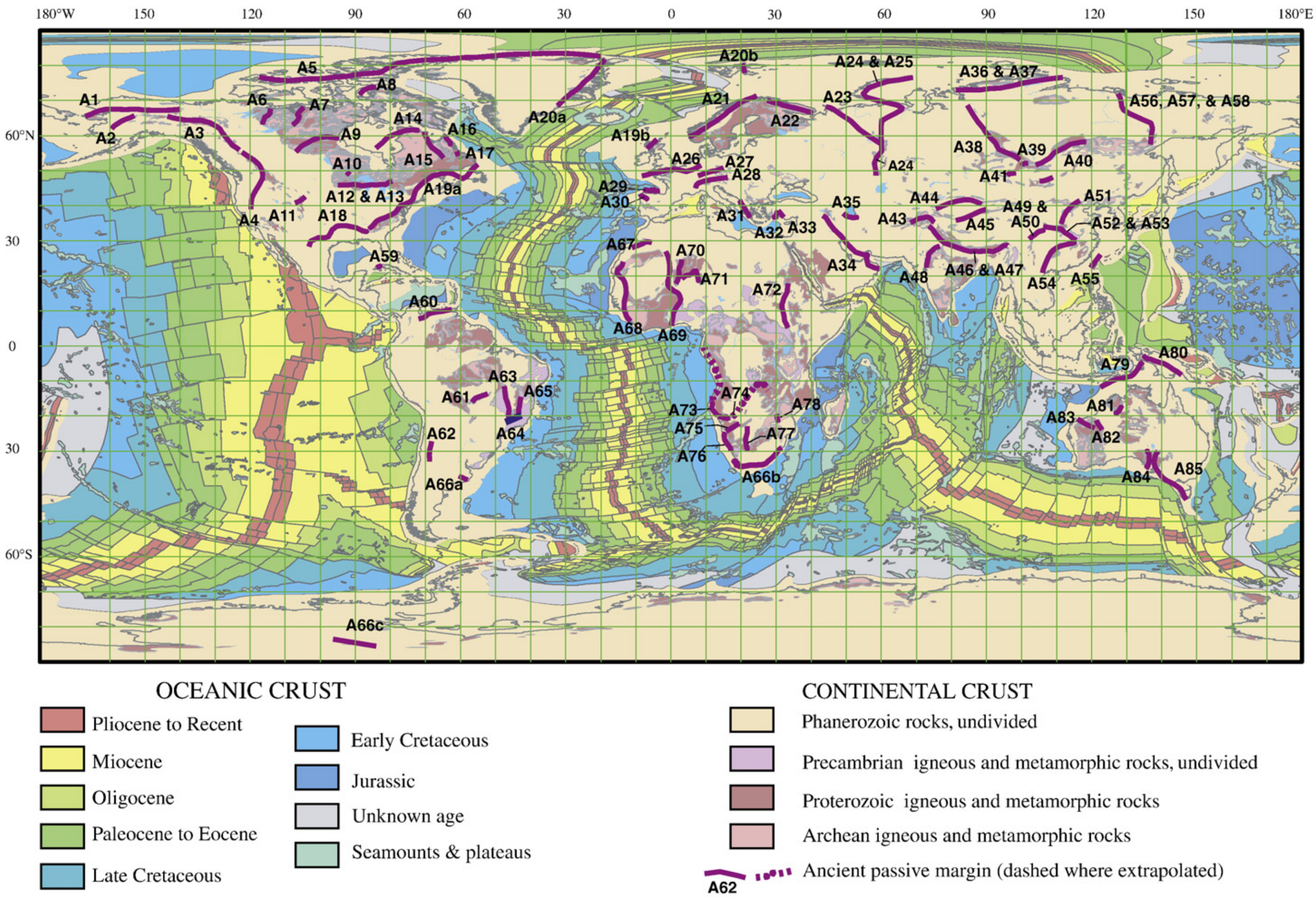


Fig. 6. World map showing ancient passive margins. The ancient margins are labeled A1 to A85 from northwest to southeast. A few superimposed passive margins are shown as a single line but have more than one number. Base map from Commission de la Carte Géologique du Monde (2000).

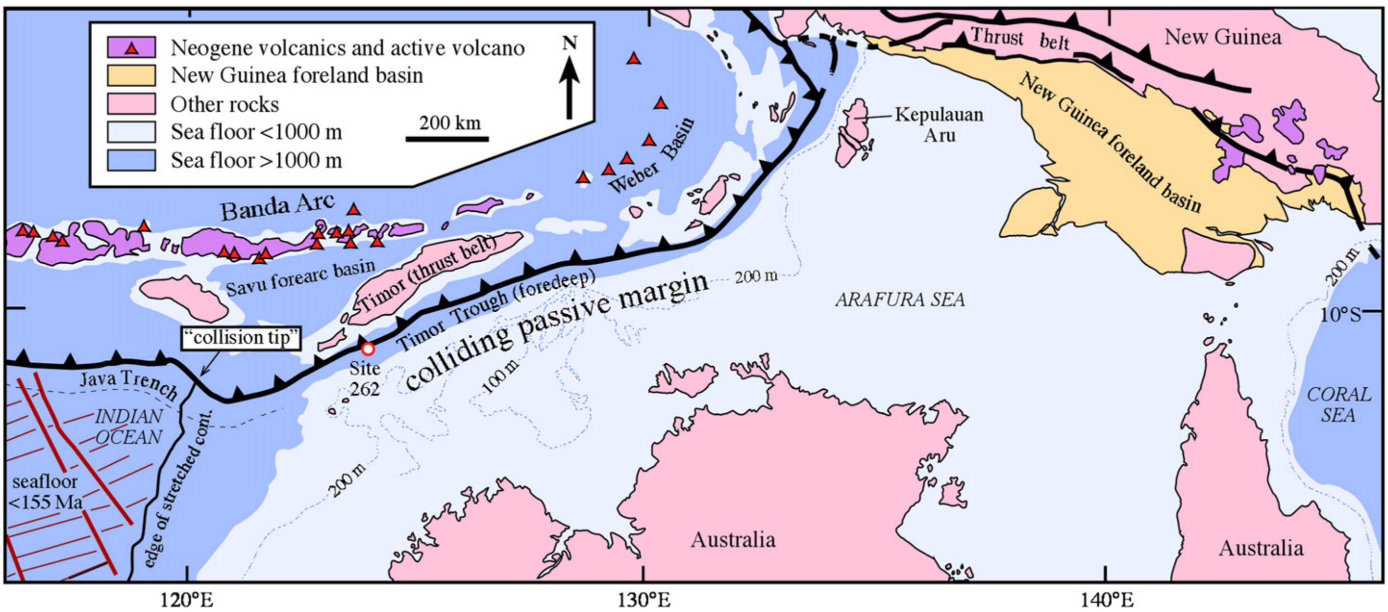


Fig. 7. Generalized tectonic map of the ongoing collision zone between Australia's northwestern passive margin and the Banda Arc. West of the point labeled "collision tip", oceanic crust of the Australian plate is subducting beneath the Banda Arc at the Java Trench, and the Australian passive margin is obliquely approaching the subduction zone. Immediately east of the collision tip, collision has just begun and involves stretched continental crust. Farther east, collision has been underway since 4 Ma and a collisional orogen has grown to heights approaching 3 km. Timor Trough is an underfilled syncollisional foredeep that can be traced into the Java Trench. Adapted from Hamilton (1979).

basement and in thrust sheets (Fig. 8a). Upper Neoproterozoic rift-related sedimentary and igneous rocks are recognized at many places along the length of the margin, and are attributed to rifting on the basis of sedimentary facies, age, igneous components, and chemistry (Rankin et al., 1989; Williams et al., 1995). In Newfoundland (Fig. 8b), Grenville basement is unconformably overlain by upper Neoproterozoic arkoses, conglomerates, quartzites, graywackes, and columnar and pillowed lavas (Williams et al., 1995). Abrupt changes in thickness and stratigraphic order are consistent with local, fault control of deposition. Tholeiitic mafic dikes are related to the flows and attest to significant extension; the dikes strike northeast, parallel to the eventual passive margin (Williams et al., 1995). Rift-related igneous rocks are as old as 617±8 Ma (Hare Hill alkaline granite; Williams et al., 1995), and as young as 550.5+3/−2 Ma (lavas of the Skinner Cove Formation; Cawood et al., 2001). The youngest detrital zircons in rift-related strata, at 572 Ma (Cawood and Nemchin, 2001), are not much older than the youngest lavas.

The succeeding Cambrian to Ordovician platformal sequence is one of the world's classic miogeoclinal prisms. The platformal strata thicken to the east, deepen to the east, and blanket slightly older rift-related rocks—all pointing toward a passive-margin setting. In Newfoundland, the lower part of the sequence consists of sandstone, siltstone, shale, and limestone, and the upper part is limestone and dolomite. Being gradual, the transition from extension-driven to thermally driven subsidence is impossible to pinpoint, but clearly it had taken place by late Early Cambrian (ca. 520 Ma; Knight et al., 1995,

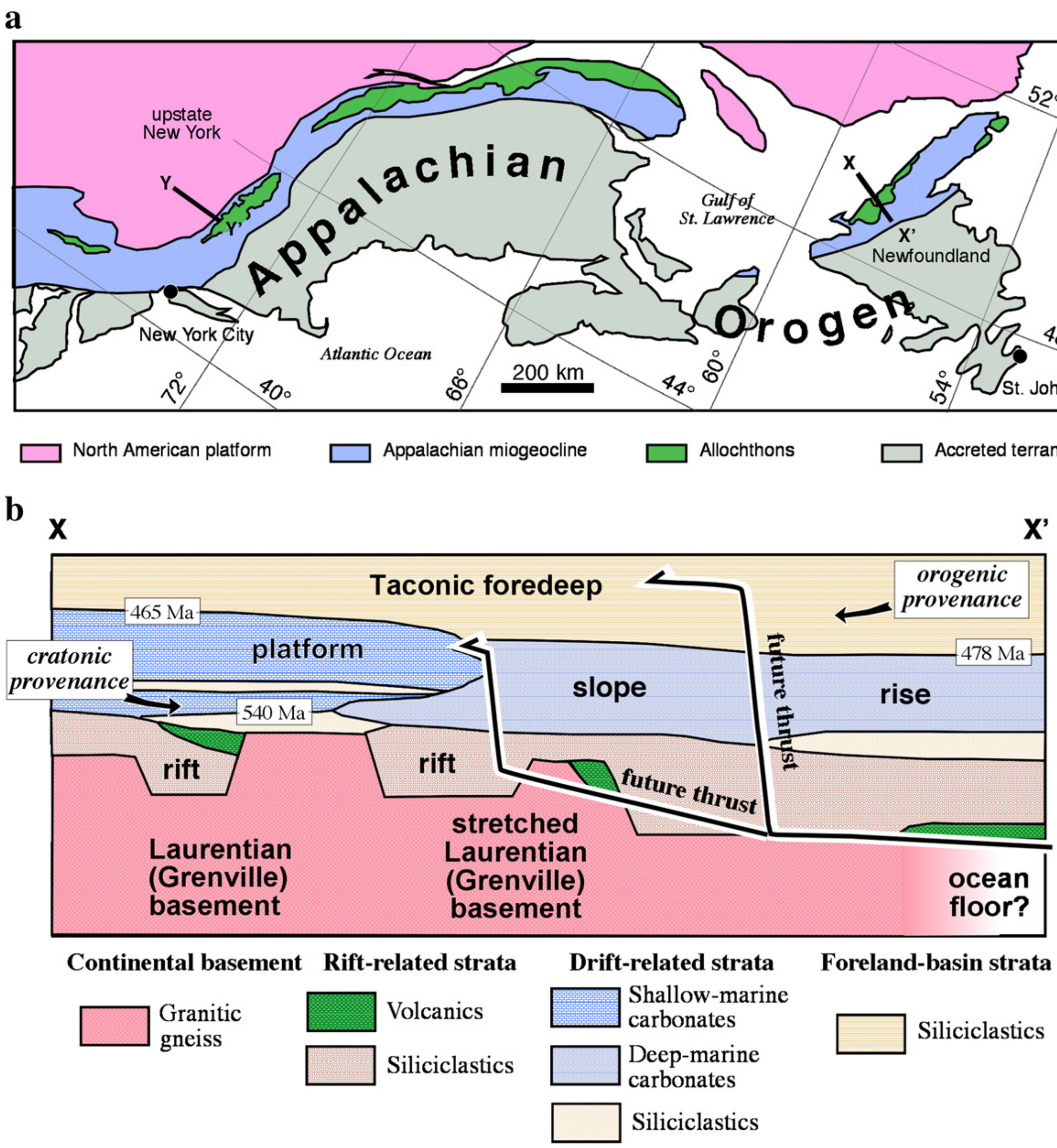


Fig. 8. (a) Map of the Northern Appalachian orogen showing the lines of two cross-sections, from Cawood and Nemchin (2001). (b) Stratigraphic cross-section across the Cambrian–Ordovician passive margin in Newfoundland, adapted from Cawood and Nemchin (2001). Note the sequence of rift-passive margin, and foreland basin, and the west-to-east deepening in the passive-margin succession. (c) Stratigraphic cross-section of the foreland basin in northern New York emphasizing age relations, adapted from Bradley (1989) and Landing and Bartowski (1996). The foreland-basin succession is diachronous, younger toward the craton; this is a hallmark of migrating foredeeps related to arc–passive margin collision. The flysch sequence coarsens and thickens upward and eastward, and the most proximal deposits are olisotostromal melanges that directly flank the thrust front and date thrust emplacement. The column on the right shows allochthonous deep-water deposits of the passive margin, where sedimentation was continuous and synorogenic flysch sedimentation commenced earlier than on the former platform. Ages of Events 1 through 4 here are a few million years older than in Newfoundland.

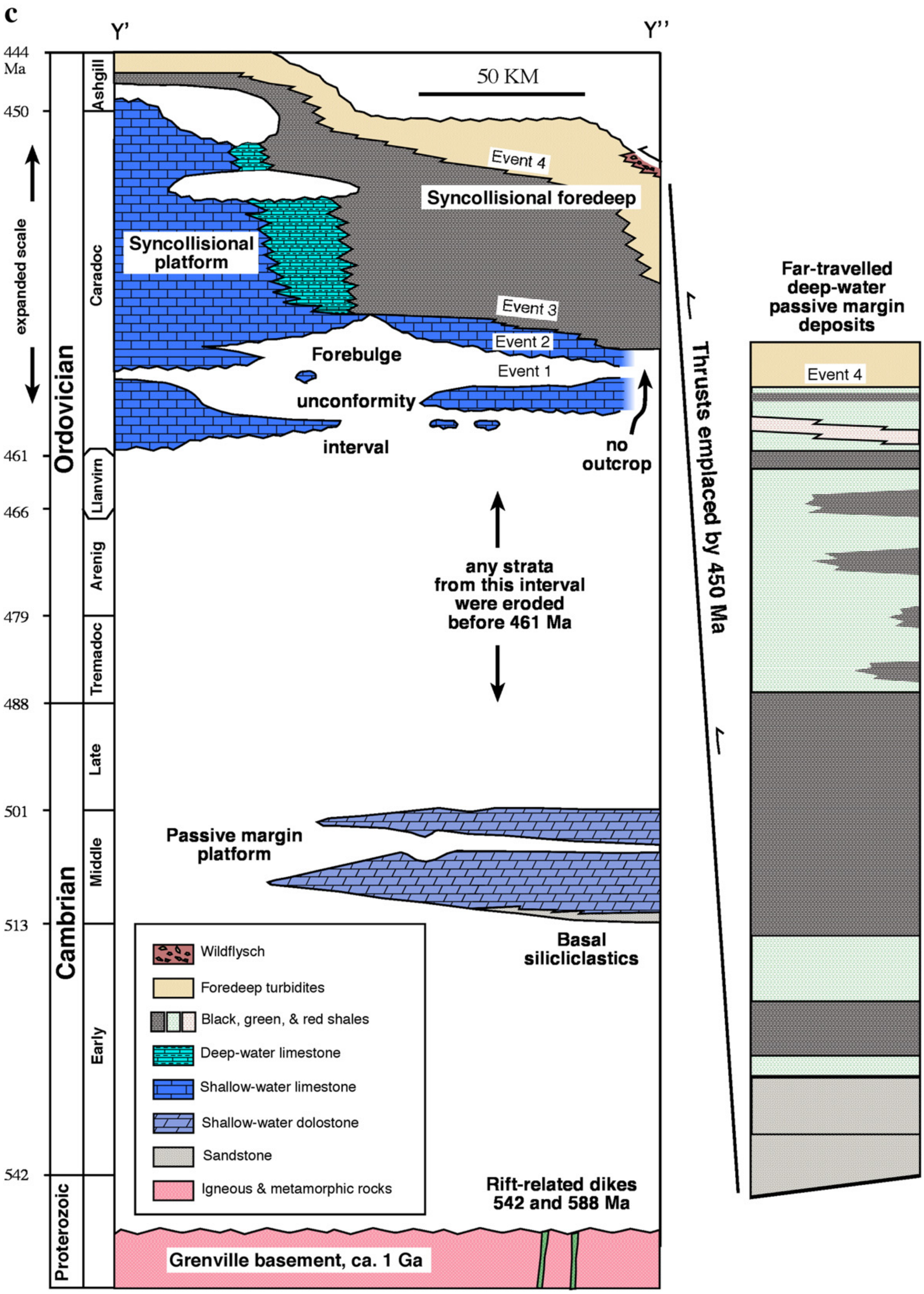


Fig. 8 (*continued*).

p. 85). In light of the bracketing ages of 550 and 520 Ma, the rift-drift transition is here placed at ca. 540 Ma.[2] Paired with the shallow-water platform was an area of coeval, deep-water (slope-rise) deposition (Fig. 8b). These strata are important because they establish that the platform faced deep water for many tens of millions of years. The deep-water rocks occur in thrust sheets (allochthons in Fig. 8a) that were transported cratonward over age-equivalent platformal strata, which is a pattern commonly seen in ancient arc–passive margin collisions.

The collisional demise of the passive margin, known as the Taconic (or Taconian) orogeny, is recorded stratigraphically in both the original

[2] Similarly, Cawood et al. (2001) picked the age of the rift-drift transition at 540–535 Ma. Bond et al.'s (1984) much older age pick for the rift-drift transition in the Appalachians (600±25 Ma) was based on subsidence curves calibrated to a long-abandoned time scale.

deep-water and shallow-water realms. In the deep-water belt, impending collision was signaled by an influx of outboard-derived siliciclastic turbidites; as reviewed by Bradley (1989), this event took place in late Arenig time (Early Ordovician, ca. 478 Ma) in Newfoundland. Assuming that the most distal deep-water sections were deposited on oceanic crust, this transition would predate collision along that part of the margin. On the carbonate platform, the slightly younger event sequence was: (Event 1) epeirogenic uplift and emergence, (Event 2) renewed carbonate deposition, deepening upward into (Event 3) black shale deposition as the platform drowned, and then (Event 4) deposition of easterly-derived turbidites, which were shed from a submarine thrust belt that advanced from the east. Fig. 8c shows the diachronous nature of facies belts and the corresponding numbered events within the Taconic foreland, as revealed by a transect across northern New York (the equivalent strata in Newfoundland are under water). Event 1 has been attributed to uplift along a forebulge some 100–200 km landward of the orogenic front (*e.g.*, Knight et al., 1991). Events 2 and 3 were accompanied by margin-parallel normal faulting related to flexure of the passive-margin plate into the trench (Bradley and Kidd, 1991), perhaps augmented by whole-lithosphere extension caused by slab pull resisted by continental drag (Schoonmaker et al., 2005). Event 4 represents the Taconic foreland basin, which in Newfoundland comprises as much as nearly 2 km of deep-water turbidites ("flysch") (Hiscott, 1995). Platform drowning in Newfoundland took place in Llanvirn time (ca. 465 Ma; see Bradley, 1989, for original sources) and this is picked as the age of the passive margin-to-foreland-basin transition in Newfoundland. It should be stressed, however, that the time elapsed between Events 1 and 4 at any given location was only a few million years.[3]

The thrust allochthons that overrode the passive margin carried rift-, passive margin-, and foreland-basin deposits, all of them native to the Laurentian margin. Higher in the thrust stack are ophiolites, which unequivocally record the former existence of an ocean basin (Iapetus) to the east of the Laurentian margin. The ophiolites, including the very well-studied Bay of Islands ophiolite, have yielded U–Pb zircon ages from 508±5 to 484±5 Ma (see Williams et al., 1995 for original sources). Thus the ophiolites are substantially younger than the oldest seafloor (ca. 540 Ma) inferred to have lain immediately offshore Newfoundland as the oceanic part of the Laurentian plate. The Bay of Islands ophiolite is instead interpreted to have formed by seafloor spreading in a supra-subduction setting (*e.g.*, Jenner et al., 1991). Amphibolites from the metamorphic sole of the Bay of Islands ophiolite yielded a $^{40}Ar/^{39}Ar$ amphibole ages of 469±5 and 464±9 Ma (see Williams et al., 1995 for original sources); the metamorphic sole is interpreted as a relict of the paleo-subduction zone during or just before emplacement of the ophiolite onto the passive margin. In both Newfoundland and the northeastern United States, rocks of the former passive margin are now flanked to the east by Ordovician volcanic and plutonic rocks; arc magmatism in both areas spanned pre-, syn-, and immediately post-collisional times (Karabinos et al., 1998; Zagorevski et al., 2006). Waldron and van Staal (2001) presented evidence that the Ordovician arc that collided with Laurentia's passive margin was built on a ribbon microcontinent that had rifted away from Laurentia at ca. 540 Ma. In this model, the *outer* margin of this inferred ribbon microcontinent was a somewhat older passive margin, formed ca. 600–570 Ma (Waldron and van Staal, 2001).

The Taconic collision was accompanied by regional metamorphism of both basement and sedimentary cover of the Laurentian margin. In the northeastern United States, Taconic regional metamorphism reached kyanite grade and took place at 465±10 Ma ($^{40}Ar/^{39}Ar$ amphibole; Laird, in Drake et al., 1989). This age correlates remarkably well with the end date of the passive margin based on stratigraphic evidence. Along the Taconic suture, blueschist and eclogites have been discovered (Laird, in Drake et al., 1989), but only in a few small enclaves that would likely have been overlooked in a less thoroughly mapped region. These high-pressure rocks confirm the presence of a subduction zone during the Taconic collision.

The Cambrian–Ordovician of the Appalachians thus serves as a thoroughly studied example of a passive margin of moderate duration whose start date and end date are constrained by multiple lines of evidence. The age of the rift-drift transition is ca. 540 Ma, bracketed between isotopically dated rift-related igneous rocks and fossil-dated strata near the base of the drift sequence. Even without age control on the igneous rocks, the age of rifting would still be fairly well constrained by the youngest detrital zircons in rift sediments. The passive-margin to foreland-basin transition is even more tightly constrained at ca. 465 Ma, with age control provided by the oldest allochthonous flysch, the age of platform drowning, and metamorphic ages of passive-margin strata. These dates imply a lifespan of about 75 m.y.—relatively short by worldwide standards.

4.3. Verkhoyansk (eastern) margin of Siberia

Between ca. 1600 and 160 Ma, the east side of the Siberian craton (Fig. 9) was flanked by what I interpret as a succession of three passive margins (margins A56, A57, and A58). The first of these lasted longer than any other passive margin–from ca. 1600 to ca. 1010 Ma–and it also stands out as one of the few margins anywhere in the world during that interval. Two younger rift-drift episodes followed, from ca. 550 to 380 Ma and then from ca. 380 to 160 Ma. Each successive passive-margin episode masked any earlier history, and the rocks of each of the three margins were involved in the Mesozoic Verkhoyansk orogeny, which produced the world's widest fold–thrust belt.

The Proterozoic history is pieced together from widely separated outcrop areas along the eastern side of the Siberian craton: a northern Verkhoyansk sector that includes the Olenek Uplift and Kharalaukh Mountains and a southern sector that includes the Sette-Daban range (Fig. 9a). Riphean strata in both areas are the remnants of a miogeoclinal prism of carbonates and siliciclastics that thicken and are more distal to the east (Fig. 9b). Maximum thickness exceeds 5 km (Khudoley et al., 2001). The ca. 1600–1050 Ma miogeoclinal succession in the south (Uchur, Kerpyl, and Lakhanda Groups) depositionally overlies a rift (Ulkian succession; not in the line of section) containing volcanic rocks as young as 1676 Ma (Pisarevsky and Natapov, 2003). Pisarevsky and Natapov (2003) interpreted the mid-Proterozoic miogeocline as having formed along a passive margin, a view supported here.[4] They bracketed the rift-drift transition between ca. 1600 and 1480 Ma in the north. In the south, the rift-drift transition would fall between 1676 Ma, the youngest rift-related volcanic rocks, and 1520 Ma, the oldest K–Ar glauconitic sandstone age from the Uchur Group. For this compilation, I place the rift-drift transition at ca. 1600 Ma in both the north and south.

The new interpretation offered here is that the Mesoproterozoic passive margin was involved in a somewhat cryptic collision at ca. 1010 Ma. In the eastern Sette-Daban, the Lakhanda Group is depositionally overlain by the Uy Group, which consists of siliciclastics that were derived from both western (Siberian) and eastern sources (Khudoley et al., 2001) (Fig. 9b). These stand out as the first easterly-derived sediments in the Riphean succession, consistent with development or arrival of a new, outboard sediment source. Moreover, thicknesses and facies in the Uy Group are locally controlled by thrust faults (Khudoley and Guriev, 2003). The Uy Group has yielded detrital zircons as young as 1070±40 Ma, and is intruded by mafic sills that are as old as 1005 Ma (Rainbird et al., 1998). A ca. 400-m.y. gap in the

[3] Events 1–4 are diachronous both across strike and along the entire length of the Appalachians (Bradley, 1989); the age progression of platform drowning is interpreted as tracking plate convergence through time along an irregular margin.

[4] An alternative view is that the mid-Proterozoic miogeocline represents a sag sequence above a failed intracratonic rift (Khudoley et al., 2001).

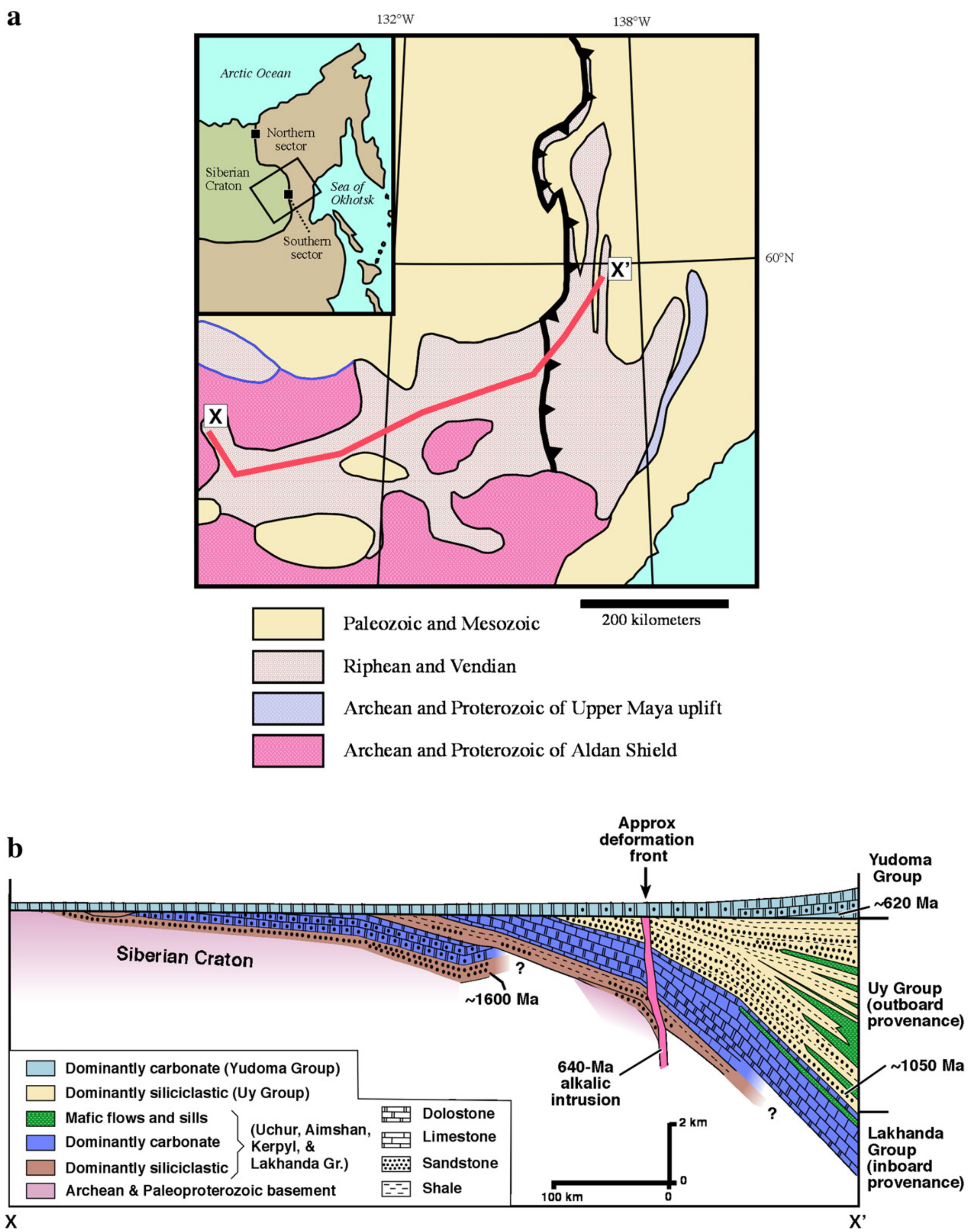


Fig. 9. (a) Generalized map of the Verkhoyansk region and eastern margin of Siberian craton, showing key features and localities mentioned in text, from Khudoley et al. (2001). (b) Stratigraphic reconstruction of Mesoproterozoic and Neoproterozoic rocks of the Siberian margin, modified from Khudoley et al. (2001). The Uy Group is here reinterpreted as a foreland-basin succession, related to orogenesis on the basin's eastern margin.

sedimentary record followed deposition of the Uy Group. The next younger rocks are Vendian strata beginning at ca. 620 Ma, which overlie an angular unconformity (Khudoley et al., 2001) that buries thrust faults. Together, these observations are consistent with the interpretation that the Uy Group is a foreland-basin succession and that the Mesoproterozoic phase of the Siberian craton's eastern passive margin ended with orogeny at ca. 1010 Ma. Whereas there is no known 1 Ga orogen east of the Sette-Daban ranges, such a feature would have departed when the next passive margin formed in the late Neoproterozoic, leaving only the foreland basin to record the collision.

An alternative view is that ca. 1000 Ma was a time of rifting (Khudoley and Guriev, 2003). Khudoley et al. (2001) interpreted an extensional tectonic setting for the Uy Group, presumably to account for mafic magmatism at the time. However, it is unclear why such rifting would be followed by 400 million years of non-deposition rather than by thermal subsidence. Instead, I suggest that the mafic sills of the Uy Group were intruded in a foreland setting and not a rift—the same interpretation invoked by Hoffman (1987) for several Paleoproterozoic examples in Laurentia.

Adopting the passive margin and collision interpretation, the start date for the Mesoproterozoic margin would have been at ca. 1600 Ma and the end date would lie somewhere between 1070 and 1005 Ma, and probably close to the younger bracketing age. An end date of 1010 Ma would correspond to a lifespan of about 590 m.y.

Two additional phases of passive-margin evolution have been suggested along the eastern margin of the Siberian craton—Neoproterozoic to Devonian and Devonian to Jurassic. These are treated as separate passive margins (margin A57 and A58) because they appear to have formed by completely separate plate-tectonic events.

In northeastern Siberia, Pelechaty (1996) documented a Neoproterozoic carbonate succession that formed along the eastern, presumably passive margin to the Siberian craton. This >600-m-thick carbonate succession (Khorbosuanka Group) predates a rifting event at around the Neoproterozoic–Cambrian boundary related to a breakup event along the Siberian craton's *northern* (Taimyr) margin (Pelechaty, 1996) (margin A37). The base of the Khorbosuanka Group is ca. 550 Ma (Pelechaty et al., 1996), my pick for the age of the rift-drift transition. Khudoley and Guriev (2003, p. 34) cited evidence from subsidence analysis for multiple rifting events between 570 and 520 Ma, of unknown significance.

The Neoproterozoic to Devonian phase of passive-margin evolution ended not with collision, but with re-rifting. This final rifting event along the eastern margin of the Siberian craton has been interpreted to have involved the separation and departure of a Lomonosov-type continental sliver (Parfenov, 1991; Sengör and Natalin, 1996, p. 554). Rifting began in the Middle Devonian in the Sette-Daban zone and was widespread along the Siberian craton's eastern margin by the Late Devonian (Khudoley and Guriev, 2003). Khudoley and Guriev (2003) reviewed evidence for rifting in the Sette-Daban zone (*e.g.*, Famennian half-grabens filled with olistostromes). Lower Carboniferous limestones were uniformly deposited across the extended terrane and are interpreted to record post-rift thermal subsidence. The margin lasted until Late Jurassic when shortening began in the Verkhoyansk foldbelt (Sengör and Natalin, 1996, p. 556). Using ages of 380 Ma for the rift-drift transition and 160 for the passive margin to foreland-basin transition, the margin had a duration of about 220 m.y. Interestingly, these are quite close to the age picks for the Brookian margin of the Arctic Alaska terrane (margin A1).

4.4. Western margin of Kaapvaal craton

The Kaapvaal craton of South Africa (Fig. 10) is one of the world's oldest continental nuclei. Development of a Neoarchean passive margin along its western side (margin A77) is inferred both from outcrop data and from high-quality seismic-reflection profiles. This margin is highlighted here because it is one of the two oldest well-dated margins, and because one aspect of the tectonic interpretation offered here is new. The Ventersdorp Group (Fig. 10b), which consists of komatiites, continental flood basalts, chemical and clastic sedimentary rocks, and felsic volcanic rocks, has been interpreted to have been deposited during the initial rifting along this margin (Tinker et al., 2002). Zircon ages from the lower part of the Ventersdorp Group are 2709±8 and 2714±8 Ma (U–Pb; Armstrong et al., 1991). Rocks of the overlying Schmidtsdrif and Campbellrand Subgroups are interpreted to represent the thermal subsidence phase of passive-margin sedimentation. These strata were deposited in a westward-thickening, west-dipping sedimentary prism (Tinker et al., 2002) and consist mainly of shallow-marine carbonate rocks (Beukes, 1984, 1986). The base of the Schmidtsdrif Subgroup is reasonably well dated by a zircon age of 2642±3 Ma (Pb-evaporation; Beukes cited in Tinker et al., 2002).

There is a lack of consensus as to what marks the demise of the passive margin. I suggest that the contact between the Campbellrand Subgroup and the overlying Asbestos Hills Subgroup (Fig. 10b) approximates this event. The Asbestos Hills Subgroup is a transgressive succession of iron formation (Kuruman and Griquatown Formations), and it is overlain by mudstone, siltstone, quartz wacke, and iron formation of the Koegas Subgroup. Hoffman (1987) interpreted similar successions (carbonates overlain by iron formation overlain by siliciclastics) along the Paleoproterozoic Wopmay (A6) and Animikie (A13) margins as recording the transition from passive margin to foreland basin. Applying a similar interpretation to the Kaapvaal margin, a fairly close age constraint for the transition is provided by a zircon age of 2465±7 Ma (Armstrong et al., 1991) from the Kuruman Formation at the base of the Asbestos Hills Subgroup. This would imply that the final position of the deformation front lay approximately 100 km to the west of any present exposures. Ages of 2640 Ma for the rift-drift transition and 2470 Ma for the demise of the passive margin yield a lifespan of 170 m.y.

Other possibilities for the age of the demise bear mention. Tinker et al. (2002) inferred that rocks of the Asbesheuwels, and Koegas Subgroups were deposited in the same regime of thermal subsidence as the underlying Campbellrand Subgroup. Whereas Tinker et al. (2002) did not specifically comment on the demise of the passive margin, their interpretation would mean that the margin endured somewhat longer than is suggested here. A much longer lifespan (about 640 m.y.) is also conceivable, although less likely. The first significant shortening of rocks of the Campbellrand Subgroup took place at ca. 1900 Ma, after deposition of the Olifantshoek Supergroup (Fig. 10b), which consists of siliciclastic and some volcanic rocks dated at 1928±4 (Pb–Pb zircon; Cornell et al., 1998). If this deformation records the end of the passive margin (a possibility raised by Maarten de Wit, oral communication, 2003), the western margin of the Kaapvaal craton would then rank as the longest-lived passive margin in Earth history.

5. Distribution of passive margins through time

The distribution of passive margins has been quite uneven through time (Fig. 5a). The peaks and valleys in the histogram serve to subdivide Earth history, the peaks representing times of relatively more continental dispersion, and the valleys times of continental aggregation. The passive-margin age distribution will be compared with various proposed supercontinents in Section 8.

The youngest group of margins in Fig. 5a extends from ca. 300 Ma to present. This includes all of the present-day passive margins as well as 13 ancient ones. The ancient margins alone show a young peak centered near 100 Ma, but this is purely an artifact that disappears when the modern passive margins of the same vintage are also included (Fig. 5a). The comparative abundance of modern passive margins is mainly due (1) to the ease with which they can be recognized, and (2) to the fact that the marine magnetic anomalies allow subdivision into sectors having different start dates. For example, the present-day eastern passive margin of Africa has been divided into five parts, but would probably end up being treated as a single margin if it was half a billion years old and now preserved in an orogenic belt.

Working back through time, the next feature of note is the pronounced late Paleozoic minimum (Fig. 5a). Only 5 passive margins can be counted during the time between 300 and 275 Ma, compared to 17 margins at the next older peak at ca. 530–505 Ma.

The next older age cluster of margins in Fig. 5a extends from 1000 to 300 Ma and has twin peaks at 530–505 Ma, and at 600–580 Ma. No single margin endured for this entire 700 m.y. interval. The younger portion of this interval includes the classic passive margins that formed on all sides of Laurentia near the Neoproterozoic–Cambrian boundary and were destroyed during subsequent collisions: the Franklinian (margin A5), Appalachian (margin A19a), Ouachita (margin A18), and northern and southern Cordilleran (margins A3 and A4). The earlier portion of this interval includes the Neoproterozoic margins of the West African craton (margins A67, A68, and A69), Sao Francisco craton (margins A63 and A64), Congo craton (margins A73 and A74), and

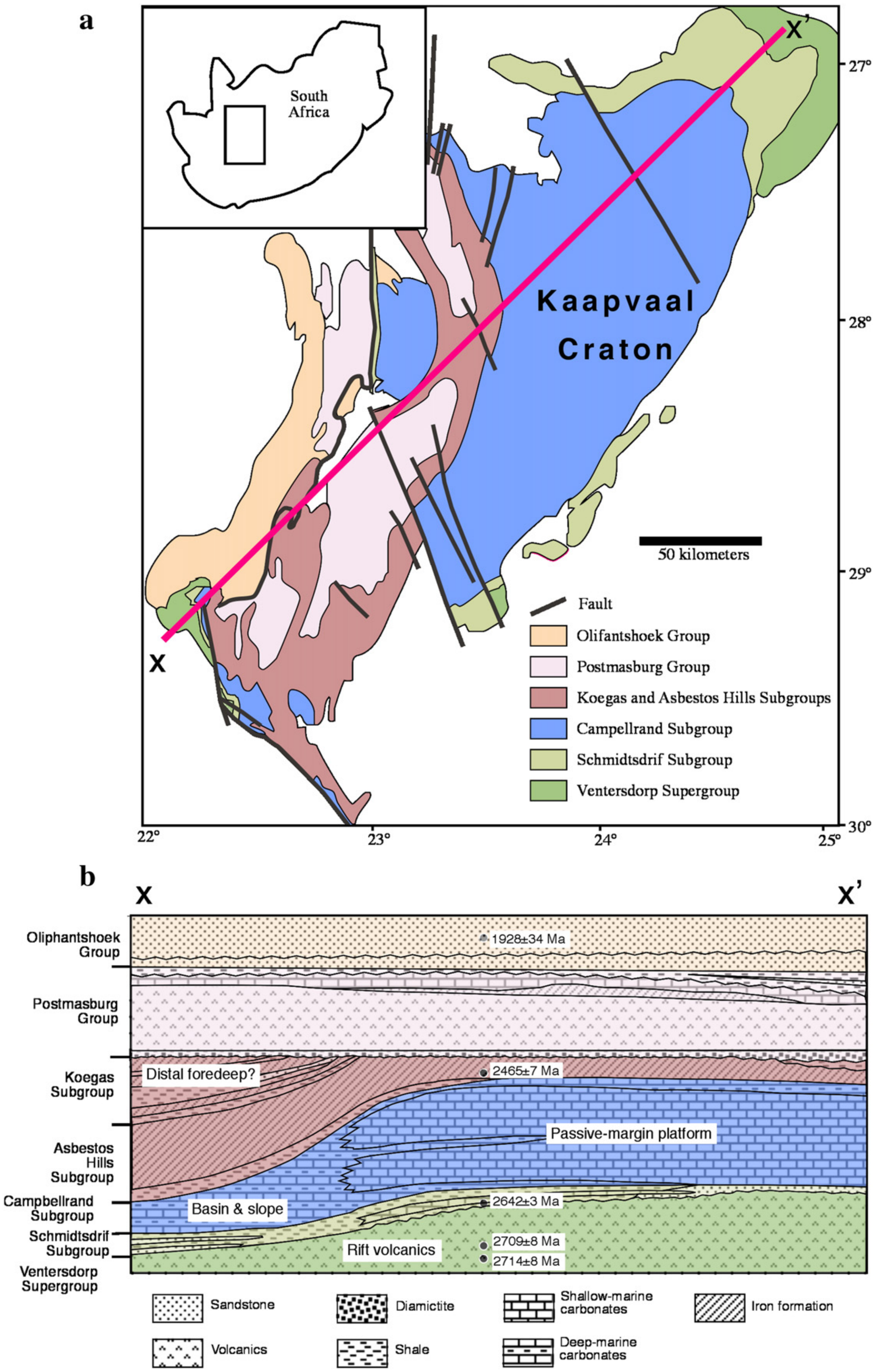


Fig. 10. (a) Generalized map of the western margin of the Kaapvaal craton, from Harding (2004). (b) Stratigraphic cross-section across the margin, from Harding (2004). Note that the siliciclastic Koegas Group is confined to areas of former off-shelf deposition, a pattern consistent with deposition in a foredeep related to a collision farther to the west.

Kalahari craton (margins A75 and A76), that were deformed in various Pan-African collision during the assembly of Gondwana.

Between 1740 and 1000 Ma, an interval approximately corresponding to Holland's (2006) "boring billion", passive margins were few in number, and from about 1740 to 1600 Ma, there appear to have been none at all. The Mesoproterozoic eastern margin of Siberia, which lasted some 590 m.y., has already been discussed. The western and southern margins of the Siberian craton (Yenisei—margin A38; and Baikal—margin A40) also fall in this age range, as do the eastern and northern margins of the Baltic craton (Uralian phase 1—margin A24; and Timan—margin A23). These five appear to be the longest-lived margins in Earth history, although they have poor quality rankings that reflect inadequate age controls and in a few cases, debatable tectonic interpretations.

The oldest well-defined peak in the passive-margin record is centered at 1900–1890 Ma and spans a range from ca. 2445 Ma to ca. 1740. Its apex corresponds to a time of continental dispersal and its end presumably corresponds to an aggregation of smaller continents into a larger one. Included in the 2445–1740 Ma group are excellent examples of passive margins bordering the Slave craton (Wopmay and Thelon; margins A6 and A7) and the Superior craton (Cape Smith-Trans-Hudson, and New Quebec; margins A14 and A15).

The oldest well-documented passive margins are the western margin of the Kaapvaal craton (margin A77, 2640–2470 Ma, discussed in Section 4.4), and the southwestern margin of the Pilbara craton (margin A83, 2685–2440 Ma). The Central Orogenic Belt of the North China craton may include an even older passive margin (margin A51, 2740–2530 Ma), but the tectonic interpretation is debatable. Together these define the oldest, minor "peak" in Fig. 5a. The oldest proposed passive margin, Steep Rock Lake (margin A10), existed sometime between 3000 and 2800 Ma but its tectonic interpretation is controversial and age controls are inadequate. If correctly interpreted by Kusky and Hudleston (1999), this margin would attest to something resembling a modern-style Wilson Cycle between 3000 and 2800 Ma.

No passive margins are known that are older than ca. 3000 Ma. This may be real, but could also involve difficulties in recognition, and lack of preservation. It would be difficult to correctly read the history of a short-lived rifted margin at the edge of a narrow Archean protocontinent that faced into a narrow ocean that opened and then immediately closed. Recycling of ancient crust could also contribute to the lack of pre-3000-Ma passive margins. Thus, the lack of margins before 3000 Ma in Fig. 5a is not definitive.

A key question is whether or not the irregular age distribution of passive margins as shown in Fig. 5a is real, or merely an artifact of an incomplete compilation. Comparisons with other secular trends, discussed in Section 9, suggest that the passive-margin age distribution is robust.

Fig. 5b shows the ages of rift-drift transitions for the 76 ancient margins in Group 1, plus all 75 modern ones. This plot reveals peaks at 500–550 and 600–650 Ma that are not obvious from Fig. 5a alone. A plot of rifting events by Condie (2002) shows a strong peak at 850–800 Ma that is barely evident in Fig. 5b; the difference between this plot and Condie's may lie in the fact that in the present study I picked the ages of rift-drift transitions, and was not too concerned with the full age range of prior rifting. Fig. 5c shows collision ages for the 81 ancient margins in Groups 1 and 2, with peaks at 300–400 and 550–600 Ma. This plot differs somewhat from a plot of collisions by Condie (2002), which shows 11 of them between 1300 and 1000 Ma, compared to just two in Fig. 5c. The discrepancy can be traced to Condie's inclusion of collisions that did not involve passive margins.

6. Lifespans of passive margins through time and implications for the tempo of plate tectonics

The overall *distribution* of passive margins through time shown in Fig. 5a is considered to be approximately correct. In contrast, corresponding plots of the *lifespans* of these margins (Fig. 11a and b) are significantly less reliable. Consider, for example, the Timanian margin of northern Baltica (margin A23). There is little doubt that it was a passive margin during the Neoproterozoic (*e.g.*, Roberts et al., 2004). The picked end date at 558 Ma seems quite reliable, as it is based on the U–Pb zircon age of a tuff from low in the foreland-basin section. On the other hand, the ca. 1000 Ma start date is extrapolated from Baltica's southern Uralian margin (margin A24), which appears to have been continuous with the Timanian margin. The Uralian margin's start date itself is equivocal and approximate. Even if the exact start date was off by 100 m.y., the Timanian margin would still contribute to the same overall peak in Fig. 5a, albeit in a different number of bins. But 100 m.y. would make a big difference to the lifespan of the margin (Fig. 11).

Keeping these caveats in mind, the lifespans of the 76 ancient passive margins in Group 1 are shown in Figs. 11 and 12. They range from about 20 to about 590 m.y. The mean lifespan of all ancient margins is 181 m.y. Subdividing the margins according to the natural age groupings suggested by Fig. 5a: (1) Seventeen passive margins that formed during the Neoarchean to Paleoproterozoic (2800–1600 Ma) have a range of estimated lifespans from 20 to 370 m.y. Their mean lifespan is a surprisingly long 186 m.y. (2) Only six margins began in the Mesoproterozoic (1599–1000 Ma); these have lifespans that range from 55 to 590 m.y. and a mean lifespan of 394 m.y. (3) Twenty-five margins with start dates in the Neoproterozoic (999–542 Ma) have lifespans ranging from 25 to 370 m.y. and a mean lifespan of 180 m.y. (4) Fifteen margins that began in the early and middle Paleozoic (Cambrian to Carboniferous) have a range of lifespans from 57 to 220 m.y. and a mean of 137 m.y. (5) Thirteen ancient margins that began in the Late Permian or thereafter have a range of lifespans of 23 to 219 m.y. and a mean of 130 m.y. (6) Finally, as noted before, the mean (partial) age of the 75 modern margins is 104 Ma and the range is 5 to 180 m.y.

At face value, the data plotted in Figs. 11 and 12 suggest that passive margins generally had longer lifespans in the Precambrian than in the Phanerozoic. This conclusion needs to be balanced against the fact that the lifespans of the modern margins are still incomplete. The Atlantic margin of West Africa, for example, is already about 170 million years old today, and it will endure for many tens of millions of years to come, until it finally collides with the Caribbean arc. A more apt comparison can be made by eliminating the modern margins, and the ancient margins of about the same vintage (*i.e.*, those that formed from Late Permian onward.) Applying this filter, the 15 Phanerozoic margins with Cambrian to Carboniferous start dates had a mean lifespan of 135 m.y., compared to the 47 margins with Precambrian start dates, which had a far longer mean lifespan of 206 m.y.

When beginning the present study, I anticipated the opposite result—shorter lifespans in the Precambrian. The shorter-lifespan hypothesis had been posed by Hoffman and Bowring (1984) and later by Grotzinger and Ingersoll (1992), as a corollary of the hypothesis that the hotter Earth of the Precambrian dissipated its heat via smaller, faster plates. Given the decline in global radiogenic heat production through Earth history, it had long been suggested that the global rate of seafloor production would have been higher in the Precambrian than the Phanerozoic (*e.g.*, Burke et al., 1976; Nisbet and Fowler, 1983; Hargraves, 1986; Pollack, 1997). At 2500 Ma, heat production would have been about two times the present value, and as recently as 1500 Ma, it would still have been about 1.5 times the present value. But heat production is not the same as heat loss. Recent thermal modeling by Korenaga (2006) has provided a strong theoretical case for an early Earth characterized by *sluggish* plate tectonics. For supporting geologic evidence, Korenaga (2006) cited the perceived slowdown in the supercontinent cycle over time. As discussd in Section 8, published supercontinent scenarios are debatable. Results of the present study provide stronger corroboration of Korenaga's (2006) findings: on average, Wilson Cycles were slower, not faster, in the Precambrian.

7. Secular changes in the geology of arc–passive margin collision

Collisions between arcs and passive margins appear to have been happening since the Neoarchean. Whereas many of the older examples are problematic owing to polyphase orogeny and destruction of the sedimentary record, a few are comparable in all important respects to Phanerozoic arc–passive margin collisions. The classic Proterozoic example is the Wopmay orogen along the western margin of the Slave craton (margin A6) (Hoffman, 1980).

7.1. High-pressure, low-temperature metamorphism

Some secular changes, however, are apparent. High-pressure, low-temperature metamorphic rocks form in oceanic subduction zones such as the Franciscan Complex of California, but also in collisional orogens where passive continental margins are partly subducted (Maruyama et al., 1996). Blueschist- and (or) eclogite-facies metamorphic rocks were noted in 17 of the arc–passive margin collisional orogens in the present compilation (Fig. 13a and Table 2). Among the oldest examples is the Neoproterozoic Dahomeyide orogen on the eastern side of the West African craton (Jahn et al., 2001) (margin A69). None of the arc–passive margin collisions older than ca. 625 Ma is known to have produced blueschists that are now exposed at the surface, whereas roughly half of the collisions since that time did. The long-standing explanation for the lack of old blueschists is that steeper geotherms on the hotter Earth precluded their formation (Maruyama et al., 1996 and references therein).

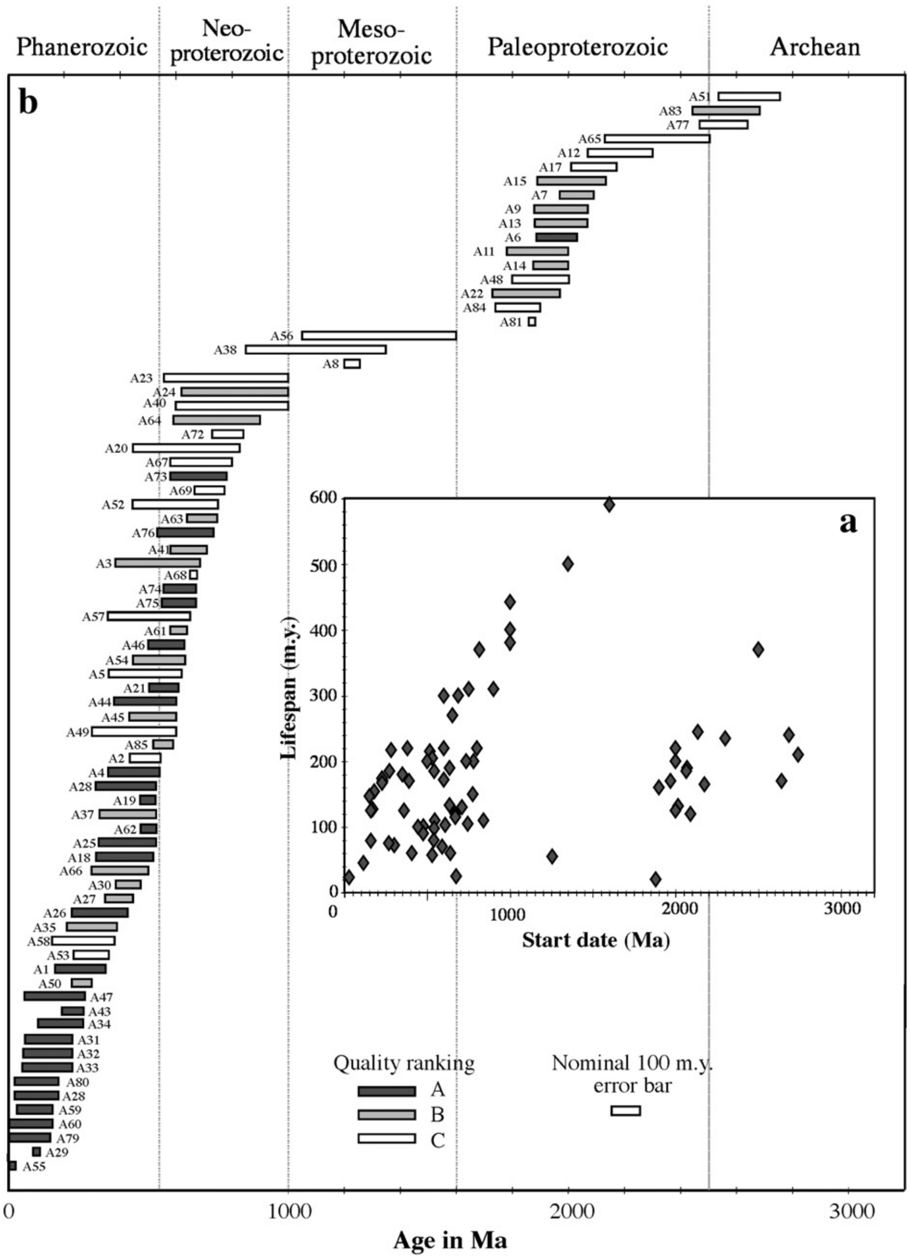


Fig. 11. Lifespans of ancient passive margins displayed in two ways, from data in Table 2. (a) Lifespan plotted against start date, showing the age of each margin at the time of its demise. Note the apparent long-term secular trend toward shorter lifespans over time and the hint of two cycles of declining lifespan. (b) Each margin is plotted as a horizontal bar extending from the start date to the end date; longer bars mean longer lifespans. Bars are arranged chronologically by start date, and keyed by color according to quality ranking. Even the longest-lived passive margins are ephemeral features when viewed at the timescale of Earth history.

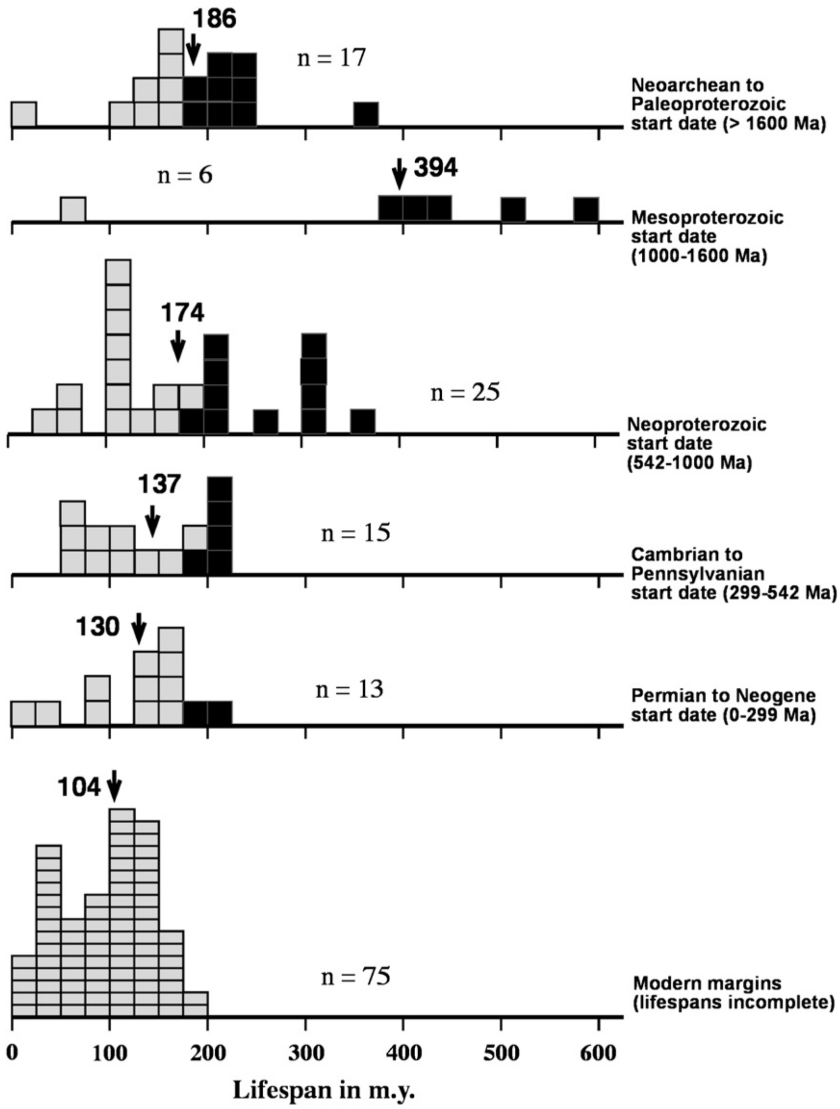


Fig. 12. Histograms showing lifespans of passive margins, grouped according to the natural divisions of geologic time suggested by Fig. 5a. The mean lifespan for each age group is indicated by an arrow and number in m.y. The eighteen longest-lived margins all had start dates in the Archean or Proterozoic. Black bins represent margins that lasted longer than the oldest modern margin, *i.e.*, 180 m.y. or older.

7.2. Foredeep magmatism

Although collisional foredeeps are not generally recognized as sites of magmatic activity, the phenomenon is more common than is generally appreciated. Synorogenic foredeep magmatism was first noticed and discussed at length by Hoffman (1987), who identified predominantly mafic magmatism in the forelands of six Paleoproterozoic collisional orogens in Canada. Two of the best-documented of these are at the margins of the Slave and Superior cratons (*e.g.*, Wopmay and Penokean forelands; margins A6 and A13). In the Penokean foreland, mostly mafic submarine volcanic rocks are interbedded with euxinic, deep-water shales and turbidites (Hoffman, 1987). In the Wopmay foreland, mafic sills are intruded into, and folded with, flysch and molasse facies (Hoffman, 1987).

The present compilation has unearthed some additional examples of foredeep magmatism, bringing the number of recognized instances to fifteen (Fig. 13b and Table 2). The oldest is along the Pilbara craton's southwestern margin (margin A83), where the Woongarra Volcanics (ca. 2491 Ma) were deposited immediately before the Boolgeeda Iron Formation, which is thought to have been deposited on a forebulge (Martin et al., 2000). As Hoffman (1987) recognized, the youngest instance of foredeep magmatism is the Penghu Islands volcanic field in the foreland of the Neogene Taiwan collision (margin A55). In addition, it seems likely that many instances of foredeep magmatism have been overlooked, because mafic magmatism has often been interpreted as *prima facie* evidence for rifting.

Although at first it seemed to be mainly a Proterozoic phenomenon, the incidence of foredeep magmatism does not show a clear secular trend (Fig. 13b). Proterozoic occurrences do, however, seem to be larger. The general problem of magma genesis in or below the lower (foreland) plate of collision zones has not been systematically tackled, and it remains an important problem nearly two decades after Hoffman's (1987) seminal paper.

8. Comparisons with postulated supercontinents

"*Supercontinents* are assemblies of all or nearly all the Earth's continental blocks" (Rogers and Santosh, 2003). A *supercraton* (Bleeker, 2003) is an assembly of several continents, but not all of

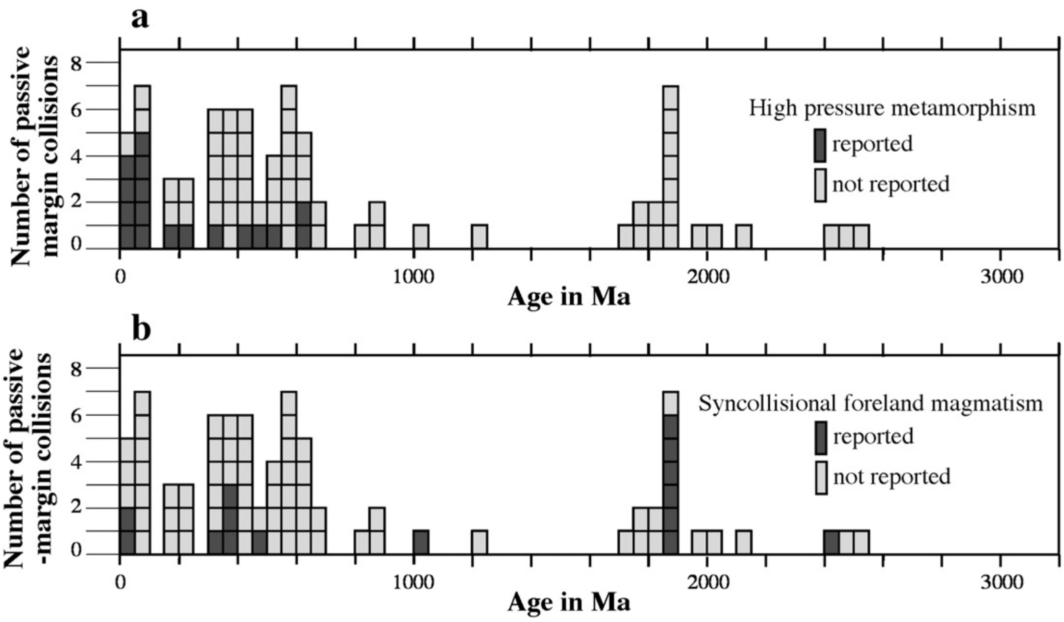


Fig. 13. Histograms showing end dates of passive margins, highlighting instances of (a) syncollisional blueschist metamorphism, and (b) syncollisional foredeep magmatism. From data in Table 2.

them. From first principles, passive margins should relate to the supercontinent cycle as follows: (1) the assembly of a supercontinent should correspond to a decrease in the world's population of passive margins; (2) during the tenure of a supercontinent, relatively few passive margins are to be expected (because the global length of continental margins of all sorts would be reduced); and (3) during supercontinent breakup, passive margins should increase in number. Fig. 14e shows the age distribution of hypothesized supercontinents and other large continental groupings, modified from Rogers and Santosh (2003). Of the variables that are compared with the passive margin age distribution in Fig. 14, supercontinents are not as independent, because various ones have been postulated, in part, on the basis of passive margins. A comparison between Fig. 14a and e shows many of the expected correspondences but also discrepancies for which some tentative explanations are offered.

8.1. Pangea

The most striking correlation between Fig. 14a and e involves Pangea, which formed and broke up during that part of Earth history having by far the clearest geologic record. The maximum of extent of Pangea (ca. 310 to ca. 180 Ma; from reconstructions in Scotese, 2004) coincides with the low in the passive-margin age distribution at 300 to 275 Ma. The increase in passive margins since then clearly equates to the breakup of Pangea. Likewise, the decline in the number of passive margins from ca. 500 to ca. 350 Ma coincides with the assembly of Pangea. The existence of Pangea is unassailable, and the correlation with the passive-margin record is excellent. The supercontinent record before Pangea is contentious and correlations with the passive-margin record are likewise uneven.

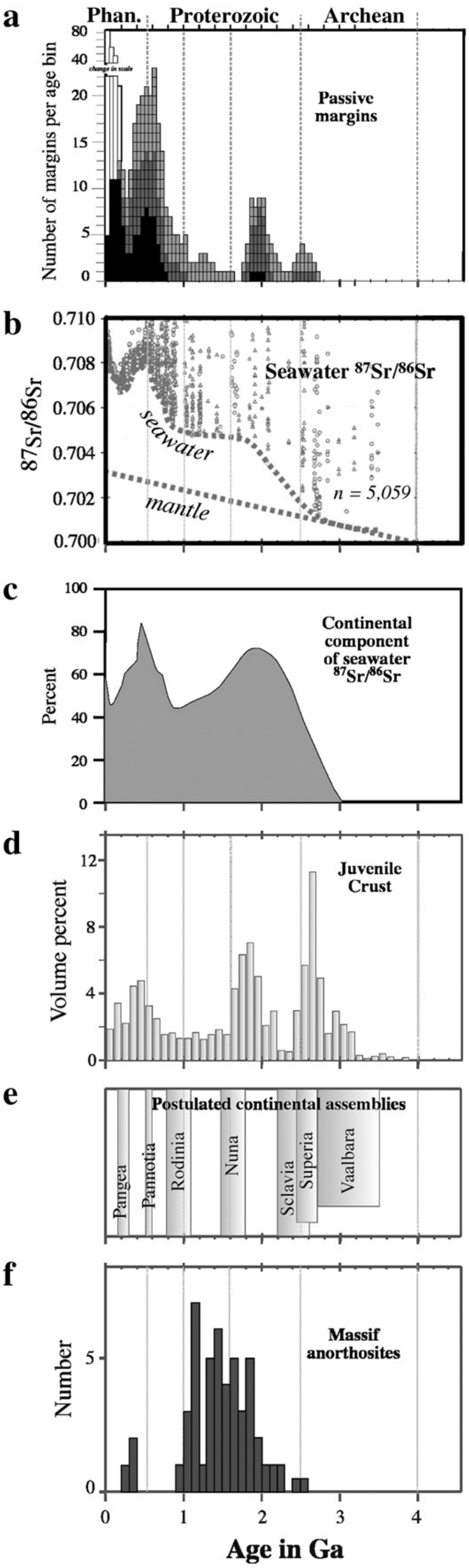
8.2. Pannotia

A reconstruction by Dalziel (1997) at ca. 545 Ma shows a single Pannotia supercontinent featuring East Gondwana,[5] West Gondwana, Laurentia, Baltica, and Siberia. Dalziel (1997) suggested that this supercontinent was distinct from Rodinia (see next paragraph) and was rather short lived (ca. 600 to 540 Ma). The number of passive margins steadily climbed starting at 1000 Ma, peaked at 600–590 Ma (19 margins), declined to a minor low at 555–545 Ma (13 margins), and peaked again at 530–520 Ma (18 margins) (Figs. 5a and 14a). Thus, the time of Pannotia as proposed by Dalziel (1997) includes the time when passive margins were at their most abundant, which clearly does not follow the Pangean analogue. If anything, the passive-margin record would squeeze Pannotia into a very brief interval from 555 to 545 Ma. But even this is unsatisfactory: some Neoproterozoic passive margins were meeting their ends in Gondwana (*e.g.*, northern and western margins of the Kalahari craton, A75 and A76) well after two of the passive margins surrounding Laurentia had already gotten started (margins A3 and A5). Thus, Pannotia was never a supercontinent according to the strict definition.

8.3. Rodinia

A broad consensus supports the existence during the early Neoproterozoic of a supercontinent, which has been called Rodinia by most researchers, or Paleopangaea by Piper (2000). The most convincing evidence for such a supercontinent would be a bombproof reconstruction like that for Pangea. Instead, several very different Rodinia reconstructions have been proposed (*cf.* Hoffman, 1991A; Dalziel 1997; Karlstrom et al., 1999; Piper, 2000), which share the assumption that all the world's continents were gathered into a single supercontinent. Condie (2003) summarized Rodinia's timing as follows: assembly between 1300 and 950 Ma, supercontinent at 950 to 850 Ma, and breakup between 850 and 600 Ma. Similarly, Rogers and Santosh (2003) gave the age of Rodinia as ca. 1100 to 800 Ma. The passive-margin record agrees broadly with the concept of a late Mesoproterozoic Rodinia supercontinent that broke up during the Neoproterozoic: a time of few margins at ca. 1200 to 1050 Ma was followed by an increase from ca. 1050 to 600 Ma. On the other hand, the general scarcity of passive margins during the Mesoproterozoic, when Rodinia was supposedly being assembled, cannot be readily explained by analogy with the Pangea case.

[5] Gondwana has been treated as a supercontinent by some researchers but in fact it only contained about half the world's continental area.



8.4. Nuna

An older supercontinent, referred to as Nuna (Hoffman, 1997) or later as Columbia (Rogers and Santosh, 2002), is proposed to have existed long before Rodinia, during the late Paleoproterozoic. There is ample evidence that a number of preexisting smaller cratons came together during this interval (*e.g.*, Hoffman, 1991A), but the case that almost all the world's continents were gathered into a *single* supercontinent is not compelling. The issue remains to be settled because the various proposed configurations are quite different (*cf.* Hoffman, 1997; Rogers and Santosh, 2003; Zhao et al., 2004). Hoffman (1997) presented a strong case that Laurentia, Greenland, and Baltica had come together by 1800 Ma. According to Rogers and Santosh (2003), what they referred to as Columbia came together ca. 1800 Ma and broke apart ca. 1500 Ma. According to Zhao et al. (2004), this supercontinent came together (in a very different configuration) by a series of collisions from 2100 to 1800 Ma, grew by subduction–accretion until ca. 1300 Ma, and then broke up, to reform soon thereafter as Rodinia. The passive-margin record is broadly consistent with the idea of a late Paleoproterozoic supercontinent, but not with the timing of breakup in either the Rogers and Santosh (2003) or the Zhao et al. (2004) scenarios. An abundance of passive margins between ca. 1850 and 2050 Ma was followed by a precipitous drop between ca. 1850 and 1750 (Fig. 5a). The drop corresponds to the postulated assembly of Nuna. But the passive-margin record summarized in Fig. 5a provides no independent confirmation of the idea that during the Mesoproterozoic, a supercontinent broke up and then reformed into Rodinia.

8.5. Sclavia, Superia, and Vaalbara

These are Archean continental groupings, or supercratons, that are based on a comprehensive global assessment by Bleeker (2003). Sclavia includes Canada's Slave craton and various other cratons that rifted away from it during the Paleoproterozoic; it is proposed to have existed from ca. 2600 to 2200 Ma (Bleeker, 2003). The slightly older Superia includes Canada's Superior craton plus various objects that rifted from it during the Paleoproterozoic; it is proposed to have existed from ca. 2700 to 2450 Ma (Bleeker, 2003). Vaalbara includes the Kapvaal and Pilbara cratons and is propose to have been together from ca. 3470 to ca. 2700 Ma (Bleeker, 2003). None of these meet the strict definition of a supercontinent. The passive-margin record is consistent with all three, subject to minor age adjustments discussed below.

8.6. Proposed scenario

The passive-margin record suggests a somewhat different history of continental aggregation and dispersal than has been previously published. Except for Pangea, none of these groupings can be definitively labeled supercontinents because there is no way to confirm that any of them contained almost all of the world's continental crust in a single entity. Indeed, casual usage of the term "supercontinent" provides circular support for the concept of a supercontinent cycle. Semantics aside, the passive margins clearly confirm that times of continental dispersion have alternated with times of continental aggregation during at least the latter half of Earth history.

Fig. 14. Comparisons between geologic time series: (a) distribution of passive margins from Fig. 5a. (b) Seawater $^{87}Sr/^{86}Sr$, from Veizer and McKenzie (2003). (c) Normalized seawater $^{87}Sr/^{86}Sr$, showing just the continental contribution, from Shields (2007). (d) Juvenile crust, from Condie (2005). (e) Supercontinents, from sources cited in Section 8. (f) Massif anorthosites, adapted from data tabulated by Ashwal (1993). Note the remarkable positive correlation between the passive-margin distribution and the normalized strontium curve, and the negative correlation between the passive-margin and anorthosite distributions.

The patchy record of passive margins in the Archean is consistent with Bleeker's (2003) scenario for three supercratons, subject to minor changes in breakup ages as per Table 2: Vaalbara from ca. 3470 to ca. 2685, Superia from ca. 2700 to ca. 2300, and Sclavia from ca. 2600 to ca. 2090 Ma. Breakup of these supercratons during the first half of the Paleoproterozoic led to a peak in the population of passive margins at ca. 1900 Ma. One by one, each of these margins collided with something, leaving no passive margins at all from ca. 1740 to ca. 1600 Ma. The passive-margin record is consistent with the purported coalescence of Nuna by this time, with Laurentia at its core—but whether or not Nuna met the strict definition of a supercontinent, or was merely a supercraton, remains to be demonstrated. This allows the possibility that Nuna and at least one other supercraton formed around the end of the Paleoproterozoic.

The passive-margin record provides little support for previous hypotheses of wholesale breakup of a single supercontinent during the Mesoproterozoic, whose pieces came back together at ca. 1000 Ma in a new configuration called Rodinia. Instead, I suggest that Nuna grew by lateral accretion of juvenile arcs during the Mesoproterozoic (*e.g.*, Karlstrom et al., 2001), and that two or more supercratons came together during a series of Grenvillian collisions (ca. 1190–980 Ma; Rivers, 1997). This grouping equates to the supercontinent Rodinia of previous workers—but whether or not *it* was truly a single entity remains debatable, because the fit of the Rodinia continents remains equivocal.

Rodinia began to breakup even as it was supposedly forming: the first Uralian, Timanide, and Baikal margins (A24, A23, and A40) all formed at ca. 1000 Ma, while in eastern Laurentia, Grenvillian collision was nearing its end. Many breakups followed during the early to mid-Neoproterozoic, and the population of passive margins reached its maximum in the ancient record at 610–590 Ma. As discussed above, the hypothesized supercontinent Pannotia seems to have been an ephemeral grouping of some but not all of the continents. A number of passive margins began before and ended after the putative 600–540 Ma tenure of Pannotia (Fig. 11b). Almost all of the continents finally had come together by about 300 Ma to form Pangea, which broke up starting 180 Ma. Even Pangea, the archetypal supercontinent, did not quite contain all the world's continents (*e.g.*, Arctic Alaska microcontinent).

8.7. Implications for continental reconstructions

With a few straightforward exceptions,[6] age assignments for the passive margins in the present study were independent of continental reconstructions in the published literature. The data summarized in Table 2 thus can be used to evaluate and refine continental reconstructions. Conjugate margins that have come to be separated by the opening of an Atlantic-type ocean should come in matched pairs, like the present-day margins of eastern North America and western Africa. Where is the passive margin that matched the Innuitian margin of Arctic North America (margin A5)? The missing conjugate margin should have a start date of about 620 Ma, as well as basement geology consistent with having broken away from the Canadian Arctic. The matching margin might flank a craton, but the possibility cannot be ignored that it formed along a ribbon microcontinent—the sort of object that is not as widely recognized or well publicized, but just as important for this kind of research.

[6] The start date of the Jurassic margin of Venezuela (margin A60) was borrowed from the start date of its supposedly conjugate margin, in Cuba. Likewise, the start date for the northern margin of the Kalhari craton (A75) was borrowed from the supposedly conjugate southern margin of the Congo craton (A74). The start- and end dates for the now-dismembered Neoproterozoic to Paleozoic East Greenland-Svalbard margin (margin A20a and b) were inferred by combining evidence from both. The start- and end dates for the now-dismembered Cambrian to Pennsylvanian Sierra de la Ventana-Cape-Ellsworth margin (margin A66a, b, and c) were inferred by combining evidence from each.

9. Comparisons with other aspects of Earth history

9.1. Isotopic composition of seawater strontium

The age distribution of passive margins shows a striking correlation with fluctuations in the isotopic composition of $^{87}Sr/^{86}Sr$ in seawater (*cf.* Fig. 14a, b, and c). Overall, the global $^{87}Sr/^{86}Sr$ ratio has increased through geologic time owing to the inexorable decay of the world's initial allotment of ^{87}Rb to ^{87}Sr. Fluctuations in the $^{87}Sr/^{86}Sr$ isotopic ratio of seawater track the shifting balance between sources of primitive mantle strontium (low $^{87}Sr/^{86}Sr$), which mostly enters seawater via hydrothermal circulation at mid-ocean ridges, and evolved strontium (high $^{87}Sr/^{86}Sr$), which mostly enters the sea at continental margins (see review by Veizer and McKenzie, 2003, and references therein). The curve in Fig. 14b shows the raw data (from Veizer and McKenzie, 2003); the curve in Fig. 14c (from Shields, 2007) is normalized so as to show just the continental contribution to seawater Sr. The highs at 1900–1890, 610–520, and 150–0 Ma in the passive-margin distribution have close counterparts in the normalized strontium curve. The lows at >2750 and 300–275 Ma also are reflected in the strontium curve, but slightly offset. Between the two curves, there is only one significant mismatch: both show broad mid-Proterozoic lows, but they are skewed, with the minimum being centered at about 1740–1600 Ma for passive margins, compared to 1000–900 Ma for strontium. This discrepancy may indeed be real, but it should be noted that all but one of the passive margins that formed between 1700 and 1000 Ma have quality rankings of C, and moreover, that the strontium data are sparse (Fig. 14b). Regardless of this one discrepancy, the overall similarity between the two time series is striking—especially when one remembers how utterly independent the datasets are: the strontium curve is based on thousands of isotopic analyses of calcium carbonate, whereas the passive-margin curve is a compilation of tectonic interpretations of 76 orogenic belts. The broad form of the passive-margin age distribution appears to be correct and cannot be written off to poor preservation or an inadequate compilation.

Why should such a correlation exist? It would appear that classic Wilson Cycles involving the formation and destruction of passive margins are an effective way to involve large volumes of old continental basement in the sedimentary cycle (Richter et al., 1992). In contrast, plate interactions involving only isotopically primitive rocks (*e.g.*, oceanic crust, intraoceanic arcs, and recently formed continental crust) would contribute scant evolved strontium to the sea, no matter the intensity of orogeny, the extent of synorogenic exhumation, or the tempo of plate tectonics. A contributing factor may be the size of continents (Halverson et al., 2007): a single supercontinent with an arid interior would probably transmit less radiogenic strontium to the sea than would an equal area of smaller, dispersed continents having a greater perimeter. Significantly, neither the rift-drift-transition histogram (Fig. 5b) nor the arc–passive margin collision histogram (Fig. 5c) is as good a match for the normalized strontium curve (Fig. 14c) as the overall passive-margin histogram (Figs. 5a and 14a).

9.2. Juvenile crust

The age distribution of juvenile continental crust shows prominent maxima at ca. 2700–2600, 1900–1800, and 500–300 Ma, and minima before 3200 Ma, at 2400–2200 Ma, at 1700–700 Ma, and at 300–200 Ma (Fig. 14d) (Condie, 2005). Comparison with the passive-margin record (Fig. 14a) shows a fairly good correspondence at first order, except for the last hundred million years.

9.3. Massif anorthosites

The age distributions of passive margins and massif anorthosites (*cf.* Fig. 14a and f) show an extraordinary negative correlation. The massif anorthosite distribution, which was constructed from a tabulation by Ashwal (1993), shows a broad peak between 2.0 and 1.0 Ga that coincides with the broad Mesoproterozoic low in the passive-margin distribution (the “boring billion” of Holland, 2006). Hoffman (1989) suggested that the anorthosite pulse was a consequence of thermal blanketing of the mantle by a large, stationary Laurentian continent (which would now be referred to as Nuna, or Columbia). The passive-margin record is consistent with this idea.

10. Less common fates of passive margins

As previously noted, the ancient passive margins in this synthesis had one of three fates: (1) collision, (2) re-rifting, and (3) conversion to a convergent margin. Here I elaborate on the last two.

10.1. Re-rifting

Re-rifting (Sengör, 2004) involves the departure of a ribbon continent or microcontinent from an existing passive margin. For this to happen, a new spreading ridge must break through near the ocean–continent boundary. Re-rifting is a more common process than is generally appreciated. The best young example is the Barents Shelf of the Arctic, which already was a passive margin in the Cretaceous when a new spreading center developed near the old ocean–continent boundary, leading to the separation of the Lomonosov ribbon microcontinent (*e.g.*, Kristoffersen, 1990). Ribbon microcontinents are believed to have rifted from the eastern Laurentian passive margin in the Neoproterozoic to Cambrian (margin A19a; Waldron and van Staal, 2001), and from the eastern passive margin of Siberia in the Devonian (margin A57; Parfenov, 1991; Sengör and Natalin, 1996; p. 554). The Tethyan realm abounds with passive margins that appear to have formed by the breakaway of ribbon microcontinents, which left the Gondwana margin on the south, and drifted north to collide with Eurasia (Sengör et al., 1988; Stampfli et al., 1991).

10.2. Conversion to a convergent margin

Direct conversion of a passive to a convergent margin (Fig. 1b) is rare indeed, as Burke et al. (1984) first showed and the present work confirms. The notion that this is commonplace dates at least as far back as Dewey and Bird (1970) in their pioneering paper that tied plate tectonics to orogenic geology. The present compilation shows that a passive margin can attain great age *without* converting directly to an Andean-type margin; eleven of them lasted at least 300 m.y.

That it is *possible* for a passive margin to convert directly to a convergent margin is demonstrated, however, by the history of the northern margin of Iberia, facing the Bay of Biscay (margin A29). It formed in the Early Cretaceous (ca. 115 Ma) when Iberia broke away from Europe. In the Late Cretaceous (late Senonian, ca. 70 Ma), convergence across the ocean–continent boundary began, and an accretionary wedge formed (Vergés and García-Senz, 2001). Significantly, a full-fledged, enduring subduction zone never did form, as there is no magmatic arc and the margin eventually lapsed into inactivity.[7] The Iberian passive margin thus lasted a mere 45 m.y., making it one of the shortest-lived ones in the entire dataset. When this passive margin failed in compression, it was not because it had grown old.

The first proposed instance of conversion from passive margin to convergent margin was the Appalachian margin during the Ordovician (Bird and Dewey, 1970) (margin A19a). This scenario for the Taconic orogeny was quickly abandoned after Stevens (1970) advanced the more compelling arc–passive margin collision interpretation that is described in Section 4.2. The northern part of the Canadian Cordilleran passive margin (margin A3) is interpreted here as having collided with an arc during the Devonian, but an alternative (implied, for example, by Nelson et al., 2002) is that it converted to a convergent margin without first colliding with something. Direct conversion from passive- to convergent margin has been proposed for three other margins that weren't included in the present synthesis, for want of detailed information. The early Paleozoic Gondwanan margin of Chile is proposed to have converted directly from a passive margin to a convergent one during the Ordovician (Bahlburg and Hervé, 1997). The North Afghan platform has been interpreted as the site of an Ordovician? to Early Devonian passive margin that developed into an arc in Late Devonian and Mississippian time, apparently without a collision in between (Brookfield and Hashmat, 2001). A similar history has been proposed for the southern margin of the Iranian plate during the Late Triassic (Sheikholeslami et al., 2008). A geologic model for this process, based on well-documented case studies, would be a valuable contribution.

11. Summary

1. Passive margins have existed somewhere on Earth almost (but not quite) continually since the Neoarchean.
2. The oldest postulated, although controversial, passive margin in the compilation is Steep Rock Lake (Superior Province), ca. 3000–2800 Ma. Modern-style passive margins appear to be absent from the rock record before that time, either because they never existed, have been deformed beyond recognition, have been eroded away, or now rest in the deep crust or mantle.
3. Passive margins are unevenly distributed through the latter half of earth history, with peak abundances at ca. 1900–1890, 610–520, and 150–0 Ma and low abundances at ca. 1740–1000 and 300–275 Ma.
4. The distribution of passive margins through time correlates with parts of the proposed supercontinent cycle, but not with others. Good correlations are seen with the assembly of Nuna, the breakup of Rodinia, and the assembly and breakup of Pangea. The passive-margin record is not obviously consistent with the proposed breakup of Nuna, the assembly of Rodinia, or the assembly or breakup of the putative Pannotia. An alternative scheme is proposed in which Rodinia formed by collision of at least two supercratons, which each had existed through the Mesoproterozoic.
5. The age distribution of passive margins appears to be robust as it shows a remarkable match for the continental component of the seawater $^{87}Sr/^{86}Sr$ secular curve.
6. The still unfinished lifespans were determined for all of the modern-day passive margins. These have an aggregate length of 105,000 km, a mean age of 104 m.y., and a maximum age of 180 m.y.
7. Seventy-six ancient margins have a mean lifespan of 178 m.y. and a range of lifespans from 25 to 590 m.y.
8. The mean lifespan of 47 Precambrian margins was 206 m.y., compared to 15 Paleozoic margins that had a mean lifespan of only 137 m.y. The five longest-lived passive margins, all of them Mesoproterozoic, had lifespans exceeding 350 m.y. It is hard to avoid the conclusion that passive margins lasted longer during the Precambrian. This unexpected finding conflicts with the notion espoused by earlier workers that the tempo of plate tectonics was faster in the Precambrian than at present, because the plates were smaller, moved faster, or both. The longevity of Precambrian

[7] The northern Iberian margin thus defies pigeonholing into either a “modern” or an “ancient” passive margin: it originated as a true passive margin, briefly became a convergent margin, and has since been inactive, though not “passive” according to my definition. Thus it is lumped with the “ancient” margins in this paper.

passive margins is consistent instead with Korenaga's (2006) recent modeling of mantle evolution, which suggested that plate tectonics was more sluggish in the Precambrian.

9. Passive-margin collisions have produced–or at least preserved–high-pressure, low-temperature metamorphic conditions only since ca. 625 Ma.
10. Additional research is warranted for all margins, but a few are particularly important. Steep Rock Lake (margin A10) and the Belingue greenstone belt (margin A78) may be the world's oldest passive margins, but tectonic interpretations are controversial and age control is inadequate. The longevity of the Mesoproterozoic passive margins of eastern, western, and southern Siberia (margins A38, A56, and A40) and eastern and northern Baltica (margins A24 and A23) needs to be carefully assessed.

Acknowledgments

This study was preceded by research in the early 1980s by Kevin Burke, Bill Kidd, and Lauren Bradley into the question of whether or not passive margins convert spontaneously to convergent margins. I concur with them: not often. I especially wish to thank Paul Hoffman for sharing his insights on many of the Proterozoic margins, and David Rowley both for ideas about many of the Phanerozoic margins, and for a set of preliminary age picks for the modern passive margins. I have been helped with various age picks and (or) interpretations by Tanya Atwater, Wouter Bleeker, Kevin Burke, Kevin Chamberlain, Bill Collins, John Dewey, Maarten de Wit, Yildirim Dilek, David Evans, Karl Karlstrom, Tim Kusky, Xiang-Zing Li, Jim Pindell, Sergei Pisarevski, Ali Polat, Celal Sengör, and Fred Ziegler. Graham Shields shared a pre-publication plot of the continental component of seawater strontium through time. Many of the start- and end dates were based on geochronology on key units that were sought out by Sam Bowring and his students. Keith Labay provided GIS support. Figs. 8b and 9b were modified from cross-sections by Peter Cawood and Alexander Khudoley. Reviews by Tim Kusky, Rich Goldfarb, Kent Condie, and Paul Hoffman substantially improved the manuscript. Suggestions and information on additional margins will be gladly received.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.earscirev.2008.08.001.

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Supplementary Data for:

Passive margins through earth history[1]

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Appendix A. Explanatory notes for ages of modern passive margins in Table 1. References cited herein are provided under the heading "Further Reading."

M3 and M4. Lena East and Lena West refer to the Arctic margin of Siberia near the Lena River where the Arctic mid-ocean ridge comes onshore. Sources: Kovacs et al. (1985); Grantz et al. (1989). The oldest anomaly on either margin is ca. 52 Ma. The mean age would then be ca. 26 Ma since the seafloor at the ridge itself formed at 0 Ma.

M5. For the north coast of Alaska, the age is assumed to be that of the north coast of Canada, following the widely accepted "windshield wiper" model (e.g., Grantz et al., 1998).

M16. The magnetic anomalies that intersect the south side of the Grand Banks range in age from 170 to 129 Ma. Source: Klitgord and Schouten (1986, their Fig. 1).

M18. The magnetic anomalies that intersect the north side of the Bahama platform range in age from ca. 170 to ca. 146 Ma. Source: Klitgord and Schouten (1986, their Fig. 1).

M51. The magnetic anomalies that intersect the east coast of the Arabian Peninsula range in age from ca. 75 to ca. 33 Ma. Source: Commission de la Carte Géologique du Monde (2000).

M59. Seafloor spreading began about 28 Ma. Source: Tamaki et al. (1992).

M60. Oldest anomaly is C11, giving an age of 31 Ma. Source: Briais et al. (1993).

M61. Oldest anomalies are C7 in the northeast and C6 in the southwest (about 27 and 19 Ma, respectively), giving a mean age of 23 Ma. Source: Briais et al. (1993).

M65. Naturaliste Plateau (west of Perth), north side. Anomalies are identified as M1 to M5 giving a mean of ca. 127 Ma. Source: Pyle et al. (1995, their Fig. 1).

M66. Naturaliste Plateau (west of Perth), west side. Assuming that the easternmost anomaly is M5, the age of this sector is ca. 130 Ma. Source: Pyle et al. (1995, their Fig. 1).

M71. The southern margin of Chatham Rise—which is in two halves, split by edge of map—is ca. 80 Ma. Source: Commission de la Carte Géologique du Monde (2000).

M72. The Antarctic passive margin is assigned breakup ages based on the better documented ages of the matching passive margins to the north. The Pacific sector matches M70 and the eastern part of M71. Its mean age is approximate.

M73. Using the same rationale, the age of this sector is the mean, weighted by length, of M38, M39, M44, and M57. Minor margins of Sri Lanka are ignored for this calculation.

M74. Using the same rationale, the age of this sector is taken as the age of M66.

M75. Using the same rationale, the age of this sector is taken as the age of M67.

Appendix B. Geologic histories of ancient passive margins.

This section provides documentation for the passive margins listed in Table 2. For each margin, the main goals were: (1) to assess the case for a passive-margin interpretation citing appropriate literature; (2) to determine the start date; (3) to determine the end date; (4) to document the fate of the margin; and (5) to highlight issues needing further attention. Margin numbers, with the prefix "A" for ancient, are ordered from northwest to southeast in Figure 6. In cases where a single passive margin clearly has been broken into two or three parts by later seafloor spreading, the margin is given a single number modified by a lower-case letter, as margins 19a and 19b for the Appalachians and Scotland, respectively. Two successive passive margins (*i.e.,* where a collision or re-rifting event interrupted) in the same approximate location are shown by a single line in Figure 6 but get two different numbers. All compass directions are in the present-day reference frame. Lengths of passive margins correspond to great-circle distances between the two ends of a margin as determined using ARC-GIS on the Carte Géologique du Monde (Commission de la Carte Géologique du Monde, 2000).

A1. Brookian margin of Arctic Alaska terrane

The Brooks Range orogen formed along the site of a passive margin of the Arctic Alaska microcontinent, which subsided from late Paleozoic to Jurassic. A

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Devonian to earliest Carboniferous siliciclastic succession (Endicott Group) is now widely regarded as a product of rifting (Moore et al., 1994), although an earlier interpretation treated it as a foreland-basin succession akin to coeval strata in the Canadian Arctic (Nilsen, 1981). It is overlain by platform carbonates of the Lisburne Group, as old as the Kinderhookian of the earliest Carboniferous (Dumoulin et al., 2004). This corresponds to the latter half of the Tournaisian, ca. 350 Ma. As Arctic Alaska drifted out of the tropics, the Mesozoic part of the passive margin sequence was dominated by cherts and siliciclastic rocks rather than carbonates. Brookian orogenesis, due to collision of an arc that approached from the south (present direction), is marked by an influx of flysch from southerly sources (Okpikruak Formation). Based on stratigraphic grounds, collision began at least as early as Berriasian (earliest Cretaceous) (Moore et al., 1994, p. 121). However, $^{40}Ar/^{39}Ar$ data show that continental margin rocks were already being metamorphosed to blueschist conditions as early as 170 Ma (Christiansen and Snee, 1994), so this is taken as the end date of the passive margin. Thus, the lifespan was about 180 m.y.

A2. Farewell terrane, Alaska

The Farewell terrane of interior Alaska is a dismembered microcontinent that includes a Paleozoic carbonate platform (Nixon Fork subterrane of Bundtzen et al., 1997) and a coeval off-shelf succession to the east (Dillinger subterrane of Bundtzen et al., 1997). The margin flanked and was built on a Proterozoic basement block containing 2040 to 2085 Ma granitoids, 980-Ma rhyolites, and 850-Ma orthogneisses (Bradley et al., 2007). During most of the Ordovician (ca. 488 to ca. 450 Ma), the Nixon Fork was quite clearly a passive margin. In the northern Kuskokwim Mountains, platform carbonate deposits of this age range are >3km thick and show a classic exponentially declining subsidence rate consistent with a passive margin setting (Dumoulin et al., 1998). Whereas the existence of a passive margin seems clear, the details of its initiation and demise both are problematic.

The early history must be pieced together using information from the Lone Mountain area, 100 km to the south. Here, an Ordovician platformal succession appears to be broadly equivalent to that just described, though poorly exposed. It is seen to overlie a late Neoproterozoic? to Cambrian succession of quartzites and carbonates 600+ meters thick (Babcock et al., 1994). All of these strata are consistent with a passive margin setting; older rift deposits have yet to be identified as such. The seven youngest detrital zircons from the Windy Fork and Lone Mountain Formations, near the base and midway through this section, have a mean age of 537 Ma (Bradley et al., 2008 and author's unpublished SHRIMP data, 2008); the base of the section and the rift-drift transition are therefore likely to be late Neoproterozoic, ca. 545 Ma.

The demise of the margin is also a problem. The platform was dominated by carbonate deposition in the Cambrian and Ordovician. Deep-water carbonates and shales signal a prolonged Silurian drowning event, which finally ended in the Devonian when platformal conditions were re-established. Meanwhile, in the deep-water Dillinger subterrane, the Silurian was marked by an influx of siliciclastic turbidites with interbedded tuffs (Terra Cotta Mountains Sandstone) and detrital zircons that are foreign to the Farewell terrane (author's unpublished data, 2008). This has the earmarks of an orogenically derived flysch succession, and I interpret it to represent the foredeep of an arc-passive margin collision zone. Accordingly, I place the demise of the margin at ca. 435 Ma, just older than the 433±2 U-Pb zircon age of a newly dated tuff interbedded with turbidites (author's unpublished SHRIMP data, 2007). Start- and end dates for the margin of 545 and 435 Ma yield a nominal lifespan of 110 m.y.; the quality ranking is C.

A3 and A4. Cordilleran margin of western Laurentia, northern and southern sectors

The Cordilleran margin of western Laurentia formed by rifting in the Neoproterozoic, and, according to the most compelling model, collided with an arc during the mid-Paleozoic Antler orogeny. The age of the rift-drift transition and the tectonic interpretation of the end of passive margin conditions have both been subject to much debate. Accordingly, I discuss two sectors of the margin separately: northern Canada, and southern Canada-western United States.

Passive margin evolution along the southern sector was assessed by Bond and Kominz (1984), who used tectonic subsidence analysis to pick an age for the rift-drift transition at about the Neoproterozoic-Cambrian boundary (542 Ma according to the time scale of Gradstein and Ogg, 2004, but ca. 575 Ma according to the time scale used by Bond and Kominz, 1984). The age of breakup is best constrained in the southern Canadian Rockies, where strata of the Windermere Supergroup and overlying Hamill Group are interpreted as rift deposits. Basalt near the base of the Windermere Supergroup is ca. 736 Ma and basalt at the top of the Hamill Group is 570±5 Ma (U-Pb zircon) (Colpron et al., 2002), implying a lengthy rifting episode. Rocks of the Hamill Group are unconformably overlain by the Gog Quartzite, which is the lowest unit of a seaward-thickening prism of Cambrian and younger carbonates, and which is interpreted as the passive margin succession. The base of the Gog Quartzite is about at the Neoproterozoic-Cambrian boundary (Colpron et al., 2002). This accords with Bond and Kominz's (1984) placement of the rift-drift transition, albeit subject to recalibration of the time scale. Thermal subsidence of the miogeocline continued until Devonian time.

The southern sector of the margin collided with an arc during the Devonian Antler orogeny. The arc-collision model was first developed for the Antler orogeny in Nevada, where deep-water continental margin strata were thrust eastward onto the lower Paleozoic platform. Convergence began offshore in latest Devonian and platform drowning is dated as mid-Kinderhook (Early Mississippian: earliest Tournaisian) (Johnson and Pendergast, 1981). For the southern sector of the

Cordilleran margin, I place the end date at 357 Ma, giving a lifespan of 185 m.y.

Passive margin evolution along the northern sector began much earlier than in the south (Colpron et al., 2002). In the Mackenzie Mountains, the ~6-km-thick, Neoproterozoic Windermere Supergroup is interpreted to include both rift and passive-margin deposits. Strata of the Windermere Supergroup rest on those of an older Neoproterozoic carbonate platform (Mackenzie Mountains Supergroup) that has been interpreted as an intracratonic basin (Batten et al., 2004). Rocks of the Coates Lake and Rapitan Groups, which comprise the lower one-third of the Windermere Supergroup, are interpreted as rift-related. The upper two-thirds of the Windermere Supergroup consists of kilometer-scale siliciclastic-to-carbonate grand cycles, which are interpreted as a passive margin succession (Pyle et al., 2004). The rift-drift transition is probably not much younger than syn-rift volcanic rocks in northern British Columbia that yielded a U-Pb zircon age of 689±5 Ma (see Colpron et al., 2002 for original sources). I place the rift-drift transition at 685 Ma. The passive margin succession comprises the upper two-thirds of the Windermere Supergroup, plus a Cambrian to Middle Devonian carbonate-dominated succession as reviewed by Fritz et al. (1991).

The collision model, originally developed for the southern sector, was later proposed for the northern sector as well (Smith et al., 1993; but see Nelson et al. 2002 for a different view). In the northern Canadian Rockies and adjacent Alaska, outboard-derived, Devonian siliciclastic rocks of the Imperial, Tuttle, and Nation River Formations, and the Earn Group represent the inferred foreland basin (Smith et al., 1993). Strata of the Imperial Formation in northern Yukon overlie platform carbonates and consists of shale and chert with local carbonate buildups, then deep-marine shales, and finally, turbidites; the overlying Tuttle Formation is conglomeratic and includes fluvial and deltaic facies (Gordey et al., 1991). The succession of platform drowning-flysch-molasse is entirely consistent with a foreland basin model and not readily explained otherwise. The base of the Imperial Formation is Givetian in age; accordingly, I place the passive margin to foreland-basin transition at 387 Ma. The corresponding lifespan for the northern sector is 298 m.y., rounded to 300 m.y.

A5. Northern margin of Laurentia, Innuitian orogen

An early Paleozoic passive margin can be traced across the Canadian Arctic and North Greenland. Comprehensive reviews were published by Trettin et al. (1991) and Higgins et al. (1991) for the Canadian and Greenland sectors, respectively. The passive margin included a carbonate-dominated, early Paleozoic continental terrane (Franklinian platform), and, flanking it to the north, a coeval deep-water slope-rise succession. The oldest rocks of the passive-margin package are not exposed, but are seen seismic profiles across Melville Island, where several kilometers of presumably Neoproterozoic strata have been imaged (Harrison, 1995). Recent work in Ellesmere Island has suggested that the oldest exposed platformal unit, the mixed carbonate and siliciclastic Kennedy Channel Formation, is late Neoproterozoic in age (Dewing et al., 2004). The long-distance correlations proposed by Dewing et al. (2004) suggest that the Kennedy Channel Formation is not much younger than the Marinoan Glaciation, ca. 635. In North Greenland, the oldest known deep-water strata are near the Neoproterozoic-Cambrian boundary (Surlyk and Hurst; 1984) and show that the passive margin was well established by ca. 542 Ma. For this synthesis I place the rift-drift transition at ca. 620.

The passive margin became a foreland basin with collision of the Pearya arc (Trettin et al., 1991). The earliest stratigraphic record of collision is a latest Ordovician to earliest Silurian (ca. 444 Ma) influx of orogen-derived turbidites that buried slope and rise deposits in northernmost Greenland and Ellesmere Island (Surlyk and Hurst, 1984). These strata represent the initial, underfilled stage of the Ellesmerian foreland basin, which would continue to receive orogenic sediments for 80 m.y. From Silurian to mid-Devonian, the flysch basin advanced to the south, causing the platform to retreat; in Greenland, collapse of the shelf was accompanied by normal faulting (Surlyk and Hurst, 1984), which Bradley and Kidd (1991) attributed to flexural extension. Synsedimentary backstepping of the foreland basin shows that the influx of flysch was related to plate convergence and bending directly to the north (and thus marks the demise of the passive margin). This argues against an otherwise viable alternative—that the flysch spread across the Innuitian continental apron from the distant Caledonian collision along strike to the east—which would mean that its arrival would not bear on the end date of the passive margin.

The suggested start- and end dates of ca. 620 and ca. 444 Ma yield a duration of 176 m.y., rounded to 180 m.y. to allow for the guesswork involved. Despite a well-constrained end date, the start date is so approximate that the quality ranking is only C.

A6. Wopmay orogen and the Coronation margin of Slave craton, Canada

The Wopmay orogen in northwestern Canada was one of the first Precambrian mountain belts to have been interpreted in terms of arc-passive margin collision (*e.g.*, Hoffman, 1980; Hoffman and Bowring, 1984). The Coronation passive margin formed during the Paleoproterozoic on the western side of the Archean Slave craton (Hoffman et al., 1970; Hoffman, 1973). The age of rifting is constrained by a U-Pb zircon age of 2019 Ma from a rhyolite at the top of the mainly mafic Valiant Formation (S. Bowring, quoted by P. Hoffman, written communication 2008). The oldest passive-margin clastics (Odjick Formation) directly overlie the Valiant Formation, so the rift-drift transition cannot be much younger than the dated rhyolites; I place it at ca. 2015 Ma. Volcanic rocks dated at ca. 1890 Ma were previously ascribed to rifting (Hoffman and Bowring, 1984) but are now attributed to intraarc extension at the time of collision (P. Hoffman, written communication, 2008). The passive margin itself is represented by the Epworth Group of mainly shallow-water carbonates (Hoffman and Bowring, 1984). The demise of the Coronation margin is marked by drowning of the platform, mafic magmatism (Hoffman, 1987), and influx of orogenically derived turbidites (Recluse Group), which represent a foreland basin related to collision of the Hottah arc terrane. A tuff near the base of the Recluse Group yielded a U-Pb zircon age of 1882 Ma; I place the demise of the margin just before this, at ca. 1883 Ma. These ages suggest a lifespan of 132 m.y., which is remarkably close to the lifespan of 125 m.y. originally estimated by Hoffman et al. (1970), long before modern geochronological controls could be brought to bear.

A7. Thelon orogen and Kimerot platform, Canada

The Thelon orogen is the approximately coeval mirror image of the Wopmay orogen, on the east side of the Slave craton in northern Canada. The Paleoproterozoic rifted margin is represented by the Kimerot Group, which includes a lower siliciclastic unit and an upper carbonate unit, each as much as 250 m thick (*e.g.*, Tirrul and Grotzinger, 1990). The age of the rift-drift transition is inferred by extrapolation from the Great Slave Lake region, where breakup on the southeast margin of the Slave craton can be estimated at ca. 2090 Ma (P. Hoffman, written communication, 2008). This is based on U-Pb zircon ages from 2185±7 to 2094±10 Ma on the alkalic to peralkalic Blatchford complex (Hoffman et al., 1984). Drowning of the Kimerot platform beneath siliciclastics rocks of the Kilogehok basin has been interpreted as the result of arc-passive margin collision. The oldest dated ash bed in the foreland-basin sequence is 1969±1 Ma (Bowring and Grotzinger, 1992), and I place the age of initiation of the foredeep slightly earlier, at 1970 Ma. Thus the Kimerot platform, and by inference the passive margin, had a lifespan of about 120 m.y.

A8. Borden Basin, Canada

The Borden Basin of Baffin Island is one of a series of Mesoproterozoic basins along Laurentia's northern margin. I here follow Hoffman's (1989) interpretation that the basin is not simply a rift, but rather a feature that evolved through rift, passive-margin, and foreland basin phases. Basement is the Archean to Paleoproterozoic Rae craton. The strata of interest comprise the Bylot Supergroup, which consists of three groups from base to top: Eqalulik Group (mafic volcanic and siliciclastic rocks), Uluksan Group (carbonate rocks), and Nunatsiaq Group (siliciclastic rocks) (Sherman et al. 2002). Rifting is dated at 1267 Ma based on a U-Pb baddeleyite age from igneous rocks correlated with basalt at the base of the Eqalulik Group (LeCheminant and Heaman, 1989). The Uluksan Group is interpreted as a passive margin platform; samples from this sequence yielded a somewhat imprecise Pb/Pb calcite age of ca. 1204±22 Ma (Kah, unpublished, cited in Sherman et al., 2002). An unconformity with pinnacle reefs near the top of the Uluksan Group may be related to a forebulge—the first distal effect of collision. The Nunatsiaq Group records drowning of the platform and submarine fan sedimentation. A collisional orogen would have lain to the west of present exposures, obscured beneath Phanerozoic cover; the hypothesized ocean that closed has been referred to as the Poseidon Ocean (Jackson and Ianelli, 1981). I place the start date for the passive margin at 1255 Ma and the end date at 1200 Ma, giving a duration of about 55 m.y. Detrital zircon data from the Uluksan Group are needed to test Hoffman's (1989) model for the evolution of the Borden Basin; non-Laurentian zircons would be key evidence.

A9. Hearn craton, southeast side, Canada

The Wollaston Supergroup on the southeast side of the Hearn craton represents a metamorphosed Paleoproterozoic passive margin that was destroyed during the Trans-Hudson orogeny. The following discussion is based on Yeo and Delaney (2006). The Courtney Lake Group, at the base of the supracrustal succession, is composed of arkose, conglomerate, quartzite, minor pelite, and bimodal volcanic rocks. A rhyolite porphyry yielded a U-Pb zircon TIMS age of 2075±2 Ma, which is similar to the age of the youngest detrital zircons from associated arkoses. A rift setting is inferred for the Courtney Lake Group. The overlying Souter Lake Group consists mostly of mature quartzites. It is seen as representing the drift stage of a passive margin that faced the Manikewan Ocean to the southeast. The rift-drift transition is not very tightly bracketed but for present purposes I place it at ca. 2070 Ma. A classic foreland-basin succession began with an unconformity attributed to a forebulge on the thrust-loaded passive margin plate. This was followed by a succession of carbonates (Karin Lake Formation), then graphitic metapelite (George Lake Formation), and finally an upward-coarsening turbidite succession (Bole Bay, Thomson Bay, and Roper Bay Formations). Detrital zircons as young as ca. 1880 Ma date the passive margin to foreland-basin transition. The ensuing Trans-Hudson Orogeny (ca. 1860-1780 in this region) involved four episodes of ductile deformation and high-grade metamorphism. The passive margin had a duration of about 190 m.y.

A10. Steep Rock Lake platform, Wabigoon Province, Superior craton, Canada

Neither the age constraints nor the tectonic interpretation of this possible passive margin are sufficiently robust for the main purpose of this study. Nonetheless, Steep Rock Lake in Canada's Superior craton is the oldest purported passive margin in the compilation. Basement is an Archean tonalite—the Marmion batholith—dated at 3002 Ma (U-Pb zircon; Tomlinson et al., 2003). It is unconformably overlain by a succession of 0 to 150 m of conglomerate, sandstone, and pelite (Wagita Formation), which in turn is overlain by 0 to 500 m of stromatolite-bearing carbonates (Mosher Carbonate) (Wilks and Nisbet, 1988). The irregular upper contact of the carbonates is interpreted as a karst surface, which is overlain by the 100 to 400-m-thick Joliffe Ore Zone of lateritic iron formation (Wilks and Nisbet, 1988). To this point, there is no disagreement that the succession is stratigraphic, and this is the part that, according to one interpretation discussed below, would represent passive-margin deposition. The iron ores are overlain either stratigraphically (Wilks and Nisbet, 1988) or structurally (Kusky and Hudleston, 1999) by the Dismal Ashrock, a mixture of ductilely deformed mafic and ultramafic volcanic and sedimentary rocks. The Dismal Ashrock has some age control: Tomlinson et al. (2003) reported two populations of zircons from a Dismal Ashrock lithology described as komatiitic lapilli tuff that are 2999-2989 Ma and 2780 Ma. Because komatiite is not normally a zircon-bearing igneous rock type, these zircons must ultimately have other origins (*i.e.,* inherited, detrital, or tectonically mixed, from the Marmion batholith for the older population and possibly from a 2780-Ma tonalite 80 km to the east for the younger population). The structurally highest unit at Steep Rock Lake is the Witch Bay Volcanics, 5 km of highly deformed mafic and minor felsic volcanic rocks. These rocks are undated but their age is established at ca. 2931 Ma by correlation with similar volcanic rocks of the Finlayson Lake greenstone belt to the west (Tomlinson et al., 2003).

In the passive margin interpretation, the Dismal Ashrock represents a foredeep sequence (Hoffman, 1991b) or tectonic melange (Kusky and Hudleston, 1999) that formed during emplacement of an allochthonous Witch Bay arc over a Steep Rock Lake passive margin. The conglomerates and sandstones would then represent rift deposits and the carbonate platform a thermally subsiding passive margin. The unconformity and deep weathering at the top of the carbonates might record a forebulge at the onset of collision. The Dismal Ashrock would represent a subduction-accretion complex. Blocks and lenses of tonalitic gneiss, carbonate, and iron formation in the Dismal Ashrock (Kusky and Hudleston, 1999) would constitute pieces that were plucked tectonically from the passive-margin plate during collision. In Kusky and Hudleston's (1999) model, the passive margin would have formed after 3002 Ma and would have been involved in collision after 2780 Ma. If this model is correct and the 2780-Ma zircons are detrital, they might well approximate the age of collision, given that collision-related successions commonly incorporate only slightly older zircons. Even if this conjecture is true, the lack of a definitive start date leaves insufficient basis for estimating the lifespan of the margin, which could have been as great as 220 m.y. An alternative tectonic model is that the Dismal Ashrock is part of the Steep Rock Lake stratigraphy and is the product of plume-related magmatism on the site of a former carbonate platform (Tomlinson et al., 1999). In this model, the platform did not face an Atlantic-type ocean, but rather formed in a rift setting. Thus, for this synthesis, the Steep Rock Lake platform is regarded as an equivocal example of what *might* be the world's first passive margin, requiring more work.

A11. Medicine Bow orogen, Wyoming craton, USA

The Medicine Bow orogen of the Wyoming craton, western USA, has been interpreted as the product of arc-passive margin collision (Karlstrom et al., 1983). In the Wyoming craton, Archean basement and supracrustal rocks are unconformably overlain by a Paleoproterozoic succession that is thought to include rift (Deer Lake Group), passive margin (lower Libby Creek Group), and possible foreland-basin deposits (upper Deer Lake Group) (Karlstrom et al., 1983). The age of rifting is best constrained by a prominent mafic and ultramafic dike swarm (Kennedy dikes) dated by U-Pb at 2011±1 Ma (Cox et al., 2000). The rift-drift transition must be younger than this. Orogenesis involved collision with an arc to the south. Although stratigraphic age constraints on the age of collision are still lacking, deformation of the orogenic wedge is closely dated by the syndeformational emplacement of the Mullen Creek mafic complex at 1778±2 Ma (U-Pb zircon; Chamberlain et al., 1998). I place the start date at ca. 2000 Ma and the end date at ca. 1780 Ma, corresponding to a lifespan of about 220 m.y. for the Wyoming craton's passive margin.

A12 and A13. Southern passive margins of the Superior craton, Great Lakes Region, Canada and USA

What appear to be two successive, superimposed Paleoproterozoic passive margins are preserved on the southern edge of the Superior craton. The origin of the older margin (Huronian, ca. 2300 Ma) and the demise of the younger one (Animikie, ca. 1880 Ma) are recorded more clearly than the demise of the older margin or the origin of the younger one. Both margins were deformed during the Penokean orogeny (ca 1880-1830 Ma; Schulz and Cannon, 2007).

The Huronian margin (A12) is mainly recorded in exposures north of Lake Huron by the Huronian (or Huron) Supergroup, a southward-thickening siliciclastic sedimentary prism as great as 12 km thick (Hoffman, 1989). Young et al. (2001) subdivided the succession into informal lower Huronian and upper Huronian parts, which they interpreted as rift- and passive-margin deposits, respectively. The lower Huronian includes a variety of sedimentary units of only local distribution, including volcanic rocks, uraniferous conglomerates, diamictite, mudstone, and arkose. Down-to-south synsedimentary normal faults attest to episodic extension during deposition of the lower Huronian (Hoffman, 1989). Initial extension is recorded by the 2490-2450 Ma Matachewan dike swarm (Pye *et al.,* 1984), and by bimodal volcanic rocks in the Elliot Lake Group at the base of the lower Huronian. The regionally extensive upper Huronian units (Cobalt Group), which are nearly 5 km thick (Hoffman, 1989), show no evidence of fault-controlled deposition. According to Young et al. (2001), the upper Huronian was deposited along a passive margin, and the Gowganda Formation at its base was laid down during the rift-drift transition interval. Age controls on the Gowganda are indirect. The upper Huronian strata are cut by and therefore predate the Nipissing diabase, which has a U-Pb age of 2219±4 Ma (Corfu and Andrews, 1986). New geochronology from the glaciogenic Enchantment Lake Formation, in northern Michigan, equivalent to the Gowganda, provides additional age control. Its youngest detrital zircon population is 2317±6 Ma (Vallini et al., 2006), and its depositional age therefore can be no older. For present purposes, I assign an age of ca. 2300 to the Gowganda which would also be the start date of the margin. An alternative interpretation is that while the Huronian Supergroup does represent a passive margin, the rift-drift transition happened earlier, closer to the 2490-2450-Ma age of the Matchewan dikes (Pye *et al.,* 1984). Another possibility is that extension during lower Huronian times did not lead all the way to opening of an ocean basin.

The demise of the Huronian margin is problematic. No strata are known that might record a ca. 2200 Ma foreland basin. The Huronian Supergroup was deformed prior to, or during, emplacement of the ca. 2219 Ma Nipissing dikes (Corfu and Andrews, 1986). Some workers have interpreted this deformation as tectonic but others have favored soft-sediment deformation (Young et al., 2001). Thus, the immediate fate of the Huronian margin is unclear: it was intruded by dikes, suffered minor deformation that may or may not have tectonic significance, and stopped subsiding. I suggest that the Huronian did not experience a terminal collision, but rather that it spalled a ribbon microcontinent, leaving behind a new margin, the Animikie. In this scenario, the end date of the Huronian is same as the ca. 2065 Ma start date of the Animikie (see below). The combination of poor age control and debatable tectonic interpretations lead to a quality ranking of a low C for the Huronian margin. The suggested start and end dates imply a lifespan of about 235 m.y.

The existence of a later Paleoproterozoic passive margin (margin A13, here called the Animikie margin) on the southern border of the Superior craton is inferred from (1) evidence for regional extension long after deposition of the Huronian Supergroup ended; (2) the presence of an widespread foreland-basin succession attributed to the final days of the Animikie margin; (3) a major contractional orogenic event—the Penokean—that deformed the foreland basin; and (4) the presence of an exotic arc, the Pembine-Wausau terrane, whose arrival provides a straightforward reason for the orogeny and foreland basin. A problematic aspect of the inferred Animikie margin, however, is a paucity of pre-collisional sedimentary strata (Schultz and Cannon, 2007). The Chocolay Group of Michigan was once seen as representing the Animikie passive margin succession (Schneider et al., 2002), but the new age constraints of Vallini et al. (2006) instead link it to the Huronian margin. Schultz and Cannon (2007) offered one possible explanation: that the Animikie margin preserves no platformal sediments because it was a transform margin lacking thinned lithosphere. The time of origin of the margin is inferred from mafic dike swarms cutting basement rocks of the southern Superior craton, which include the Nipissing at 2219±4 Ma (Corfu and Andrews, 1986), the Fort Frances at 2077±4 Ma (Southwick and Day, 1983; Southwick and Halls (1987), and finally the Minnesota River Valley dikes at 2067±1 Ma (Schmitz et al., 2006). Assuming that seafloor spreading began not long after these youngest dikes intruded into the continental crust of the Superior craton, the start date was probably ca. 2065 Ma.

The Animikie foreland basin of Minnesota and western Ontario includes banded iron formation, siliciclastic turbidites, and mafic igneous rocks (Hoffman, 1987). The iron formations record platform drowning; tuff horizons within the iron formation in two locations have U-Pb zircon ages of 1878±2 and 1874±9 (Fralick et al. 1998; Schneider et al., 2002). Accordingly, I place the demise of the Animikie margin at 1880 Ma. The quoted ages imply a lifespan of 185 m.y.

A14. Northern margin of the Superior craton, Cape Smith orogen and Trans-Hudson orogens, Canada

The Paleoproterozoic northern margin of Canada's Superior craton is exposed discontinuously in the Cape Smith orogen in northern Quebec, and in the Trans-Hudson orogen in Hudson Bay and Manitoba. It has been widely interpreted as a passive margin (*e.g.*, Hoffman, 1987), although the Manitoba segment is problematic. In the Cape Smith belt, a rift sequence (Povungnituk Group) consists of 3 km of siliciclastic rocks overlain by 5 km of basalt and rare rhyolite. Rift-related igneous rocks in the Cape Smith belt have yielded U-Pb zircon ages that range from 2038 to 1918 Ma (Machado et al. , 1993, their Fig. 9). Similarly, several hundred kilometers to the south in the Hudson Bay region, diagenetic apatite in rift deposits of the Richmond Gulf Group is ca. 2025±25 Ma (U-Pb and Pb-Pb; Chandler and Parrish, 1989). In the Cape Smith belt, an allochthon of layered gabbro (Watts Group) is interpreted as representing oceanic crust, and has yielded a zircon date of 1998 Ma (Parrish, 1989). Apparently, seafloor spreading in the ocean basin was already underway before rifting had ended on the continental margin. Demise of the margin is dated by an 1870-Ma gabbro (U-Pb zircon; Parrish, 1989) that was intruded into transitional (stretched continental) crust of the Chukotak Group that had already been emplaced onto the Superior craton margin (Parrish, 1989). For this study, I place the start date at 2000 Ma and the end date at 1875 Ma, yielding a 125 m.y. lifespan for the passive margin.

A15. Eastern margin of the Superior craton, New Quebec orogen ("Labrador Trough"), Canada

The New Quebec orogen, or Labrador Trough of older literature, is a Paleoproterozoic collision zone between two Archean cratons: the Rae craton (or "Province") on the east and the Superior craton (or "Province") on the west (Hoffman, 1989). Machado et al. (1997) summarized the stratigraphy and geochronology. In the western part of the orogen, autochtonous and allochthonous strata record rifting and subsidence of a passive margin of the Superior craton. A basal rift sequence (Chakonipau Formation) of fluvial facies and basalt is intruded by a gabbro sill dated at 2169± 4 Ma (Rohon et al 1993). Higher in the section, felsic volcanic rocks dated at 2142+4/-2 Ma also would appear to predate the rift-drift transition. An overlying marine platform that includes an extensive, undated dolomite unit (Denault Formation) is interpreted as a passive margin succession (Hoffman, 1987; Hoffman and Grotzinger, 1989). The rift-drift transition is probably not much younger than 2142 Ma and I place it at 2135 Ma. An unconformity of unknown duration divides the passive margin from a younger succession consisting of iron formation, turbidites, and mafic and felsic volcanic rocks. Hoffman (1987) interpreted these rocks as a collisional foreland-basin succession. A U-Pb age of 1884±2 Ma from a mafic sill intruding turbidites provides the oldest age constraint on foreland-basin development; I place its initiation at ca. 1890 Ma. Together, these ages suggest a duration of 245 m.y. for the passive margin; the start date of the margin, however, could be younger than my age pick by tens of millions of years.

A16. Western margin of the Nain craton, Labrador, Canada

Archean basement of the Nain craton of easternmost Labrador is overlain by a 1.6-km-thick metasedimentary rock succession, the Ramah Group, that Hoffman (1987) identified as a likely passive margin to foredeep succession. As described by Knight and Morgan (1981), the lower one-third of the Ramah Group (Rowsell Harbour and Reddick Bight Formations) is a clastic-dominated shelf succession, consisting mostly of quartzite and mudstone; Hoffman (1987) interpreted this as a passive margin succession. A tholeiitic basalt flow near the base of the Rowsell Harbour Formation may constrain the age of the rift-drift transition. It has an Rb/Sr isochron age of 1892 Ma (Morgan, 1978), but the dated rock is strongly weathered and the age is probably unreliable (Mengel et al., 1991). Within the Torngat orogen to the west, arc magmatism has been dated at 1876 Ma (Bertrand et al., 1993), showing that an ocean basin existed by this time.

The upper two-thirds of the Ramah Group (Nullataktok, Warspite, Typhoon Peak, and Cameron Brook Formations) consists of a basinal succession deep-water carbonates, shales, minor pyritic iron formation, and at the top, turbiditic sandstones (Knight and Morgan, 1981). This part of the Ramah Group was interpreted by Hoffman (1987) as a foredeep succession. Mafic sills intrude rocks of the upper Ramah Group, which, if dated, may constrain the end date of the passive margin. The Ramah Group sedimentary basin was overthrust from the west by foreland thrusts of the Torngat orogen (Mengel et al., 1991), which is the product of collision between the Nain and Rae cratons. The oldest syntectonic granites related to collision are ca. 1859 Ma (Bertrand et al., 1993). The lifespan of the passive margin can only be estimated at greater than about 17 m.y.

A17. Makkovik Province, Labrador, Canada

The Archean Nain craton of eastern Canada is bordered on the north by a Paleoproterozoic orogen, the Makkovik Province. The following is from Ketchum et al. (2001). A succession of metasedimentary and mafic metavolcanic rocks, assigned to the Lower Aillik Group, is interpreted to represent a northern passive margin of the Nain craton. Although it is allochthonous with respect to the Nain craton, the Lower Aillik Group is correlated with cover rocks that rest unconformably on Nain basement. A mafic metavolcanic rock sequence in the Lower Aillik Group, dated at 2178±4 by U-Pb zircon (TIMS), has been regarded as marking the rift-drift transition (Ketchum et al., 2001), although I place the transition a few million years later, at 2175 Ma. Quartzites from low in the section have yielded only Archean detrital zircon grains, consistent with a Nain craton source. A psammite-semipelite sequence, interpreted to stratigraphically overlie the mafic rocks, was deposited after 2013 Ma, as constrained by the age of the youngest detrital zircon in a population dominated by Paleoproterozoic grains. Ketchum et al. (2001) interpreted the psammite-semipelite succession to represent a foredeep formed during the collision of an arc. An age of 2010 Ma for deposition is reasonable because the youngest detrital zircons in sandstones from most collisional foredeeps are rarely much older than the depositional age. A duration of about 165 m.y. is suggested for the Nain craton's Makkovik passive margin.

A18. Ouachita margin of Laurentia

The Ouachita margin of southern Laurentia is a continuation of the Appalachian margin, but with a different history after the Early Ordovician. The oldest platformal strata are latest Middle Cambrian (Thomas, 1991) and I place the rift-drift transition a bit earlier, at ca. 520 Ma. Unlike the Appalachians, the Ouachita margin escaped arc collision until the Carboniferous. Flysch, then molasse, inundated the carbonate platform during Atokan time (ca. 310 Ma) (Bradley and Leach, 2003). The Ouachita margin thus lasted ca. 210 m.y., nearly three times as long as the Appalachian margin, although both originated at about the same time.

A19. Appalachian margin of Laurentia (a) and northwestern Scotland (b)

The eastern (Appalachian) margin of eastern North America is described at length in Section 4.2. The start date is ca. 540 Ma and the end date ca. 465 Ma, giving a lifespan of about 75 m.y.

A20. Caledonian margin of East Greenland (a) and northeastern Svalbard (b)

East Greenland and northeastern Svalbard have closely comparable Neorproterozoic to Ordovician successions that are interpreted as portions of the same passive margin (*e.g.*, Harland et al., 1992; Fairchild and Hambrey, 1995). The two areas were first offset from one another along Devonian sinistral strike-slip faults, and later ended up on opposite sides of the North Atlantic (Harland et al., 1992). They are discussed together here.

The *Paleozoic* history seems quite clear and comparable to that of the Laurentian passive margin in the Appalachians and Scotland (margins A19a and A19b). During the Cambrian and Ordovician, carbonate-dominated miogeocline faced the Iapetus Ocean to the east. In far-traveled allochthons in central East Greenland, it reaches 4 km in thickness (Higgins et al., 2004). The demise of the margin is recorded in the extreme north of East Greenland

(Kronprins Kristians Land). In this region, at the Ordovician-Silurian boundary, east-derived turbidites flooded the former platform and heralded the advance of Caledonian thrust sheets (Hurst et al., 1983). This sets the end date of the margin at 444 Ma.

The initiation of the East Greenland-Northeast Svalbard margin presents a more difficult problem. A widely held view (*e.g.,* Smith et al., 1999, 2004) is that breakup along the East Greenland-Northeast Svalbard margin took place at about the time as the other Paleozoic margins of Laurentia, around the Neoproterozoic-Cambrian boundary. A vast thickness (about 15 km in Greenland and 12 km in Svalbard) of underlying Neoproterozoic platformal strata, however, might also have been deposited on a passive margin. The depositional history of these rocks is summarized below from Halverson et al. (2004). In both northeast Svalbard and East Greenland, the Neoproterozoic successions begin with clastics inferred to be related to rifting (Nathorst Land and Lylell Land Groups in Greenland, Planetfjella and Veteranen Groups in Svalbard). Next come platformal carbonates (Ymer Ø. and Andrée Land Groups in Greenland, Akademikerbreen Group in Svalbard). The Neoproterozoic section is completed by approximately 1 km of glaciogenic strata (Tillite Group in Greenland, Polarisbreen Group in Svalbard). Chemostratigraphy suggests that the glacial succession includes correlatives of both the Sturtian (ca. 740-710 Ma) and Marinoan (ca. 635 Ma) glaciations (Halverson et al., 2004; Maloof et al., 2006). The glaciogenic strata, in turn, are overlain by the Cambrian and Ordovician carbonates that are generally viewed as passive margin deposits.

Age constraints and tectonic interpretations of Neoproterozoic events are both equivocal. Maloof et al. (2006) combined meager geochronological control, correlations based on chemostratigraphy, and tectonic subsidence analysis to generate the ages used here. A transition from rift-driven to thermal subsidence predated the Bitter Springs Stage recognized in the carbon isotope record, and is placed at about 815 Ma. Three possible tectonic interpretations are as follows. (1) The Neoproterozoic strata were deposited along the same passive margin as the Cambrian and Ordovician strata. In this case, the 815-Ma event would be the rift-drift transition and the lifespan of the margin would thus be about 370 m.y. (2) There were two passive margins: a Neoproterozoic passive margin that formed at 815 Ma, and a Cambrian-Ordovician one that formed when a ribbon continent containing the distal part of the old margin split away. In this case, the start date of the older margin would be 815 Ma, the end date of the first margin and the start date of the second would be ca. 542 Ma, and the end date of the second would be 444 Ma. This would correspond to lifespans of about 275 m.y. for the first margin and 98 m.y. for the second. (3) The Neoproterozoic rocks represent a rift-sag sequence, that did not proceed all the way to seafloor spreading (Smith et al., 1999, 2004). In this case, the Iapetus Ocean did not open until about the Neoproterozoic-Cambrian boundary, giving dates of 542-444 Ma for the margin, and a duration of about 98 m.y. There is no evidence for the existence of either deep-water facies or a sedimentary source to the east of the Neoproterozoic carbonate platform; either of these would help discriminate between the tectonic scenarios. Given 200-400 km of Caledonian shortening (Higgins et al., 2004), such paleogeographic uncertainties are not surprising. For this paper, I tentatively adopt this first scenario, but with a quality ranking of C.

A21. Caledonian margin of Baltica

The Caledonian (western) margin of Baltica in Scandinavia formed by rifting in the late Neoproterozoic and was destroyed at about the Cambrian-Ordovician boundary by collision. Evidence for the timing of rifting was summarized by Kumpulainen and Nystuen (1985, p. 225-226) and Torsvik et al. (1996, p. 240). They put the rift-drift transition at 600 to 580 Ma. More recently, a U-Pb TIMS zircon age of 608±1 Ma has been reported from the Sarek tholeiitic dike swarm that intrudes rift-related metasedimentary rocks in the Sarektjåkkå Nappe (Svennigsen, 2001). I place the rift-drift transition at ca. 605 Ma. Demise of the resulting passive margin is recorded by eclogite-facies metamorphism of Baltic rocks at 505 Ma (Roberts, 2003). This event is called the Finnmarkian orogeny and was followed by later Caledonian orogenic phases. Collision with an arc within the Iapetus Ocean is the likely cause (Stephens and Gee, 1985). The passive margin's lifespan was about 100 m.y.

A22. Kola suture belt, northern Europe

The Kola-Karelia orogen of the northern Baltic craton has been interpreted as the product of arc-passive margin collision (Berthelsen and Marker, 1986; Zhao et al., 2002), the passive margin having been on the southern edge of the Kola craton. Melezhik and Sturt (1994) documented a long history of mainly subaerial rifting from ca. 2.6 to 2.0 Ga. In their model, rifting eventually gave way to seafloor spreading starting ca. 1970 Ma, and arc collision at ca. 1800 Ma. These age picks suggest a lifespan for the passive margin of about 170 m.y.

A23. Timanian orogen, northern Baltica

The Timanian orogen along Europe's Arctic coast includes a discontinuously exposed, Neoproterozoic passive margin that apparently was connected with the southern portion of Baltica's Uralian margin (Maslov, 2004). The ages of rifting and then passive-margin subsidence are not tightly constrained. According to Siedlecka et al. (2004, p. 176), this margin probably is as old as late Mesoproterozoic, although mainly Neoproterozoic. In the absence of definitive local age control along this portion of the Baltic margin, the age of the rift-drift transition is extrapolated at ca. 1000 Ma from the southern Urals (margin A24), farther south along the east side of Baltica. The Timanian passive margin is overlain by a succession of northerly-derived siliciclastic rocks shed from an outboard orogenic source formed during collision of an arc. These strata are interpreted as the fill of a foreland basin (Grazhdankin, 2004). A U-Pb zircon age of 558±1 Ma has been reported from an ashfall tuff in the Verkhovka Formation in the lower part of the succession (Grazhdankin, 2004), indicating that orogeny was already underway by this time. Allowing a few million years to accumulate >500 m of older foreland-basin sediments, the passive-margin to foreland basin transition is here placed at ca. 560 Ma. Thus, the margin has an apparent lifespan of about 440 m.y.

A24 and A25. Eastern (Uralian) margin of Baltica

Paleoproterozoic (ca. 2.3-1.8 Ga) metamorphic and igneous basement rocks of the eastern Baltic craton are overlain by a 12- to 15-km-thick, unmetamorphosed sedimentary succession (Willner et al., 2003). These Mesoproterozoic and Neoproterozoic strata comprise an eastwardly-thickening sedimentary prism in which three subdivisions (Lower, Middle, and Upper Riphean) are recognized (Chumakov and Semikhatov, 1981). Each of these intervals appears to have involved a cycle of extension and then thermal subsidence, but only the third cycle ended with collision.

The Lower Riphean (Burzyan Group, ca. 1650-1350 Ma) is 5500 to 6000-m-thick (Maslov et al., 1997). It begins with sandstone, conglomerate, and trachybasalt, overlain by quartzose sandstones and dolostones. The stratigraphic reconstruction of Chumakov and Semikhatov (1981, their Fig. 3) shows the Lower Riphean rocks as an eastward-thickening miogeoclinal prism, with no sign of a basin margin to the east. The Middle Riphean (Yurmatau Group, ca. 1350-1000 Ma) is 5000 to 6000-m thick and starts with sandstones, conglomerates, and bimodal volcanic rocks, overlain by sandstones, black shales, and toward the top, carbonates. The Middle Riphean also thickens eastwardly to form a classic miogeoclinal prism (Chumakov and Semikhatov, 1981, their Fig. 3), again with no sign of a basin margin to the east. Whereas Maslov et al. (1997) interpreted the Lower Riphean and Middle Riphean cycles in terms of rifting and thermal subsidence, they did not believe that either led to seafloor spreading or a true passive margin. However, given the absence of any known land to the east, that possibility cannot be discounted.

The Upper Riphean section (Karatau Group, ca. 1000 to 650 Ma) includes mostly siliciclastic rocks in the lower part and mostly carbonate rocks in the upper part. This is a well-defined passive margin (margin A24) that ended with collision. The age of the Middle Riphean-Upper Riphean boundary is bracketed between gabbroic sills that intrude the Middle, but not the Upper Riphean (1000±20 Ma, K/Ar) (Maslov et al. 1997, p. 319 and references therein), and a date of 970 Ma from near the top of the Zilmerdak Formation, the lowest formation in the Upper Riphean (Maslov et al. 1997, p. 318 and references therein). The rift-drift transition probably took place ca. 1000 Ga. The Upper Riphean passive margin is overlain by the Vendian, a late Neoproterozoic siliciclastic succession 2- to 3-km-thick (Willner et al., 2003). Heavy mineral populations in the Upper Vendian contrast strongly with those from the Upper Riphean and Lower Vendian. Perhaps most significant is an influx of detrital phengite, derived from a Neoproterozoic high-pressure metamorphic belt to the east (Willner et al., 2001), the Beloretsk terrane (Glasmacher et al., 2001). The Upper Vendian is interpreted as having been deposited in a foreland-basin environment (Puchkov, 1997), which would appear to have formed when the Baltic passive margin began to be subducted to the east under the approaching Beloretsk terrane. Willner et al. (2001) put the provenance shift at ca. 620 Ma, and Maslov et al. (1997) put the Lower-Upper Vendian boundary at ca. 610-620 Ma, by interpolating between Rb-Sr and glauconite K-Ar dates. For this paper, the demise of the passive margin is placed at 620 Ma, giving the Neoproterozoic Uralian margin an relatively long lifespan of 380 m.y.

The Paleozoic Uralian margin of the Baltic craton (margin A25) approximately follows the southern portion of the Neoproterozoic margin. However, the older margin connected to the northwest with the Timanian margin, whereas the younger margin continued to the north to Novaya Zemlaya. Brown et al. (2006) provided a modern review. Upper Cambrian to Lower Ordovician (Tremadocian) rift facies are locally preserved (Puchkov, 1997, Puchkov et al., 2002; Brown et al., 2006). The rift-drift transition is placed at ca. 477 Ma. Starting in the Ordovician, a passive margin can be recognized, characterized by platformal facies to the west and bathyal facies to the east. The margin persisted until Late Devonian when it began to collide with the Magnitogorsk arc over an east-dipping subduction zone (Brown et al., 2006). The earliest sign of collision is in the southern Urals, where easterly-derived late Frasnian flysch was deposited atop cherts of the Baltica margin (Puchkov, 1997, p. 223). The demise of the passive margin is therefore placed at 376 Ma. Puchkov (1997) has likened the initial (Late Devonian-Early Carboniferous) Uralian collision to the modern collision between Australia and the Banda Arc. The better known main stage of Uralian orogenesis took place from mid-Carboniferous to Late Permian, long after the passive margin was destroyed. The start and end dates imply a lifespan of about 101 m.y.

A26. Variscan margin of Baltica

The Variscan passive margin of Baltica was a relatively short-lived feature that formed by rifting in the Devonian and was destroyed in the Early Carboniferous. In parautochthonous sections in the Rhenish Massif, Neoproterozoic basement is overlain by rift-related siliciclastic strata as old as late Lochkovian (Franke, 2000, p. 36). The oldest remnants of Rheno-Hercynian seafloor are Emsian or slightly older (Franke, 2000, p. 39). For this study, the age of the rift-drift transition is placed at the base of the Emsian, 407 Ma. Variscan convergence was underway in one of the oceanic allochthons (Giessen Werra Südarz /Selke Nappe) by Frasnian time (Franke, 2000, p. 39). Synorogenic clastic sedimentation had begun by the late Tournasian in the parautochthon, and I put the end date of the passive margin at this time (ca. 347 Ma). The lifespan of the Rheno-Hercynian margin thus appears to have been about 60 m.y.

A27. Saxo-Thuringian block (Bohemian Massif), Germany

The Saxo-Thuringian sector of the Armorican microcontinent includes the deformed remnants of a passive margin on the south side of the Rheic Ocean (Linnemann et al., 2004). Cambrian and Ordovician rift-related strata include Cambrian conglomerates and Lower Ordovician (ca. 490 Ma) mafic volcanic rocks (Linnemann et al., 2004, their Fig. 3). The passive margin phase is represented by earliest Silurian to Tournaisian (early Carboniferous) limestones and shales. Detrital zircon and Nd-isotopic data suggest that Saxo-Thuringia was part of the Gondwana margin during this entire time span; the presence of uppermost Ordovician (Saharan) glacial deposits in Saxo-Thuringia also implies a Gondwana connection (Linnemann et al., 2004). Demise of the passive margin is recorded by an influx of Variscan flysch during the Early Carboniferous (Linnemann et al., 2004). I place the start date of the Saxo-Thuringian margin at ca. 444 Ma and the end date at ca. 344 Ma, giving a lifespan of about 100 m.y.

A28. Swiss Alps

Permian and Triassic rifting of the Hercynian basement of western Europe led to seafloor spreading in the Jurassic, forming the Piemonte Ocean, which later closed during the Alpine orogeny. Ziegler et al. (2001) placed the rift-drift transition in the Bajocian (ca. 170 Ma). The demise of the margin is stratigraphically well-constrained in the North Alpine foreland basin. Cretaceous passive margin carbonates are unconformably overlain by *the* classic upward deepening, then progradational foreland-basin succession of Flysch, and then Molasse. The unconformity likely formed at a forebulge. The foreland-basin succession began to be deposited partway through the Lutetian (ca. 43 Ma) (Allen et al. 1991). This overlaps with 50- to 40-Ma high-pressure metamorphism in the Piemonte domain (Rosenbaum and Lister, 2005). Ages of 170 and 43 Ma for the start and end dates yield a lifespan of about 127 m.y.

A29. Pyreneean-Biscay margin of Iberia

The evolution of this short-lived margin is discussed in Section 10.2.

A30. Paleozoic margin of northwestern Iberia

A telescoped, metamorphosed Ordovician to Devonian passive margin sequence is recognized in the northwestern part of the Iberian microcontinent (Gonzales Clavijo and Martinez Catalan, 2002). Preserved in a late Paleozoic (Variscan) thrust belt, the inferred passive margin formed along what was then the northern margin of Gondwana. Early Ordovician gneissic volcanics are overlain unconformably by a siliciclastic platformal sequence whose basal unit, the Santa Eufemia Formation, begins in the early Arenig of the Early Ordovician (Gonzales Clavijo and Martinez Catalan, 2002). Ordovician rift sequences have not been recognized as such, so the base of the platformal succession is taken as the rift-drift transition, which I round to 475 Ma for lack of tight age control. Silurian strata show evidence of soft-sediment deformation and include volcanic sills, flows, and tuffs; Gonzales Clavijo and Martinez Catalan (2002) interpreted the Silurian as a time of extension and thinning of the already established passive margin. Alternatively, I suggest that this inferred extension was the consequence of lithospheric flexure at the onset of collision. Devonian foreland basin deposits of flysch-like character are recognized in three thrust sheets (Almendra, San Vitero, and Rabano Formations); these units each contain metamorphic pebbles from the nascent Variscan thrust belt. Correlative foreland-basin turbidites in Portugal are early Frasnian (Late Devonian) in age (Gonzales Clavijo and Martinez Catalan, 2002 and references therein), and these are inferred to mark the demise of the margin. The end date is placed at ca. 385 Ma, giving a lifespan of 90 m.y.

A31. Apulian microcontinent, Greece, eastern margin

The Apulian microcontinent of the Adriatic region faced an ocean that formed by rifting in the Triassic and was destroyed by collision in the Cretaceous (Degnan and Robertson, 1998). Rifting from Late Permian(?) to Middle Triassic (Dercourt et al., 1986) led to development of a carbonate platform that was flanked to the east by the Pindos Ocean. Slope, rise, and abyssal facies of the Priolithos Group are regarded as having been deposited in this ocean, and the oldest of these sediments are said to be "Mid?-Late Triassic" (Degnan and Robertson, 1998), or ca. 230 Ma. The passive margin endured until the Paleocene when terrigenous turbidites of the Pindos Flysch Formation mark collision between the Apulian passive margin and an accretionary complex to the east, at the leading edge of the Pelagonian block (Degnan and Robertson, 1998). I estimate an age of ca. 60 Ma for the base of the Pindos Flysch Formation and for the demise of the passive margin. These ages imply a lifespan of about 170 m.y.

A32 and A33. Isparta Angle, Turkey

The Isparta Angle in southern Turkey contains two conjugate passive margins of Mesozoic age that face one another but have somewhat different subsidence histories (Dilek and Rowland, 1993). The region was already a carbonate platform (along the north side of Neotethys) before rifting began in Middle Triassic, at a high angle to the preexisting margin (Dilek and Rowland, 1993, p. 964). Carbonate platforms were established on both the western (Bey Daglari, A32) and eastern (Anamas-Akseki, A33) margins by Late Triassic. Accordingly, I place the rift-drift transition at ca. 227 Ma. Demise of the intervening ocean is recorded on the Bey Daglari margin by the late Paleocene to early Eocene olistostromes containing ophiolitic debris; I estimate this age at 60 Ma. Demise of the Anamas-Akseki margin is recorded by early Eocene flysch at ca. 53 Ma. Accordingly, the Bey Daglari margin existed for about 170 m.y. and the Anamas-Akseki margin for about 177 m.y.

A34. Northeastern margin of Arabia, Oman and Zagros

The Late Cretaceous orogeny in Oman is one of the better documented arc-passive margin collisions. The Oman margin can be traced northward along the Zagros orogen in Iran, where younger deformation complicates things. The passive margin formed in the late Paleozoic along the Arabian sector of Gondwana's northern margin, either through normal, Atlantic-type spreading (*e.g.,* Stampfli et al., 1991) or back-arc rifting (Sengör, 1990). Rift-related strata formed in the upper plate and are Early Permian in age (Stampfli et al., 1991). Seafloor spreading apparently was underway by earliest Late Permian time, judging from the age of the oldest dated pelagic facies that would later be thrust back onto the platform (Stampfli et al., 1991). Accordingly I place the rift-drift transition at 272 Ma, near the start of the Late Permian. The passive-margin platform, which is dominated by carbonates, spanned Late Permian to Late Cretaceous. Onset of collision is dated by a Turonian forebulge unconformity, followed by abrupt deepening and flysch sedimentation in the Coniacian (Gealey, 1977; Robertson, 1987) (ca. 87 Ma). This was followed by thrusting of deep-water facies (Hawasina Compex) and the recently formed (ca. 95 Ma; Tilton et al., 1981) Semail forearc ophiolite onto the former platform. The passive margin lasted about 185 m.y.

A35. Late Paleozoic margin of northern Iran, Alborz orogen

The Alborz orogen in northern Iran contains the telescoped remnants of a Devonian to Triassic passive margin that bordered Paleo-Tethys on the north side of Gonwana (Stampfli et al., 1991). A Neoproterozoic to Lower Ordovician platformal sequence of siliciclastics, carbonates, and minor tuffs is considered to predate rifting that led to the Alborz margin (Alavi, 1996), and may belong to an older passive margin. Mid-Ordovician to mid-Devonian mainly mafic volcanics about 1 km thick have been attributed to extension (Alavi, 1996). Overlying mid-Devonian to Middle Triassic strata are predominantly carbonates and represent a passive margin that faced the Paleotethys ocean to the north (Stampfli et al., 1991). Demise of the passive margin is marked by Upper Triassic to Lower Jurassic turbidites of the Shemshak Formation, which were derived from the north and reach 3 km in thickness (Alavi, 1996). The Shemshak is interpreted as a foreland-basin fill related to Cimmeride collision. The generalized stratigraphic section of Alavi (1996) implies a mid-Devonian rift-drift transition, which I place at ca. 390 Ma, and a Late Triassic passive-margin to foredeep transition, which I place at ca. 210 Ma. Thus the margin had a lifespan of about 170 m.y.

A36 and A37. Taimyr (northern) margin of Siberian craton

The Siberian craton's northern margin appears to have been the site of successive passive margins, first in the Neoproterozoic (margin A36) and then in the Paleozoic (margin A37). Fragmentary evidence for the older of these comes from the Taimyr Peninsula where "ophiolites and island-arc complexes were formed around 750 Ma and emplaced onto the Arctic Siberia passive margin around 600 Ma" (Vernikovksy and Vernikovskaya, 2001, p. 138). There are no direct age constraints bearing on when this margin formed, as noted by Pisarevsky and Natapov (2003). The putative end date of ca. 600 Ma is based on a range of metamorphic ages in the Stanovoy ophiolite belt. This margin is assigned a lifespan of >150 m.y. and a quality ranking of D.

A second passive margin appears to have formed on the same general location about 70 m.y. later, a margin that formed in the Cambrian and was destroyed in the Carboniferous. The rift-drift history was assessed by Pelechaty (1996) and Pelechaty et al. (1996) based on work near the junction of the Siberian craton's Taimyr (northern) and Verkhoyansk (eastern) passive margins. Latest Vendian carbonates (Khorbosuonka Group) are regarded as belonging to a preexisting east-facing continental margin sequence along the Verkhoyansk sector (margin A57). A paleokarst surface separates these strata from Lower Cambrian Nemakit-Daldyn strata which include interbedded conglomerate, sandstone, shale, limestone, and volcanic rocks. In northeastern Siberia, these rocks were deposited in a southwest-facing basin, interpreted as a rift basin. Tommotian (mid-Lower Cambrian) limestones of the Erkeket Formation record regional flooding of the craton and onlapped from north to south. Pelechaty (1996) placed the rift-drift transition at the base of the Erkeket Formation, at ca. 530 Ma. The passive margin was destroyed during a late Paleozoic orogeny. This orogeny has been dated by Late Pennsylvanian to Early Permian thrusts and Permian granitic plutonism and metamorphism in the central Taimyr zone (Zonenshain et al., 1990). This suggests an age of ca. 325 Ma for the demise of the margin, giving a lifespan of about 205 m.y.

A38. Western margin of Siberian craton, Yenisei Range

The Siberian craton's Proterozoic western margin is known from outcrops in the Pre-Sayan area, the Yenisei Range (or Yenisei Ridge), and the Turukhansk and Irgarka Uplifts (Pisarevsky and Natapov, 2003). The Yenisei Range exposes the most complete Mesoproterozoic to Neoproterozoic successions, which comprise a miogeocline that thickens and deepens to the west, away from the craton. This succession includes the Lower Riphean Korda Formation (interpreted as rift deposits), the Middle Riphean Sukhoi Pit Group, and the Upper Riphean Tungusik and Oslyanka Groups. Rocks of the Korda Formation are entirely siliciclastic, whereas all the other units are mixed carbonate and siliciclastic, with carbonates being more abundant toward the craton (Pisarevsky and Natapov, 2003). Syncollisional metamorphic biotites from the Korda Formation yielded $^{40}Ar/^{39}Ar$ plateau ages of 823 and 826 Ma (Likhanov et al., 2007). Allowing for 25 m.y. for exhumation of the dated rocks through the argon closure temperature for biotite, Likhanov et al. (2007) concluded that collision-related metamorphism was not older than 848-851 Ma. In their review, Pisarevsky and Natapov (2003) suggested that the Yenisei passive margin existed between 1350 and 850 Ma, a lifespan of about 500 m.y.

A39. Gargan microcontinent, central Asia

Kuzmichev et al. (2001) identified a Neoproterozoic passive margin on the Gargan microcontinent, part of the Precambrian Tuva-Mongolia massif in central Asia. Basement rocks of the Gargan microcontinent have not been dated by modern methods but they are likely Archean (Kuzmichev et al., 2001). The inferred passive margin succession consists of the Irkut Formation, up to 600 m of dolomitic marbles with a basal conglomerate of metamorphic debris. No rift deposits have been identified, so the base of the Irkut Formation is the best approximation of the rift-drift transition. Rocks of the Irkut Formation are overlain by dark shale and minor interbeds of sandstone and limestone of the Ilchir Formation. The top of the Ilchir Formation contains olistostromes and ophiolitic melange. An ophiolitic allochthon, probably of forearc origin, was thrust over rocks of the Ilchir Formation. This event implies a foredeep depositional setting for the Ilchir Formation; the Irkut-Ilchir contact would then represent the passive margin to foredeep transition. The one age constraint for ophiolite obduction is provided by the Sumusunur plutonic suite, dated at 785±11 Ma (U-Pb TIMS; Kuzmichev et al., 2001). The igneous rocks cut the thrust contact and thus postdate ophiolite emplacement. The start date for this passive margin is unknown, although Kuzmichev et al. (2001, p. 121) assumed that the margin formed some time in the Mesoproterozoic. Better age control is needed; the quality ranking is D.

A40. Southern margin of Siberian craton, Baikal region

The Siberia craton is generally thought to have had a passive margin on its south side during the Neoproterozoic (*e.g.*, Pisarevsky and Natapov, 2003; Vernikovsky et al., 2004). Khain et al. (2003, p. 343) placed the onset of rifting at ca. 1100 Ma and the oldest record of oceanic crust at ca. 1000 Ma, the age picked here for the rift-drift transition. The passive margin succession consists of Neoproterozoic siliciclastic and carbonate rocks, including shelf facies in the north and slope facies in the south (Vernikovsky et al., 2004). Ophiolites and island-arc rocks of the Baikal-Muya Complex (ca. 900-812 Ma) were thrust over the passive margin in "pre-Vendian or Vendian time" (Vernikovsky et al., 2004) (ca. 600 Ma). These imprecise constraints suggest a nominal lifespan of about 400 m.y for the passive margin.

A41. Dzabkhan Basin, Mongolia

The Dzabkhan (or Zavkhan, or Zavhan**)** Basin in southwestern Mongolia preserves a passive-margin carbonate sequence that evolved into a foredeep during the Tuva-Mongol arc-microcontinent (Macdonald et al., in review, 2008). Basement rocks of the Dzabkhan terrane include gneisses dated at 1868±3 and felsic volcanics dated at 850±2 and 750±3 Ma (Badarch et al., 2002). The passive-margin sequence began following the mid-Cryogenian (Sturtian) glaciation at ca. 710 Ma and ended ca. 580 Ma (Macdonald *et al.,* in review, 2008). These ages yield a lifespan of ca. 130 m.y.

A42. Idermeg terrane, Mongolia

The supracrustal succession of the Idermeg terrane in eastern Mongolia consists of marble, quartzite, conglomerate, sandstone, and limestone deposited on an older metamorphic basement (Badarch et al., 2002). Archeaocyathids and stromatolites suggest a Neoproterozoic to Cambrian age for the limestones. Middle to late Cambrian plutons intrude the succession. Badarch et al. (2002) categorized the Inderbeg terrane as a passive margin but details of its evolution and timing are unclear. I assign it a quality ranking of D.

A43. Karakorum margin, Pakistan

The Karakorum block is one of the large series of broadly related Tethyan microcontinents that split away from the northern margin of Gondwana, drifted north, and collided with Eurasia. The following is from Gaetani (1997). Ordovician and Silurian strata consist of mixed siliciclastic and shallow-marine carbonates, and are interpreted to predate the passive margin. Devonian rifting is suggested by thick basaltic lavas and tuffs but this apparently did not proceed all the way to seafloor spreading, because Carboniferous strata, of mixed siliciclastic and carbonate facies, are thin. Renewed rifting in the Early Permian is marked by thick (ca. 1 km) terrigenous clastics of the Gircha Formation. Based on Gaetani's (1997) summary, I place the rift-drift transition in the Guadalupian (Late Permian, about 268 Ma). The ensuing passive margin is represented by Upper Permian through lowermost Jurassic platform carbonates. This inferred upper-plate margin faced deeper water to the north (Gaetani, 1997, his Fig. 6). During the Early Jurassic, peritidal carbonates gave way to slope breccias, which are overlain by sandstones containing volcanic, serpentite, sedimentary, and metasedimentary detritus. These siliciclastic strata are inferred to represent an Eo-Cimmerian foreland basin that came into existence in the earliest Jurassic, ca. 193 Ma. The margin lasted about 75 m.y.

A44. Northern margin of Tarim microcontinent (southern Tian Shan), central Asia

The northern margin of the Tarim microcontinent, along the southern side of the Tian Shan range in central Asia, has been interpreted as passive margin (*e.g.*, Allen et al., 1992; Carroll et al., 1995, 2001; Zhou et al., 2001) that formed in the Neoproterozoic and collided with an arc in the Paleozoic. Along the northwestern part of the margin, Carroll et al. (2001) recognized a Neoproterozoic through Middle Ordovician megasequence that includes rift-related clastic and volcanic rocks, and overlying platformal carbonates inferred to reflect passive-margin subsidence. All but the youngest Neoproterozoic strata are likely products of a rifting environment (*e.g.*, mafic flows or sills and boulder conglomerates; Carroll et al., 2001). Along the northeastern margin of the Tarim block, the youngest Neoproterozoic (presumably rift-related) volcanic rocks are ca. 600 Ma (Shuiquan Formation; Xu et al., 2005), and I therefore place the rift-drift transition at ca. 600 Ma. There is a broad consensus that the northern Tarim passive margin met its demise by collision following north-dipping subduction beneath an arc in the central Tian Shan (Allen et al., 1992; Carroll et al., 2001; Zhou et al., 2001, Xiao et al., 2004). The timing of the *onset* of collision is uncertain. The earliest possibility is at about the Ordovician-Silurian boundary, corresponding to a dramatic shift from carbonate to clastic sedimentation. Carroll et al. (2001), however, attributed these clastics to a distant orogeny on the other side of the Tarim block. A second possibility is at about the Middle-Late Devonian boundary, when the subsidence rate increased dramatically along the north Tarim margin (Carroll et al., 2001, their Fig. 15). An angular unconformity separates the Neoproterozoic through Devonian section from overlying Carboniferous rocks, and this otherwise unexplained deformation would thus be interpreted as a result of collision. A third possibility is that the collision took place in the Carboniferous or Permian, which is consistent with the generally accepted view of the Lower Carboniferous to Lower Permian succession as the fill of a foreland basin (Carroll et al., 1995). Following the suggestion of Carroll et al. (2001) that *initial* collision began in Late Devonian, I place the demise of the margin at ca. 380 Ma. The selected start and end dates correspond to a lifespan of about 220 m.y.

A45. Kunlun margin, southern side of Tarim microcontinent, central Asia

The Tarim microcontinent in central Asia has been interpreted as having a Neoproterozoic to Paleozoic passive margin along its southern side in the North Kunlun Range. A general model involves Sinian (Neoproterozoic) breakup and early Paleozoic collision (Mattern et al., 1996, p. 708; Mattern and Schneider, 2000, p. 645; Xiao et al., 2002, p. 524; Xiao et al., 2003, p. 322). In the North Kunlun, Precambrian gneisses are overlain by a weakly metamorphosed succession of late Neoproterozoic carbonates and oceanic tholeiites, as well as lesser shales, marls, and tuffs. Mattern and Schneider (2000, p. 638) interpreted these rocks as "a rift sequence which formed during the fragmentation of a subsiding marine platform". They suggested that rifting was followed in the later Sinian by seafloor spreading. For present purposes the rift-drift transition is placed at ca. 600 Ma. An episode of south-directed subduction in the Cambrian and Ordovician led to arc-passive margin collision, which produced a mid-Silurian to Devonian foreland basin (Wei et al., 2002). The end date of the passive margin is here placed at ca. 430 Ma. The corresponding lifespan of the passive margin is about 170 m.y., subject to a large uncertainty as to the start date.

A46 and A47. Himalayan margin of India

The classic Himalayan margin of northern India (margin A47, Permian to Paleocene, see below) was preceded by an older passive margin (margin A46) that formed and was destroyed during a Neoproterozoic to early Paleozoic Wilson Cycle. The Cenozoic Himalayan orogeny has severely telescoped the Neoproterozoic margin. Myrow et al. (2003) have argued that, despite this young shortening, proximal to distal parts of this ancient margin can still be recognized in the Lesser, Greater, and Tethyan Himalaya. In the Lesser Himalaya, which can be most confidently linked to the Indian craton, the Neoproterozoic succession is about 12 km thick (Jiang et al., 2002). The poorly dated lower half includes quartzite, sandstone, argillite, carbonate rocks, and mafic volcanic rocks, which are all younger than 1 Ga, and may be related to rifting (Jiang et al., 2002, 2003A). Glaciogenic strata of the Blaini Formation, up to 2 km thick, unconformably overlie the rift package; Jiang et al. (2002, 2003A) interpreted the rift-drift transition to fall within or perhaps at the base of the glacial interval. The glacial deposits are not directly dated, but they have been correlated with the Nantuo glaciation in South China (Jiang et al., 2003B), the end of which is now precisely dated at 635 Ma (Condon et al., 2005). I place the rift-drift transition at 635 Ma. Overlying strata of the passive margin phase are assigned to the Neoproterozoic Infra Krol Formation and Krol Group; these rocks are thought to represent the thermal subsidence phase and consist of mainly shallow marine carbonate rocks of a seaward-thickening, seaward-deepening prism. The top of the Krol Group is near the Neoproterozoic-Cambrian boundary (542 Ma) (Jiang et al., 2003A). Rocks of this sequence are overlain by the Cambrian Tal Group, a 2-km-thick transgressive, then progradational succession of phosphatic chert, argillite, and sandstone (Hughes et al., 2005). Brookfield (1993) regarded the Tal Group as a clastic-dominated part of the passive margin succession. However, as depicted by Hughes et al. (2005, their Fig. 4), the Tal Group sequence is reminiscent of collisional foreland-basin successions such as the Taconic; this possibility warrants a closer look.

Demise of the passive margin is most confidently dated in the Greater Himalaya, where the carbonate platform (Karsha Formation) was inundated by siliciclastic rocks near the end of the Middle Cambrian. The Kurgiakh Formation includes Middle Cambrian shales with interbedded tuffs, and Upper Cambrian siliciclastic turbidites (Garzanti et al., 1986). Garzanti et al. (1986, their Fig. 11) interpreted the Kurgiakh clastic rocks to record the first encounter of passive margin with an advancing accretionary prism; according to that interpretation, the end date for the margin would be ca. 502 Ma (tuned to Myrow et al.'s (2004) age pick of latest Middle Cambrian for the Karsha-Kurgiakh contact). Gehrels et al. (2003) reviewed evidence for Cambro-Ordovician orogeny in the Himalaya; effects included emplacement of granite plutons, penetrative deformation, and regional garnet-grade metamorphism. A granite dated at 488 Ma (about the Cambrian-Ordovician boundary) cuts sillimanite- and kyanite-bearing schists (Marquer et al., 2000), showing that deformation began before Ordovician time. I place the start date for the passive margin at ca. 635 Ma and the end date at ca. 502 Ma, for a duration of about 133 m.y. An alternative tectonic interpretation, proposed recently by Cawood et al. (2007B), is that during Cambrian time, the passive margin converted directly to an Andean-type margin; in my view, this interpretation does not adequately account for Middle Combrian stratigraphy described above.

The classic Cenozoic Himalayan orogeny was the result of collision between the northern passive margin of India (margin A47) and an arc complex to the north. Visean (Early Carboniferous) to Sakmarian (Early Permian) rifting is recorded by thick siliciclastic sequences in extensional basins and by bimodal, alkalic volcanism (Garzanti et al., 1999). Stampfli et al. (1991) interpreted the Indian margin as an upper plate margin. The initiation of seafloor spreading and the rift-drift transition are marked by Artinskian to Kungurian (Early Permian) tholeiitic basalts and by submergence of rift shoulders (Garzanti et al. 1999). For this study, I estimate the rift-drift transition at 271 Ma, the age of the Kungurian-Roadian boundary. The age of collision has been well studied as it ties into such issues as the mechanism for Tibetan Plateau uplift and initiation of the Asian monsoon. The start of collision is marked by the arrival of ophiolitic detritus on Indian margin sediments in the early Eocene (Late Ypresian, 52 Ma) (Rowley, 1996). Thus the lifespan of India’s Phanerozoic Himalayan margin was about 219 m.y.

A48. Aravalli-Dehli orogen, India

The east side of India's Aravalli-Dehli orogen contains the remnants of a Paleoproterozoic passive margin, the Aravalli Supergroup (Banerjee and Bhattacharya, 1994; Deb and Thorpe, 2004). Banerjee and Bhattacharya (1994) interpreted the Aravalli succession in terms of rift, drift, west-directed subduction, and finally, collision between the passive margin and the subduction zone. Rifting of Archean gneissic basement is recorded by the Debari Formation, the oldest formation in the Aravalli Supergroup, which consists of siliciclastic and mafic volcanic rocks (Banerjee and Bhattacharya, 1994). The best age control is provided by Pb-Pb model ages of 2024, 2030, and 2075-2150 Ma from galenas in barite lenses within volcanic units in three widely separated locations (Deb and Thorpe, 2004). The overlying Matoon Formation is a thick succession of shallow-marine platform carbonates, which defines the passive margin itself. Coeval deep-marine facies that appear to be coeval (Roy and Paliwal, 1981) imply that the margin faced an ocean to the west. The top of the Aravalli Supergroup (Udaipur and Sajjangarh Formations) consists of turbiditic sandstones and their metamorphic equivalents (Banerjee and Bhattacharya, 1994). These rocks have been interpreted as collision-related foreland-basin deposits (Banerjee and Bhattacharya, 1994). Pb-Pb model ages on galena from several syngenetic sulfide occurrences average about 1800 Ma (Deb and Thorpe, 2004) and provide a tenuous end date. For present purposes, I place the start date at ca. 2000 Ma and the end date at ca. 1800 Ma, for a lifespan of about 200 m.y.

A49 and A50. South China craton, northwest side, Longmen Shan margin

The northwestern margin of the South China (Yangtse) craton, exposed in the western Sichuan Basin and the Longmen Shan orogen was a passive margin leading up to collision in the Triassic. The origin of this margin is unclear: Burchfiel et al. (1995) favored a Silurian breakup, whereas Jia et al. 2006) suggested that there were two passive margins, one that formed during the late Neoproterozoic, and the second one that formed around the start of the Permian, after re-rifting. I favor the latter scenario.

Late Neoproterozoic (Sinian[2]) strata rest on Mesoproterozoic basement that has yielded ages of 1043-1017 Ma (Yong et al., 2003 and references therein). The older Sinian strata consist of coarse clastic and volcanic rocks deposited in grabens, and are clearly related to regional extension (Burchfiel et al., 1995). The younger Sinian strata blanketed the western part of the craton and consist of clastic rocks, evaporites, and carbonates (Burchfiel et al., 1995). This suggests a transition from extension-driven to thermal subsidence during the late Neoproterozoic. The questions are: did rifting lead all the way to seafloor spreading, and if so, when did this take place? Adopting the model of Jia et al. (2006), I place the first rift-drift transition within the Sinian, ca. 600 Ma., and interpret overlying Cambrian and Ordovician carbonate and siliciclastic strata as part of a passive margin sequence (margin A49). Silurian strata are problematic; Silurian strata are present only in the more distal rocks of the Longmen Shan, where shales and sandstones up to 700 m thick have been reported (Burchfiel et al., 1995; Jia et al., 2006). As mentioned above, Burchfiel et al. (1995) suggested a Silurian breakup date for the margin.

Recent hydrocarbon exploration has provided evidence for a late Paleozoic extensional event along the already established passive margin. Beneath the Longmen Shan orogenic front, seismic reflection profiles show deep grabens that are thought to contain Devonian to Carboniferous strata. The grabens are buried by Permian carbonates (Jia et al., 2006) that are part of a platform that persisted until the Middle Triassic (Jia et al., 2006)[3]. Again following Jia et al. (2006), I interpret the grabens as a record of re-rifting of the passive margin, which caused renewed thermal subsidence. In this scenario, the end date of the older margin (margin A49) and the start date of the younger margin (margin A50) would be the same, ca. 300 Ma.

The demise of the Longmen Shan's younger passive margin is well documented and tightly dated. A foreland basin formed on the passive margin during the Triassic Indosinian orogeny. In what is now the thrust belt, Middle Triassic platform carbonates were uplifted and

[2] The Sinian is roughly equivalent to the Cryogenian plus the Ediacaran, ca. 740 to 542 Ma.

[3] Late Permian flood basalts of the Emeishan mantle plume (260 Ma; Fan et al., 2008) briefly interrupted platformal conditions.

eroded, then unconformably overlain by a fining- and deepening-upward (drowning) succession (Maantang Formation) (Yong et al., 2003). The unconformity is in the right place and at the right time to represent a forebulge (Yong et al., 2003). Whether reckoned at the oldest strata above the unconformity or the carbonate-to-shale transition, the transition from passive margin to foreland basin is near the base of the Carnian (Yong et al., 2003), ca. 228 Ma.

In summary, the older Longmen Shan margin has inferred dates of ca. 600 to ca. 300 Ma for a lifespan of ca. 300 m.y. The younger margin had dates of ca. 300 to 228 Ma for a lifespan of 72 m.y.

A51. Central Orogenic Belt of the North China craton

The North China craton was assembled by collision along the Central Orogenic Belt between the craton's Western Block and Eastern Block (Fig. 5), a collision that may have involved a passive margin. Two conflicting views of its evolution have been pieced together from a complex and fragmentary record. Zhao (2001; also Zhao et al., 2003, 2005) suggested that the *Western Block* had a passive margin that formed before 2500 Ma and collided with the Eastern Block ca. 1850 Ma. On the other hand, Kusky and Li (2003) suggested that the *Eastern Block* had a passive margin that formed ca. 2700 Ma and collided with the Western Block ca. 2500 Ma. They recognized deformed and metamorphosed remnants of a Neoarchean passive margin, a foreland basin, an orogenic wedge, a dismembered mafic-ultramafic complex, and a magmatic arc (Kusky and Li, 2003). Kusky and Li's (2003) model is adopted here, but owing to the controversial tectonic interpretation, complex geology, and imprecise age controls, the quality ranking is a low C. Arguments against Zhao's (2001) alternative tectonic model are presented at the end of this section.

The proposed passive margin sequence of the Eastern Block is preserved in the Qinglong fold-thrust belt (Li and Kusky, 2007). The passive margin is represented by shallow-water deposits of the Banyukou Formation and Wanzi Group. In the Taihang Mountains, the Banyukou Formation consists of >650 m of marble, calc-silicates, banded iron formation, quartzite, and metapelite. Although the rocks have been recumbently folded and metamorphosed to amphibolite facies, the protolith assemblage is consistent with deposition in a passive-margin setting, and incompatible with an active-margin setting (which would be implied by the Zhao, 2001 model). The passive-margin deposits are only indirectly dated but are clearly Neoarchean (Li and Kusky, 2007); they were deposited on ca. 2.7-2.8 Ga gneisses and are overlain by inferred foreland basin deposits that are no younger than ca. 2530 Ma (see below). These broad age brackets are supported by observations from the orogenic belt to the west (Taishan greenstone belt), where komatiites, which were erupted through continental basement, are inferred to record a mantle plume at the time of breakup (Polat et al., 2006). Amphibolites from this greenstone belt yielded a combined Nd isochron age of 2740±70 (Jahn et al., 1988). The age of the rift-drift transition can thus be estimated at ca. 2740 Ma.

The passive-margin deposits are overlapped and flanked on the west by intensely folded and thrust-faulted metasedimentary rocks of the Qinglong basin (Li and Kusky, 2007). The basin fill includes a lower turbiditic succession ("flysch") and an upper sandstone and conglomerate ("molasse") succession. The western boundary of Qinglong basin is an ophiolitic melange. Based on the protolith package, stratigraphic position, and structural setting with respect other tectonic elements, Kusky and Li (2003) and Li and Kusky (2007) interpreted the Qinglong Basin as a foreland basin related to demise of the Neoarchean passive margin. The age of the transition between inferred passive-margin deposits and inferred foreland-basin deposits is best constrained in the Wutai Mountains. Here, the Gaofan Group consists of deep-water turbidites and siliceous rocks that likely represent the deep-water apron of the passive margin. A tuff in the Gaofan Group yielded a U-Pb zircon age of 2528±6 Ma and a gabbro that intrudes the Gaofan Group yielded a U-Pb zircon age of 2523±30 Ma (see Li and Kusky 2007 for original sources). Flysch and molasse facies of the overlying Sizizhuang Formation, which are proximal to the west and distal in the east, are part of the inferred foreland basin (Li and Kusky, 2007). The Sizizhuang Formation was intruded by granite dated at 2549±22 Ma (U-Pb zircon; see Li and Kusky 2007 for original source). On this evidence, I estimate the demise of the passive margin at ca. 2530 Ma—an age that accounts for all the geochronology within error. In this interpretation, the tuff in the Gaofan Group was derived from the approaching arc (see below), and the granite and gabbro were intruded at the start of orogeny.

The colliding arc in Kusky and Li's (2003) model is represented by rocks of the Zunhua and Wutaishan belts, which lie to the west of the foreland-basin succession. The Zunhua belt is characterized by dismembered ultramafic bodies, one of which yielded an Re-Os chromite age of 2547±10 Ma (Kusky et al., 2004). Ultramafic rocks in this belt have geochemical signatures consistent with a suprasubduction zone origin (Polat et al., 2006). The Wutaishan greenstone belt includes granites and comagmatic volcanic rocks that range from 2560 to 2515 Ma (Wilde et al., 2005).

In their alternative model for the Central Orogenic Belt, Zhao (2001) also invoked a collision involving a passive margin, but here the similarities end with the Kusky and Li (2003) model. Zhao (2001) suggested that the passive margin was on the Western Block rather than the Eastern Block. This passive margin existed from an unspecified time in the Neoarchean until ca. 1850 Ma in the late Paleoproterozoic, when the ocean basin between the Western Block and Eastern Block was finally consumed by subduction. Zhao's (2001) tectonic model thus implies a lifespan of >650 m.y. for the passive margin, unsurpassed in Earth history. This model fails to account for regional metamorphism of both the Eastern and Western Blocks—presumably including rocks of the postulated passive margin—at ca. 2.5 Ga. There is no actualistic way to explain a metamorphic event partway through the lifespan of a passive margin. As for the Paleoproterozoic orogenesis that has been emphasized by Zhao (2001) and his coworkers (*e.g.*, Zhao et al., 2003, 2005; Zhao and Kröner, 2007), an alternative view invokes collision along the North Hebei orogen on the northern margin of the North China craton (Kusky et al., 2007). In this new model, the Central Orogenic Belt originated during the Neoarchean and was overprinted in the Paleoproterozoic.

In summary, adopting the Kusky and Li (2003) tectonic scenario, I estimate the start date for the passive margin at ca. 2740 Ma and the end date at ca. 2530 Ma, giving a lifespan of about 210 m.y.

A52 and A53. South China craton, north side, Qinling-Dabie orogen

The suture zone between the South China (Yangtze) and North China (Sino-Korean) cratons has been widely publicized because of its Triassic ultra-high-pressure metamorphic rocks. This orogen shows evidence for two Phanerozoic collisions and two superimposed passive margins on the northern side of the South China craton (Meng and Zhang, 1999).

Initial rifting along the northern side of the South China craton is recorded by a widespread bimodal igneous suite with U-Pb zircon ages ranging from ca. 782 to 746 Ma, the youngest robust age being 756 Ma (Rowley et al., 1997, p. 200). I place the rift-drift transition at ca. 750 Ma. A 5- to 8-km-thick succession of Neoproterozoic (Sinian), Cambrian, and Ordovician strata represent the passive margin (margin A52) (see Rowley et al., 1997 for original Chinese-language citations). The first of two collisions appears to have begun in the Silurian, as evidenced by a 10-km-thick Silurian-Devonian flysch sequence in the South Qinling belt (at that time, still part of the South China craton). Influx of this flysch sequence corresponded to a marked change in provenance: southerly (South China craton) sources during late Neoproterozoic to Ordovician, followed by northerly sources during the Silurian-Devonian (Gao et al., 1995). The transition from passive margin to foreland basin is here placed at 440 Ma.

Meng and Zhang (1999) argued that Devonian rifting and then seafloor spreading in the southern foreland of the Silurian-Devonian collision zone led to breakaway of the South Qinling belt from the South China craton. A breakup unconformity separates Devonian rift-related rocks from Carboniferous-Permian drift-related strata on the new northern edge of the South China craton (Meng and Zhang, 1999); the rift-drift transition is placed at 360 Ma. Demise of this second margin (margin A53) is recorded by development of a "flysch foreland basin" in the middle Triassic (Anisian-Ladinian boundary, rounded to 235 Ma) (Liu et al., 2005). This is slightly older than the age of ultra-high-pressure metamorphism in the Dabie Shan sector of the Qinling-Dabie orogen, which has been dated at 218.4±2.5 Ma (U-Pb zircon; Ames et al., 1996) and involved rocks of the South China craton's northern margin.

In summary, the older passive margin on the South China carton started ca. 750 Ma and ended ca. 440 Ma, giving a nominal lifespan of 310 m.y. The younger margin started ca. 360 Ma and ended ca. 235 Ma, giving a nominal lifespan of 125 m.y.

A54. Nan Ling margin, southeast side of South China craton

The Nanling passive margin on the southeast side of the South China (Yangtze) craton formed during the Neoproterozoic and was destroyed during the Late Ordovician. Neoproterozoic rift-related rocks are widespread across the craton and include granites, mafic-ultramafic plutons, and sedimentary-volcanic rift sequences (Wang and Li, 2003). The youngest pre-drift unit, the Datangpo Formation, has an U-Pb zircon SHRIMP age of 654.5±3.8 (Zhang *et al.,* in press). The rift-drift transition is thought to correspond to the next higher unit in the stratigraphy, the glaciogenic Nantuo Formation. A tuff in the lower Nantuo Formation has a U-Pb zircon SHRIMP age of 636.3±4.9 Ma (Zhang *et al.,* in press), and a tuff at the top of the post-glacial cap dolostone (lower Doushantuo Formation) has a U-Pb zircon TIMS age of 635.2±0.5 Ma (Condon *et al.,* 2005). The rift-drift transition on the southeastern passive margin is placed at ca. 635 Ma. Above the glacial deposits is a widespread platform-carbonate sequence that was deposited across both the South China craton and its southeastern passive margin. Platformal conditions lasted from late Neoproterozoic (upper Sinian System) through Middle Ordovician. The demise of the passive margin in the Ashgillian (ca. 445 Ma) is recorded by platform drowning and, in distal southeasterly sections, by the first influx of southeasterly derived graywacke turbidites. Xu et al., (1997, p. 475) interpreted this as recording the northwestward advance of the Guanxiang orogen. In the area of the inferred orogenic sediment source, Devonian strata rest with angular unconformity on Sinian through Silurian strata, which were deformed during the Guanxiang orogeny (Xu et al., 1997). The duration of the margin was ca. 190 m.y.

A55. Taiwan

The Luzon arc is currently colliding with the passive margin of China at Taiwan. The passive margin is a young one. Clift et al. (2001, p. 500-501) put the rift-drift transition at ca. 28 Ma based on seismic reflection profiles calibrated by nanofossils at ODP Site 1148. A forebulge unconformity at the Miocene-Pliocene boundary (5 Ma) marks the onset of collision (Chou and Yu, 2003). These dates suggest a lifespan of 23 m.y. for the passive margin.

A56, A57, and A58. Eastern margin of Siberia

The eastern (Verkhoyansk) margin of the Siberian craton has been described at length in the main body of the text. Three distinct passive margins were located in about the same place; they are treated as separate passive margins because they formed by completely separate plate-tectonic events. The first margin started at ca. 1600 Ma and ended at ca. 1010 Ma—a very long lifespan of about 590 m.y. The second margin started ca. 650 Ma and ended ca. 380 Ma. The third margin started ca. 380 Ma and ended ca. 160 Ma.

A59. Guaniguanico terrane, Cuba

The Guaniguanico terrane of western Cuba consists of a carbonate-dominated Mesozoic sequence regarded as a displaced passive-margin fragment, originally close to the Yucatán Peninsula of Central America. The rift-drift transition in the Guaniguanico terrane is dated at ca. 159 Ma by Oxfordian tholeiitic basalts (Pszczólkowski, 1999). The drift stage is represented by Upper Jurassic, Lower Cretaceous, and Upper Cretaceous shallow-water carbonates that graded offshore into deeper-water facies (Pszczólkowski, 1999). In Campanian time, distal parts of the margin began to receive volcaniclastic detritus from the encroaching the Greater Antilles arc (Pszczólkowski, 1999), marking the beginning of the end for this margin at ca. 80 Ma. The lifespan was about 79 m.y.

A60. Northern margin of Venezuela

The northern margin of Venezuela formed during the Jurassic as part of the breakup of Pangea. The age of the rift-drift transition is based mainly on a comparison with the Guaniguanico terrane of Cuba (Algar, 1998), which sits next to northern South America on Pindell's (1985) Pangea fit. The start date of the Cuban margin is ca. 159 Ma. In Venezuela, the transition from passive margin to foredeep is marked by a rapid increase in the rate of subsidence at around the Eocene-Oligocene boundary (ca. 34 Ma) (Erikson and Pindell, 1993). This gives a lifespan of about 125 m.y.

A61. Araras margin of Amazonia, Paraguay orogen

The southeast side of the Amazonia craton is interpreted as the site of a Neoproterozoic passive margin that was deformed during the growth of a Neoproterozoic to Cambrian orogenic belt, the Paraguay orogen. No rift-related deposits have been documented. The existence of a passive margin is most clearly recorded by the Ediacaran Araras Formation, an eastward-facing, eastward-deepening platform carbonate succession preserved on the craton and within the orogen (de Alvarenga et al., 2004). The Araras Formation overlies glaciogenic strata—the Puga Formation on the craton and the Cuiabá Group in the fold belt—whose C and Sr isotopes suggest correlation with the Marinoan glaciation (de Alvaregna et al., 2004), which ended at 635 Ma (Condon et al. 2005). The glacial facies show the same west-to-east facies distribution as the carbonates. The passive margin thus appears to have been in existence at least by 635 Ma. I place the rift-drift transition slightly earlier, at ca. 640. A pre-glacial unit of phyllite, quartzite, and marble is also present in the foldbelt (de Alvagenga et al., 2004); these poorly documented strata may also record passive-margin deposition, in which case the start date would be older still. The Araras carbonates are overlain by nearly 3 km of sandstone, mudstone, and siltstone of the Alto Paraguay Group, interpreted to represent a foreland basin. Recently, a second glacial interval—the Serra Azul Formation, has been discovered between the Araras Formation and Alto Paraguay Group (de Alvarenga et al., 2007). It has been correlated with the 580-Ma (Bowring, in Hoffman et al., 2004) Gaskiers glaciation of Newfoundland. An end date of ca. 580 Ma is thus assigned to the Amazonia margin, giving a lifespan of about 60 m.y.

A62. Cuyania terrane, Argentine Precordillera

The Cambrian-Ordovician platformal succession of the Cuyania terrane, Argentine Precordillera (Ramos, 2000) is strikingly similar to that in the Appalachian-Ouachita system (Astini et al., 1995). The oldest platformal strata are Lower Cambrian and their base is taken to approximate the rift-drift transition at ca. 530 Ma. The passive margin subsided until Early Ordovician. Platformal drowning at the onset of arc-passive margin collision began in latest Arenig (ca. 473 Ma), which is only slightly older than in the Appalachians. These dates give the Precordilleran passive margin a relatively brief lifespan of 55 m.y.

A63. Sao Francisco craton, west side, Brasiliano orogen

The Brasiliano Orogeny involved a collision between a Neoproterozoic passive margin on the west side of the Sao Francisco craton, and terranes to the west. The ocean that closed is referred to as the Goianides Ocean. Campos Neto (2000, p. 344) provided a recent review and tectonic model. As noted in the section on the Aracuai-Ribeira margin, extension-related magmatism in the Sao Francisco craton has been dated at 906 Ma. Along the western passive margin of the craton (which presumably was also established during this episode), Campos-Neto (2000, p. 343) placed the rift-drift transition within the undated Paranoá Group. A long-lived carbonate platform (Bambui Group) then developed. An unaltered glaciogenic cap carbonate at the base of the Bambui carbonate platform sequence has yielded a Pb-Pb isochron age of 740±22 Ma (Babinski et al., 2007). I place the rift-drift transition slightly earlier, at ca. 745 Ma. The Bambui platform is overlain by a collision-related foreland-basin succession, the Tres Marias Formation (Campos-Neto 2000, p. 344-345), for which good age control is lacking. Distal portions of the passive margin (Araxá and Andrelandia groups) were metamorphosed at ca. 640 to ca. 630 Ma (Valeriano et al., 2004, p. 53), dating the demise of the margin. Similar age constraints are provided by the youngest detrital zircon grain in the Araxa Group (ca. 643 Ma), which was intruded by tonalite at 638±11 Ma (Piuzana et al., 2003); the narrow age gap suggests that this part of the Araxa Group was deposited in a foredeep during collision. The start- and end dates of the passive margin are placed at ca. 745 and ca. 640 Ma, giving a duration of about 105 m.y.

A64. Sao Francisco craton, east side, Araçuai and Ribeira orogens, Brazil

The Araçuai and Ribeira orogens in Brazil are interpreted as the product of ocean closure and collision involving a passive margin on the east side of the Sao Francisco craton (Pedrosa-Soares et al., 2001; Heilbron and Machado, 2003). I have combined timing constraints from the northerly (Araçuai) and southerly (Ribeira) sectors of the orogen. In the Araçuai belt, mafic magmatism related to rifting of the Sao Francisco craton is as young as 906±2 Ma (Machado, cited in Pedrosa-Soares et al., 2001). Rift- and subsequent passive-margin sedimentation are recorded by the glacially influenced Macaúbas Group (Pedrosa-Soares et al., 2001, p. 310). I place the rift-drift transition at 900 Ma. Within the ocean to the east of the Sao Francisco craton, an arc (Oriental terrane) has yielded U-Pb zircon ages ranging from 790 to 620 Ma (Heilbron and Machado, 2003). Collision between the Sao Francisco craton and the arc began ca. 590 Ma, judging from dates on the oldest syntectonic plutons in the orogenic wedge (Heilbron and Machado, 2003, their Fig. 13). Using these numbers, the Araçuai-Ribeira margin would appear to have lasted about 310 m.y.

A65. Sao Francisco craton, south side, Transamazonian orogen, Brazil

In the southern part of the Sao Francisco craton of Brazil, the Minas Supergroup is interpreted to represent a passive margin to foreland-basin sequence that was caught up in the Transamazonian orogeny. The following is from Teixeira et al. (2000, p. 124) and references therein. Basement rocks are as young as 2612+3/-2 Ma. Rift deposits have not been identified. The inferred passive margin deposits (Caraca, Itabira, and most of the Piracicaba Groups) include siliciliclastics rocks, banded iron formation, and carbonate rocks. The youngest detrital zircon in the Caraca Group is 2580±7 (Hartman et al., 2006). Carbonates of the Itabira Group yielded a Pb/Pb whole-rock isochron age of 2420±19 Ma, which must be somewhat younger than the rift-drift transition, which I place, splitting the difference, at ca. 2500 Ma. The Sabará Formation, at the top of the Piracicaba Group, is regarded as flysch that was deposited in a Transamazonian foreland basin (Machado et al. 1996), or intrarc basin (Hartmann et al., 2006). The youngest detrital zircon ages in the Sabará Formation (2125±4 Ma) (Machado et al. 1996) are virtually identical to the zircon age of a tonalite pluton within the orogen (2124±1 Ma) (Noce et al., 1998). Orogeny was underway by this time. Thus, the end date of the margin was probably ca. 2130 Ma. These start and end dates suggest a lifespan of 370 m.y.

A66. Sierra de la Ventana, Argentina (a) and Cape Fold Belt, South Africa (b), and Ellsworth Mountains, Antarctica (c)

On a restored fit of the South Atlantic continents, the late Paleozoic Sierra de la Ventana orogen and Cape Fold Belt line up as parts of du Toit's (1937) Samfrau geosyncline. Ordovician to Carboniferous strata in the two belts, and in a third belt in Antarctica, were deposited along what was originally a single passive margin, and are discussed here under the same heading. In Argentina, Cambrian rift-related igneous rocks range from 531±4 to 509±5 Ma (Rapela et al., 2003). An overlapping sequence of Ordovician to Devonian marine siliciclastic strata represents the thermal subsidence phase of the inferred passive margin (Rapela et al., 2003). Precise age control is lacking but the rift-drift transition is probably near the Cambrian-Ordovician boundary. Demise of the Argentinian sector of the passive margin corresponds to the onset of Late Carboniferous to Permian foreland-basin deposition in the Sauce Grande Basin (beginning at ca. 295 Ma; Lopez-Gamundi, 2006).

Correlative passive-margin deposits of similar age and facies are also known from the Cape Fold Belt of South Africa. Here, alluvial fan facies of the Kansa Subgroup are interpreted as rift deposits; an age cluster of detrital zircons at 518±9 Ma establishes a Midle Cambrian or younger depositional age (Barnett et al., 1997). The overlying Table Mountain Group represents the thermal subsidence phase of a siliciclastic-dominated passive margin, from Early Ordovician to mid-Carboniferous time (Shone and Booth, 2005). The overlying Karoo Group is widely interpreted as foreland-basin succession, which began with the Dwyka tillites at ca. 300 Ma (Catuneanu et al., 2005).

A third segment of the same early Paleozoic passive margin is known from the Ellsworth Mountains of Antarctica. Here rifting is represented by the Lower? and Middle Cambrian Hertitage Group, which includes volcanic rocks that have extensional geochemical signatures (Curtis et al., 1999). Next came a silicilclastic-dominated passive margin, represented by the Upper Cambrian to Devonian Crashsite Group (Curtis and Lomas, 1999). The passive margin is overlain by foreland-basin succession that begins with the 1-km thick, glaciogenic Whiteout Conglomerate of Permo-Carboniferous age (Matsch and Ojakangas, 1992). Age controls are imprecise.

Taking the South American, South African, and Antarctic segments together, the rift-drift transition can be placed at ca. 500 Ma and the passive margin to foreland basin transition at ca. 300 Ma, giving a lifespan of about 200 m.y.

A67. Northern margin of West African craton, Anti Atlas, Morocco

The West African craton was surrounded by passive margins that formed in the Neoproterozoic and were destroyed in the late Neoproterozoic to Cambrian. Along the northeastern margin of the craton in the Anti-Atlas of Morocco, the passive margin succession is assigned to the Taghdout Group (Thomas et al., 2002). Basalts in the lower Taghdout Group, and affiliated mafic intrusions of the Ifzwane Suite, are related to rifting (Thomas et al., 2002, p. 221) and are ca. 800 Ma. Development of an ocean basin to the north is inferred from ophiolitic and island-arc fragments later caught up in the Anti-Atlas collision (Thomas et al., 2002, p. 225). Specifically, plagiogranite within the Tasriwine supra-subduction ophiolite has a U-Pb zircon age of 762±1 Ma (Samson et al., 2004), showing that an ocean existed north of the West African craton by this time. Collision between the passive margin and an arc (the Saghro arc of Saquaque et al. 1989) is inferred to have followed an interval of north-dipping subduction (Hefferan et al., 2000). A belt of melange, ophiolitic fragments, and blueschist marks the subduction zone (Hefferan et al., 2002). The demise of the passive margin is dated by the post-kinematic Bleida granodiorite, which cuts collision-related fabrics and is dated at 579.4±1.2 Ma (Inglis et al., 2004). For this study, I place the start date of the passive margin at ca. 800 Ma and the end date at ca. 590 Ma; these ages imply a lifespan for the passive margin of about 210 m.y.

A68. Western margin of West African craton, Mauritanide orogen

In contrast to the northern and eastern margins of the Neoproterozoic West African craton, the coeval western margin is not universally regarded as having been a passive margin. The evolution of this edge of the craton is recorded in sectors known as the Mauritanide, Bassaride, and Rokelide orogens, from north to south. According to one view, the west side of the Neoproterozoic West African craton faced into a rift that never proceeded as far as seafloor spreading. This has been suggested for the Rokelides by Culver et al. (1987) and for the Bassarides and Mauritanides by Lécorché et al. (1989). Alternatively, the existence of ocean floor and thus a passive margin *sensu stricto* was favored for the Bassarides by Ponsard et al. (1988), and for the Mauritanides by Villeneuve and Cornée (1994). The latter interpretation is adopted here.

The Mauritanide margin of the West African craton is severely telescoped and tectonized, having been buried and shuffled by thrusts during three orogenies. These are dated as follows: Pan-African I—ca. 650 Ma; Pan-African II—ca. 575-550 Ma, and Hercynian—ca. 300 Ma (Lécorché et al., 1989). Three unhappy consequences of the polyorogenic history are that the deposits of the passive margin phase are now tectonites, that the original paleogeographic relations and affinities of rock units are obscure, and that no foreland-basin deposits can be unequivocally related to the demise of the Neoproterozoic passive margin during the Pan-African I event.

In the Mauritanides, rifting is dated by the parautochthonous Bou Naga alkalic complex, which, along with Archean basement rocks of the West African craton, occurs in a window beneath Mauritanide thrusts. The youngest age from the alkalic complex is 676±8 Ma (Pitfield et al., 2004, p. 307 and references therein), and accordingly, I place the rift-drift transition at ca. 675 Ma.

The end date is constrained by geochronology from within the orogen. The Pan African I is believed to represent emplacement of a Neoproterozoic arc over the margin of the West African craton (*e.g.*, Villeneuve and Cornée, 1994). This orogeny is best expressed in the Bassaride portion of the belt where sedimentary rocks were metamorphosed at 660 to 650 Ma and are overlain unconformably by younger Neoproterozoic tillites (Lécorché et al., 1989). I place the end date at ca. 650 Ma, implying a very short lifespan of 25 m.y.

Calc-alkaline plutonic rocks in the Mauritanides that likely represent the colliding arc include ca. 665-Ma gabbros and granodiorites of the Gorgol Noir Complex; metavolcanic and metasedimentary rocks of arc affinity include the Mbout Supergroup (Pitfield et al., 2004, p. 506). Virtual overlap between the inferred ages of rifting and arc magmatism suggest that the Mauritanide passive margin faced a short-lived back-arc basin, whose closure resulted in the event known as Pan African I.

A69. Eastern margin of the West African craton, Dahomeyide and Gourma orogens

The eastern margin of the West African craton was a passive margin that formed and was destroyed within the Neoproterozoic, its demise brought on by collision with an arc (Caby et al., 1981). Evidence for the evolution and timing of this margin comes from the Pan-African Dahomeyide orogen and Volta basin in Nigeria, and from the Pan-African Gourma orogen farther to the north in Mali. As reviewed by Affaton et al. (1991), the Volta Basin is an eastward-thickening prism consisting of three supergroups representing successive tectonic environments. At the base, the 1-km-thick Boumbouaka Supergroup (ca. 1100 to ca. 700 Ma) is regarded either as the fill of an epicontinental cratonic basin (Affaton et al., 1991) or as rift-related (Drouet, 1997, p. 11). The overlying Penjari (or Oti) Supergroup, about 2.5-km thick, includes tillites, various other siliciclastic rocks, and carbonates in its lower part, and is interpreted as a passive-margin succession (Affaton et al., 1991; Bertrand-Sarfati et al., 1991). The onset of passive margin deposition is poorly dated but estimated at ca. 700 Ma by Drouet (1997, p. 11). A somewhat older, indirect age constraint is provided by the 726+7/-3 Ma Techalré pluton in the Tilemsi magmatic arc in Mali (Caby et al., 1994). This arc is thought to have formed above the subduction zone where ocean floor attached to the West African craton was consumed (Caby et al., 1994). This finding is consistent with a paleogeographic reconstruction at 750 Ma by Villeneuve and Cornée (1994) that shows an established passive margin by this time. On this slim evidence I place the rift-drift transition at ca. 775 Ma.

The youngest Neoproterozoic strata in the Volta Basin, assigned to the Tamale Supergroup (about 500-m thick), include outboard-derived molassic sandstones and have been interpreted as the fill of a Pan-African foreland basin (Affaton et al., 1991; Bertrand-Sarfati et al., 1991). The demise of the margin is not tightly dated on stratigraphic grounds. However, new Rb-Sr, Sm-Nd, and $^{40}Ar/^{39}Ar$ dates on high-pressure metamorphic rocks within the Gourma orogen show that the continental margin of the West African craton had entered a subduction zone by ca. 625 Ma (Jahn et al., 2001). The tightest age constraint comes from a $^{40}Ar/^{39}Ar$ plateau age of 623±3 Ma (Jahn et al., 2001). I place the demise of the margin at ca. 625 Ma. This implies a lifespan of about 150 m.y., although a considerably longer duration seems possible.

A70 and A71. Western and southern margins of the LATEA craton, Hoggar, Africa

The Central Hoggar is made up of four terranes that came together during the Paleoproterozoic and have been grouped by Liegeois et al. (2003) under the acronym "LATEA"; the following is from their paper. During the earlier Neoproterozoic, the western margin of LATEA (margin A70) has been interpreted as a passive margin that was overridden by a magmatic arc, the Iskel terrane. Early events in the evolution of this hypothesized margin have not been documented, and the start date is unconstrained. The end date is about 870 Ma based on the 870-850 Ma age range of syn-kinematic to post-kinematic plutons in the arc terrane. This margin has a quality ranking of D.

A second passive margin (margin A71) has also been hypothesized for the south side of the Central Hoggar (Liegeois et al., 2003). As above, early events are unconstrained and a start date cannot be set. The demise of the margin at ca. 685 Ma was marked by thrust emplacement of ophiolites and eclogites. This margin also has a quality ranking of D.

A72. West Gondwana margin, East African orogen

East and West Gondwana collided in the Neoproterozoic to form the East African orogen. As reviewed by Stern (1994), deformed and metamorphosed remnants of inferred passive margin deposits have been identified along the West Gondwana margin in Sudan in the north and Kenya in the south. In the Keraf zone of Sudan, polydeformed slope- and basin-facies carbonate rocks are assigned to this passive margin. Their C- and Sr-isotopic signatures suggest a depositional age of ca. 750 Ma (Stern, 1994), and they were metamorphosed to granulite facies at about 720 Ma (Stern, 1994). Metamorphism was the consequence of collision of an island arc system preserved in eastern Sudan (Haya terrane) and across the Red Sea in Saudi Arabia (Asir terrane) (Kroner et al., 1991). Stern (1994) argued that the age of breakup of the passive margin is approximated by U-Pb ages of 840-870 Ma from the oldest arc rocks, on the assumption that it is the arc that broke away from Sudan to form the passive margin. For present purposes, I assign a start date of about 840 Ma and an end date of about 730 for the passive margin, for a nominal duration of 110 m.y.

A73. Kaoko Belt (Northern Coastal Branch of Damara orogen), Africa

The Kaoko Belt marks the former site of the Adamastor Ocean, which lay between the Congo and Rio de la Plata cratons. The western margin of the Congo craton has been interpreted as a passive margin (Otavi platform) that met its end during the Pan-African Damaride orogeny. Timing constraints are not as good as for the southern margin of the Congo craton (margin A74). The age of the rift-drift transition perhaps corresponds to the upward transition, at ca. 780 Ma (Halverson et al., 2005), from siliciclastic rocks of the Nosib Group to carbonate rocks of the Ombombo Subgroup of the Otavi Group (P. Hoffman, 2008, written communication). In the adjacent Kaoko orogen, synkinematic orthogneisses dated at 580-576 Ma (Goscombe et al., 2003) show that collision was underway by this time. Superposed folds at the junction of the Kaoko Belt and Inland Branch of the Damara orogen confirm that the Kaoko orogen is older (Maloof, 2000). A start date of the passive margin at ca. 780 and an end date at ca. 580 Ma yield a lifespan of about 200 m.y.

The Neoproterozoic western passive margin of the Congo craton is also exposed farther north, from Angola to Gabon (Tack et al., 2001). Rift-related magmatism has been dated at 999 to 912 Ma (Tack et al., 2001). Strata of the overlying West Congolian Group are interpreted as the deposits of a passive margin platform (lower part) and a westerly-derived Pan African foreland basin (upper part) (Tack et al., 2001). Except the rift-related igneous rocks, there are no tight age controls.

A74. Southern margin of Congo craton, Inland Branch of Damara orogen, Africa

The Inland Branch of the Damara orogen marks the suture between the Kalahari craton to the south and the Congo craton to the north; the Neoproterozoic ocean that is inferred to have opened and then closed between the same two cratons is called the Khomas Sea (Stanistreet et al., 1991). In Namibia, the southern margin of the Congo craton was the site of the south-facing Otavi passive margin, which bends to the north near the Atlantic coast and becomes the Kaoko margin (A73). The geometry and geochronology of rift- and drift-related deposits of the Congo margin were recently summarized by Hoffman et al. (2007). Early rifting is dated at 759±1 and 746±2 Ma (Hoffman et al., 2007). The rift-drift transition is placed between the Gruis and Ombaatjie Formations; the age of this boundary has been estimated at ca. 670 Ma (Halverson et al., 2005). The passive margin itself was the site of a long-lived, 3-km-thick carbonate platform: the Tsumeb Subgroup of the Otavi Group (Hoffman et al., 2007). The passive margin met its end in the very late Neoproterozoic with collision between the Congo craton and a short-lived arc, now located in the Central Damara Zone; south of this arc lay the Kalahari craton with its own passive margin (see below, margin A75). Arc-related magmatism has been dated at 558±5 Ma (de Kock et al., 2000), and syntectonic, probably syncollisional granites have been dated at 549±11 Ma (Johnson et al., 2006). The end date for the passive margin can therefore be set at about 555 Ma for a duration of about 115 m.y.

Farther east along the south side of the Congo craton, the Neoproterozoic history of the Katangan sector of the passive margin is broadly compatible with that just discussed (Wendorff, 2005), but age constraints and tectonic interpretations are not as robust. For this synthesis, I therefore rely on the Namibian evidence.

A75. Northern margin of Kalahari craton, Inland Branch of Damara orogen, Africa

The other side of the Khomas Sea has also been interpreted as a passive margin, conjugate to the southern margin of the Congo craton (margin A74), described immediately above. This passive margin bends to the south near the Atlantic coast and becomes the Gariep margin (A76). Borrowing age control from the Congo margin in Namibia, the rift-drift transition is placed at ca. 670 Ma. In the Gogabis area, the most easterly outcrops in Namibia, the passive-margin succession is represented by the Witvlei Group carbonate platform. Paleogeographic maps show that the oldest northerly-derived strata on the Kalahari craton are in the Schwarzrand Subgroup of the Nama Group (Germs, 1983, p. 101). The oldest of four ash beds in the Schwarzrand Subgroup yielded a U-Pb zircon age of 545 Ma (Grotzinger et al., 1995). I place the start date of the passive margin at 670 Ma and the end date at 550 Ma, giving a lifespan of 120 m.y.

A76. Gariep Belt (Southern Coastal Branch of Damara orogen), Africa

Rocks of the Gariep Belt mark the site of closure of the southern arm of the Adamastor Ocean, between Africa's Kalahari craton on the east and South America's Rio de la Plata craton on the west. Rifting is recorded by the Rosh Pinah Formation, which consists of lacustrine and alluvial fan facies and bimodal volcanic rocks (Jasper et al., 1995). The volcanic rocks are dated at 741 ± 6 (Frimmel et al., 1996). An overlying transgressive sequence of shallow-water carbonates and siliciclastics is interpreted to record post-rift, thermal subsidence of the passive margin (Germs, 1983). The highest of several ash beds from near the top of the Schwarzrand Subgroup (and not far below foreland-basin siliciclastics of the Fish River Subgroup) has been dated at 539 Ma (U-Pb zircon; Grotzinger et al., 1995). A platform-wide erosional unconformity immediately below the dated ash may record forebulge uplift. I place the start date of the passive margin at 735 Ma and the end date 535 m.y., giving a lifespan of about 200 m.y.

A77. Western margin of Kaapvaal craton, South Africa

The southwestern passive margin of the Kaapvaal craton has been described at length in Section 4.4. The start date is ca. 2640 Ma and the end date is ca. 2470 Ma, giving a lifespan of about 170 m.y.

A78. Belingwe margin, Zimbabwe craton

The Zimbabwe craton, one of the world's oldest continental nuclei, is flanked to the south by the Belingwe greenstone belt. A recent review by Hofmann and Kusky (2004) captures the current state of knowledge and offers a well-reasoned interpretation of the stratigraphy and tectonics. Unfortunately, age control is still not adequate and shear zones complicate what was once regarded as a relatively straightforward stratigraphic succession (Kusky and Winsky, 1995). The Belingwe greenstone belt includes an older (ca. 2.9-2.8 Ga) greenstone succession and a younger one (ca. 2.6 Ga) that includes strata of a postulated passive margin, called the Ngezi Group. The lowest unit in the Ngezi Group, the Manjeri Formation, rests on Zimbabwe craton basement that is as old as 3.5 Ga. The Manjeri Formation consists of up to 250 m of fluvial to shallow marine conglomerate, sandstone, siltstone, banded iron formation, and limestone. A Pb-Pb isochron age of 2607±49 Ma provides some control for the age of a stromatolitic limestone from the lower part of the Manjeri Formation. In the upper part of the Manjeri Formation, rapid changes in facies and thickness suggest a rifting environment (Hofmann and Kusky, 2004). The Manjeri Formation is overlain by a volcanic-dominated succession assigned to the Reliance and Zeederbergs Formations, consisting of nearly 4 km of basalt, komatiite, minor andesite, and some sedimentary rocks. A komatiite yielded a whole-rock Pb-Pb age of 2692±9 Ma. The uppermost strata in the Ngezi Group belong to the ~1.3-km-thick Cheshire Formation, which includes a lower carbonate member and an upper siliciclastic member. The carbonates were deposited in an eastward-deepening ramp setting and are capped by a karst horizon. The siliciclastic section (deep-water conglomerate and shale) records platform drowning. As shown by Hofmann et al. (2001), the Cheshire Formation has all the earmarks of a passive margin to foreland basin transition. The rift-drift transition would appear to correspond to the base of the Cheshire Formation and the passive margin to foredeep transition to the mid-Cheshire Formation influx of siliciclastics. Unfortunately, the only direct age control on the Cheshire Formation is a Pb-Pb isochron age of 2601±49 Ma on stromatolitic limestone (Bolhar et al., 2002). This is indistinguishable from the age of the stromatolitic limestone in Manjeri Formation, *below* the interpreted rift sequence. The duration of this proposed Late Archean passive margin cannot be estimated from available evidence, except to suggest that it existed around 2600 Ma.

A79. Northwestern margin of Australia at Timor

The ongoing collision between the northwestern passive margin of Australia and the Banda Arc was described in Section 4.1. The start date is ca. 151 Ma and the end date is about 4 Ma, for a lifespan of about 147 m.y.

A80. Northern margin of Australia in New Guinea

The northeastern passive margin the Australian continent is in the late stages of arc-collision in New Guinea. The rifting history is complex. Pigram and Symonds (1991, their Fig. 2) assigned the post-breakup unconformity a Bajocian age in western New Guinea and a Pliensbachian-Sinemurian age in eastern New Guinea. I place the age in the middle at ca. 180 Ma. Platform drowning at onset of collision was diachronous; the drowning sequence began in the Late Oligocene, ca. 26 Ma (Pigram et al., 1989). These ages yield a lifespan of about 154 m.y. for the passive margin.

A81. Halls Creek orogen, Western Australia

The Halls Creek orogen of northwestern Australia has been interpreted in terms of collision between a passive margin on the west side of the North Australian craton that collided with a magmatic arc (Tyler et al., 2001). The inferred passive margin succession is the Halls Creek Group of the Eastern Zone of Blake et al. (2000). The Saunders Creek Formation, at the base of the Group, consists of quartz-rich sandstone and conglomerate. The overlying Biscay Formation features mafic and lesser felsic volcanics, siliciclastics, carbonates, and chert. From these lithologies, the Biscay Formation might be interpreted to include both rift- and passive-margin facies. A felsic volcanic rock at 1880±3 Ma (Blake et al., 2000) is probably close to the age of the rift-drift transition, although Tyler et al. (2001) put this event 30 m.y. earlier. The overlying Olympio Formation consists of mainly of turbidites, interbedded with alkaline pillow lavas dated at ca. 1857 and ca. 1848 Ma, and intruded, prior to deformation, by mafic sills (Blake et al., 2000). The main folding of these units took place during the Halls Creek orogeny at ca. 1835 Ma (Blake et al., 2000). This implies a foreland-basin setting for the Olympio Formation, and an end date of about 1860 Ma for the passive margin, yielding a lifespan of perhaps 20 m.y. In this model, the Olympio volcanic rocks and sills would also have been emplaced in a foredeep.

A82. Rudall Complex, Western Australia

The Rudall Complex of Western Australia preserves the deformed and metamorphosed remnants of a Paleoproterozoic passive margin. Two principal terranes make up the Rudall Complex: the Talbot and Connaughton terranes (Smithies and Bagasa, 1997; Betts and Giles, 2006). The Connaughton terrane consists of mafic gneiss and schist derived from tholeiitic basalts, and paragneiss derived from chemical and clastic sedimentary rocks (Bagas, 2004); this assemblage has been interpreted as the pre-collisional eastern passive margin of the Plbara craton (Smithies and Bagasa, 1997, their Fig. 7). Metamorphosed quartzites and turbidites of the Talbot terrane deepened and thickened to the east and are interpreted as a foreland basin succession related to the Yapungku orogeny, when the passive margin collided with an arc (Smithies and Bagas, 1997, their Fig. 7). The start date for the passive margin is poorly constrained but likely was older than 2000 Ma (Smithies and Bagas, 1997; Betts and Giles, 2006). The end date is probably slightly before 1777±7 Ma, the age of the oldest syntectonic intrusions that intrude the inferred passive-margin deposits (Bagas, 2004); I set the end date at ca. 1780 Ma. A lifespan cannot be determined but was likely greater than 220 m.y. The quality ranking is D.

A83. Southern margin of Pilbara craton, Australia

Australia's Pilbara craton is among the world's oldest continental nuclei. Its southwestern margin, the Hamersley Basin (or McGrath Trough; Martin et al., 2000), has been interpreted as a passive margin that spanned the Archean-Proterozoic boundary. The 7-km-thick Fortescue Group, mostly mafic lavas plus some siliciclastic rocks, has been interpreted as the product of two successive, differently oriented rifting events, the second one having led to seafloor spreading to the southwest of the craton (Blake and Barley, 1992). The oldest rift-related volcanic rocks are ca. 2775 Ma (Arndt et al., 1991). Blake and Barley (1992) placed the rift-drift transition at the contact between the youngest basalts (Bunjinah Formation) and overlying mudstones (Jeerinah Formation, the youngest unit in the Fortescue Group). The rift-drift transition is at ca. 2685 Ma, based on a zircon age of 2684±6 Ma from a tuff near the base of the Jeerinah Formation (Arndt et al., 1991). Most of the passive-margin succession belongs to the Hamersley Group and consists of iron formations, limestones, and mudstones (Blake and Barley, 1992). The lower part of the Hamersley Group is dated by tuffs at 2629, 2597, and 2561 Ma (age constraints reviewed by Martin et al., 1998). Blake and Barley (1992) noted that the upper part of the Hamersley is condensed and they suggested that this interval represents a distal arc collision. Alternatively, I suggest that the condensed interval can be explained as a consequence of the exponential decay of the subsidence rate, such that little new accommodation space was created in the later part of the passive margin's lifespan. Intercalated between iron formations, just below the top of the Hamersley Group, is a rhyolite sequence (Woongarra Volcanics) dated at 2449 Ma. The youngest unit in the Hamersley Group, the Boolgeeda Iron Formation, has been interpreted as having been deposited on a forebulge during initial thrust loading of the margin (Martin et al., 2000). If this is correct, then the Woongarra Volcanics would also be a distal product of collision. The Boolgeeda Iron Formation grades upward into siliciclastic turbidites (Kungarra Formation of the Turee Creek Group), interpreted as a foreland-basin succession related to the Ophthalmian orogen to the south (Martin et al., 2000). The passive-margin to foreland basin transition can be tentatively placed at the base of the Boolgeeda Iron Formation, perhaps a few million years after 2449 Ma; I place it at ca. 2445 Ma. The ages selected here imply a lifespan of about 240 m.y. for this passive margin, which is longer than any entirely Phanerozoic margin in the present compilation.

A84. Northeast margin of the Gawler craton, Kimban orogen, southern Australia

The Kimban orogen of southern Australia records collision between the Gawler and North Australian cratons in the late Paleoproterozoic, with an inferred passive margin on the northeastern margin of the Gawler craton (Betts and Giles, 2006, their Fig. 10). The passive margin is represented by clastic and chemical metasedimentary rocks of the Hutchinson Group (Betts and Giles, 2006, p. 99), deposited between 1900 and 1850 Ma (Swain et al., 2005). The start date for the margin can be placed at ca. 1900. The Kimban Orogeny spanned ca. 1740-1690 Ma (Betts and Giles, 2006, p. 99); the end date for the margin is therefore placed at ca. 1740 Ma. These ages imply a lifespan of 160 m.y.

A85. Tasman orogen, Tasmania and Australia

The eastern margin of Australia appears to be the first Precambrian passive margin to have been identified as such (Sprigg, 1952). After the initial stint as a passive margin, eastern Australia had a long and complex history involving several Paleozoic arc collisions. In Tasmania, evidence suggests that a relatively short-lived passive margin lay along the eastern margin of Australia in the late Neoproterozoic and was destroyed in the Cambrian (Crawford and Berry, 1992). Rifting of Proterozoic continental basement produced rift basins along with basalts transitional to MORB. Emplacement of the Rocky Cape dike swarm is dated by K/Ar at 590±8 and this age is taken here as approximating the rift-drift transition. Somewhere to the east, an intra-oceanic arc formed above an east-dipping subduction zone, leading inevitably to collision. The timing of collision is bracketed by a 520 Ma zircon age on an arc-like tonalite (which presumably pre-dated collision) and the Middle Cambrian to early Late Cambrian age of the post-collisional sedimentary rocks of the Tyndall and Denison Groups (Crawford and Berry, 1992). These dates give a duration of about 70 m.y. for the passive margin in Tasmania.

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