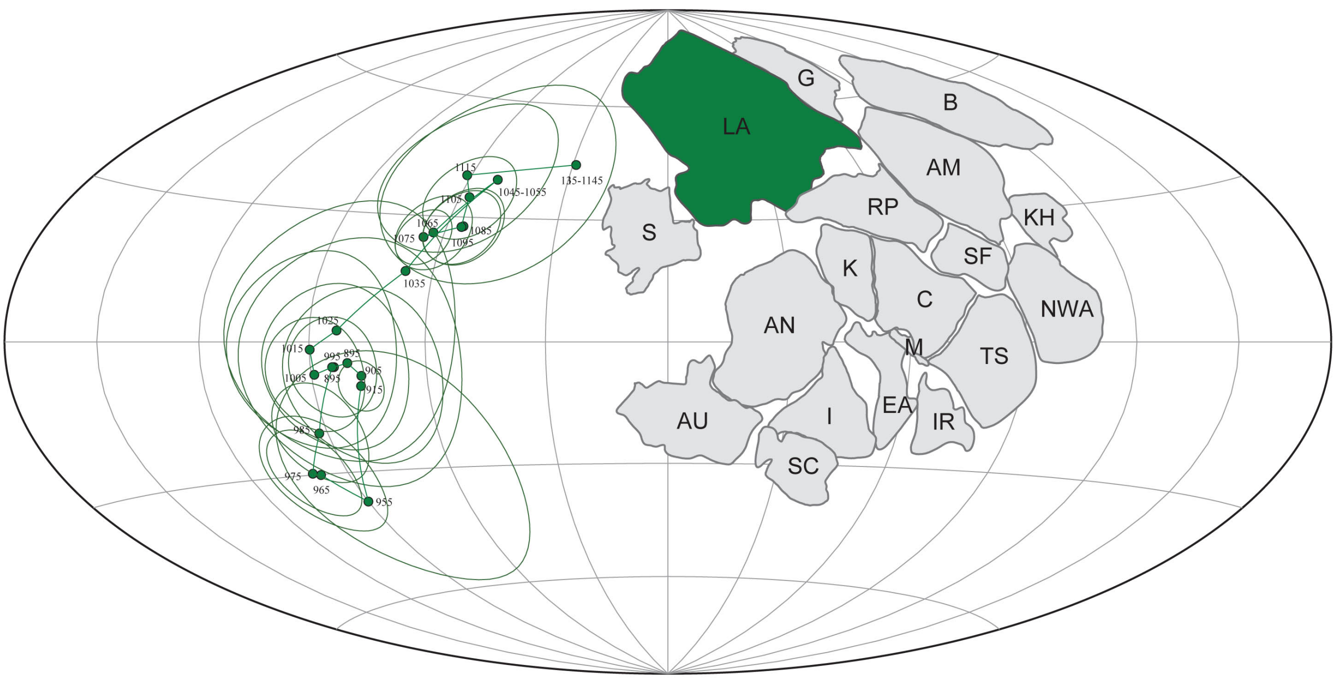
Reconstructing Rodinia by Fitting Neoproterozoic Continental Margins

By John H. Stewart



1,200-850 Ma

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Cover: Paleomagnetic poles for Laurentia (green) for the time period 1,200-850 Ma, and an averaged apparent polar wander path (APWP) derived from these poles

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John H. Stewart

U.S. Geological Survey, Menlo Park, CA 94025

Abstract

Reconstructions of Phanerozoic tectonic plates can be closely constrained by lithologic correlations across conjugate margins by paleontologic information, by correlation of orogenic belts, by paleomagnetic location of continents, and by ocean floor magmatic stripes. In contrast, Proterozoic reconstructions are hindered by the lack of some of these tools or the lack of their precision. To overcome some of these difficulties, this report focuses on a different method of reconstruction, namely the use of the shape of continents to assemble the supercontinent of Rodinia, much like a jigsaw puzzle. Compared to the vast amount of information available for Phanerozoic systems, such a limited approach for Proterozoic rocks, may seem suspect. However, using the assembly of the southern continents (South America, Africa, India, Arabia, Antarctica, and Australia) as an example, a very tight fit of the continents is apparent and illustrates the power of the jigsaw puzzle method.

This report focuses on Neoproterozoic rocks, which are shown on two new detailed geologic maps that constitute the backbone of the study. The report also describes the Neoproterozoic, but younger or older rocks are not discussed or not discussed in detail.

The Neoproterozoic continents and continental margins are identified based on the distribution of continental-margin sedimentary and magmatic rocks that define the break-up margins of Rodinia. These Neoproterozoic continental exposures, as well as critical Neo- and Meso-Neoproterozoic tectonic features shown on the two new map compilations, are used to reconstruct the Mesoproterozoic supercontinent of Rodinia. This approach differs from the common approach of using fold belts to define structural features deemed important in the Rodinian reconstruction. Fold belts are difficult to date, and many are significantly younger than the time frame considered here (1,200 to 850 Ma).

Identifying Neoproterozoic continental margins, which are primarily extensional in origin, supports recognition of the Neoproterozoic fragmentation pattern of Rodinia and outlines the major continental masses that, prior to the breakup, formed the supercontinent. Using this pattern, Rodinia can be assembled by fitting the pieces together.

Evidence for Neoproterozoic margins is fragmentary. The most apparent margins are marked by miogeoclinal deposits (passive-margin deposits). The margins can also be outlined by the distribution of continental-margin magmatic-arc rocks, by juvenile ocean-floor rocks, or by the presence of continent-ward extending aulacogens.

Most of the continental margins described here are Neoproterozoic, and some had an older history suggesting that they were major, long-lived lithospheric flaws. In

particular, the western margin of North America appears to have existed for at least 1,470 Ma and to have been reactivated many times in the Neoproterozoic and Phanerozoic. The inheritance of trends from the Mesoproterozoic by the Neoproterozoic is particularly evident along the eastern United States, where a similarity of Mesoproterozoic (Grenville) and Neoproterozoic trends, as well as Paleozoic or Mesozoic trends, is evident.

The model of Rodinia presented here is based on both geologic and paleomagnetic information. Geologic evidence is based on the distribution and shape of Neoproterozoic continents and on assembling these continents so as to match the shape, history, and scale of adjoining margins. The proposed model places the Laurasian continents—Baltica, Greenland, and Laurentia—west of the South American continents (Amazonia, Rio de La Plata, and Saõ Francisco). This assembly is indicated by conjugate pairs of Grenville-age rocks on the east side of Laurentia and on the west side of South America. In the model, predominantly late Neoproterozoic magmatic-arc rocks follow the trend of the Grenville rocks. The boundary between South America and Africa is interpreted as the site of a Wilson cycle, in which Rodinia fragmented in the Neoproterozoic, forming an ocean that then closed in the late Neoproterozoic. Although many have proposed a similar model for East Gondwana, the interpretation presented here suggests that the East Gondwana continents were previously assembled at least as early as the Mesoproterozoic.

The validity of the model is tested by drawing upon paleomagnetic data. Paleomagnetic poles from the continents of Amazonia, Baltica, Congo, Kalahari, Siberia, and possibly Australia (the main components of the model) are compatible with the reconstruction.

Introduction

Concepts concerning the existence of a major Mesoproterozoic supercontinent have been debated since the early 1980s. Some models show many relatively small plates dispersed over large parts of the globe (Cordani and others, 2003a; Lu and others, 2008b; Meert and Torsvik, 2003; Pesonen and others, 2003), whereas other models show a grouping of major continents into a loosely or tightly fitting assemblage forming a supercontinent (Bond and others, 1984; Condie, 1997, 2003; Dalziel, 1991, 1997; Hoffman, 1991; Li and others, 1995; Piper, 2000, 2004; Pisarevsky and others, 2003; Stewart and Glen, 2005; Waggoner, 1999). The name "Rodinia" was proposed by McMenamin and McMenamin (1990) to describe this supercontinent, and they suggested a specific grouping of continents in their proposal. Subsequently, the term "Rodinia" has been used by various authors for a wide range of different configurations. McMenamin and McMenamin (1990) define Rodinia as resulting from a "major, one-billion-year-old episode of continental collision and supercontinent formation," and the name is used in the same sense here. Such a definition is in agreement with the usage of many geologists, although such a widespread practice leads to conferring the same name on quite different configurations of the proposed supercontinent.

The purpose of this report is to identify Neoproterozoic continents using the distribution of sedimentary rocks, as well as continental-margin magmatic rocks, to define the margins of Rodinia. This approach differs from the commonly used method of using fold belts and associated other sedimentary and igneous features to define the important structural features of the Rodinian crustal fragments. Fold belts are difficult to date, and many are significantly younger than the Neoproterozoic time frame considered here. The distribution of sedimentary rocks is considered to be a more reliable indicator of major structural features, namely the structures that outline continents. This information is further used to propose a new Proterozoic reconstruction.

Recognition and Classification of Neoproterozoic Continental Margins

Active rifts to passive margins (miogeoclines)

Rift margins were common during the Neoproterozoic, the supposed time of breakup of the Rodinian supercontinent. These margins initiate by extension and consist, in their lower part, of oceanward thinning crust, as well as grabens and associated mafic and siliceous igneous rocks. Once established, these margins are characteristically succeeded by miogeoclinal margins containing oceanward-thickening wedges of shallow-water continental shelf rocks—a miogeocline. A typical Neoproterozoic miogeoclinal belt extends along the Neoproterozoic western margin of North America from northwestern Mexico (Stewart and others, 2002), across the western United States (Link and others, 1993; Stewart, 1970, 1991; Stewart and Suczek, 1977) and western Canada (Gabrielse and Campbell, 1992). Characteristically these continental shelf (miogeoclinal) rocks contain diamictite that are interpreted to be glaciogenic. Models of Neoproterozoic rift margins in western North America are presented by Stewart (1972, 1991), Ross (1991), and Ross and others (1995).

Margins exhibiting multiple episodes of rifting or reactivation

A continental margin is commonly considered to have been formed by a single rift event. It has become increasingly clear, however, that many margins have undergone multiple episodes of rifting. For example, rifting in western North America (see more details under "Laurentia," below) probably began about 1,470 Ma (Evans and others, 2000) and formed a continental margin along which the Mesoproterozoic Belt and Purcell Supergroups were deposited (Burchfiel and others, 1992; Burke and Dewey, 1973). Another major rift occurred along most of the North American Cordillera at about 750 Ma (Gabrielse and Campbell, 1992; Link and others, 1993; Stewart, 1972; Stewart, 1978; Stewart, 1991), and a somewhat older event (770 Ma to ~750 Ma) has been proposed in central and northern Utah (Dehler and others, 2005). Finally, a controversial rifting event in western North America, at about 600 to 650 Ma, is indicated by studies of thermally driven subsidence (Levy and Christie-Blick, 1991a; Link and others, 1993).

The concept of multiple times of rifting along a given continental margin may seem unusual, because clear-cut examples of this type of reactivation of continental margins in modern-day or even Phanerozoic plate tectonics are not evident. At least some of the multiple rifting events may be related to Wilson cycles, in which rifting occurs, the continents drift apart, an ocean is formed, and the continents then drift back together and the ocean is consumed. In this case, magmatic-arc rocks would be expected on one, or perhaps both, rift margins. Such arc rocks occurring along the eastern margin of South America appear to be related to a Wilson cycle in which South America and Africa drift apart and are then reassembled (Zhao and others, 2002).

Another possible example lies along the eastern margin of Australia (as described and referenced later), where microcontinents appear to have been removed sequentially in the Mesoproterozoic and Cenozoic. The simplest explanation of these multiple-rift margins is that they formed at continental/oceanic interfaces—perhaps along the trend of a major tectonic flaw, where repeated rifting and (or) injection of igneous magmas were more likely to have occurred.

Aulacogens

Aulacogens are structural troughs extending into continents at a high angle to the trend of the continental margin (Burke, 1977; Burke and Dewey, 1973). They occur throughout the Proterozoic and Phanerozoic (Lobkovsky and others, 1996; Sengor and Natal'in, 2001; Shpunt, 1988) but may be more widespread in the Neoproterozoic than at other times. Most of these troughs are failed arms of three-armed rift systems, and thus in themselves indicate the existence of a continental margin formed by the other two arms of the aulacogen. The aulacogens are characterized by thick accumulations of sedimentary rocks and minor igneous rocks. Neoproterozoic aulacogens are recognized in North America, western and eastern Baltica, Siberia, India, Africa, South America, and Antarctica. All are described below, under descriptions of the individual continents. The abundance of extension-related aulacogens in the Neoproterozoic enhances the concept that the Neoproterozoic was a time of crustal fragmentation.

Long-lived miogeoclinal margins (Mesoproterozoic continuing into Neoproterozoic)

A miogeocline is defined as a structural feature consisting of a wedge-shaped continental margin deposit, similar to Cretaceous and younger deposits along the Atlantic margin of the United States (Dietz and Holden, 1967), that forms along rifted continental margins (the Cordilleran miogeocline).

Most Neoproterozoic miogeoclinal margins appear to have begun forming roughly at 850 to 740 Ma, presumably as the result of the fragmentation of the Rodinian supercontinent. However, the Proterozoic continental margins in western and northern North America, in the Ural Mountains, in Siberia, and perhaps in China are of different origin. In these areas, deposition on continental margins appears to have begun in the Mesoproterozoic (as early as 1,470 Ma) and to have continued into the Neoproterozoic. Recognized unconformities, in these cases, do not appear to represent times of major orogenic events but merely interruptions in sedimentation due to relatively minor structural dislocations. The impression is that these margins existed for a long time and marked the boundary between continental and oceanic domains. Continents bounded by these long-lived margins may have moved independently, rather than as coherent parts of Rodinia.

Rifted tectonic slivers, ribbon continents, and microcontinents

Rifted tectonic slivers, "ribbon continents," and microcontinents are not easily recognized in the Neoproterozoic, although they are often inferred to explain multiple rifting events, for example in the Cordillera of western North America. In the southeastward Canadian Cordillera, Colpron and others (2002) suggest, in one of two hypotheses, that a major Neoproterozoic rift was followed about 170 m.y. later by a second rift that completed the separation of a ribbon continent (microcontinent) from the older continental margin.

The concept that slivers, ribbon continents, and microcontinents formed by rifting from a continental margin is best exemplified by Mesozoic and Cenozoic examples (Gaina and others, 1998; Gurnis and Mueller, 2003; Mueller and others, 2001). In particular, along the eastern margin of Australia, at least three microcontinental blocks rifted away from the continental margin in the Mesozoic and Cenozoic, their separations energized, in most interpretations, by subduction or by a mantle plume. An alternate interpretation is that repeated rifting occurred along an elongate zone of weakness at the boundary between continental and oceanic crust, perhaps enhanced by mantle upwelling.

Outboard and accreted terranes

Outboard and accreted terranes are widespread along Mesozoic and Cenozoic continental margins (Jones and others, 1983; Silberling and others, 1984) but appear to be sparsely distributed, with the exception, perhaps, of outboard terranes related to subduction along Neoproterozoic margins. Outboard terranes of known or possible Neoproterozoic age are recognized in the Cordillera de Mérida (Bella Vista Greenschist and associated granitoids) of northern South America (Case and others, 1990); in the Klamath Mountains of northern California, where they consist of ophiolites (Mankinen and others, 2002; Wallin and others, 1991; Wallin and others, 2000) and sediments with Ediacaran fossils (Lindsley-Griffin and others, 2003); and in Alaska, where they include sedimentary rocks of the Nixon Fork terrane (Patton and others, 1994), granitic rocks (Karl and Aleinikoff, 1990; Patrick and McClelland, 1995), and metamorphic assemblages that generally are poorly dated and in which the distribution of Neoproterozoic rocks is poorly constrained [the Alexander terrane of Early Cambrian and possibly Neoproterozoic-age rocks of the coastal Cordillera of Canada (Gehrels, 1990); the Ruby terrane of west-central Alaska (Patton and others, 1994); and the Seward terrane of northwest Alaska (Till and Dumoulin, 1994)]. The time of accretion of the Neoproterozoic terranes is probably mostly Paleozoic and Mesozoic, but not Neoproterozoic. Other outboard or accreted terranes appear to be widespread in central Asia (Khain and others, 1997).

Collisional margins (sutures)

Plate convergence in the Mesoproterozoic, presumably related to the subduction of oceanic crust, led to igneous activity and continental collision. This activity has produced belts of metamorphic and igneous rock such as Grenville and related rocks (~1,200 to 900 Ma) and the Pan-African orogeny of the late Neoproterozoic in the southern continents. Neoproterozoic collisional margins are difficult to recognize, probably because the Neoproterozoic is mostly a time of fragmentation rather than collision. Nevertheless, Wilson cycles resulting in the opening and closing of oceans imply collision during ocean closure. Such a Neoproterozoic Wilson cycle is apparent between Africa and South America (Zhao and others, 2002), and probably elsewhere in the southern continents during the Pan-African orogeny.

Magmatic arcs

Magmatic-arc rocks are distributed in the African-Nubian shield, in the East African orogenic belt, in a part of northwest Africa, in the eastern United States, perhaps in Yucatan (Mexico), and in Europe (Plate 1). Many of these rocks lie near continental margins and probably formed there, whereas other rocks, particularly in Europe, occur as relatively small outcrops scattered over large regions. The magmatic-arc rocks probably formed in a variety of settings, from continental-margin volcanic terranes to outboard terranes formed by back-arc spreading.

Inherited margins

Some Neoproterozoic continental margins appear to be inherited from older margins. Mesoproterozoic Grenville-age rocks (1,200 to 900 Ma) formed by continental collision- and subduction-related processes along Mesoproterozoic continental margins. The Neoproterozoic margins mimic the location and trend of the Mesoproterozoic margins and, thus, appear to have been inherited from the location of these older margins (Plate 2).

Some Neoproterozoic continental margins (eastern and northern North America, arctic Ural Mountains, Siberia, and perhaps China) were initiated in the Mesoproterozoic and had continuous or intermittent miogeoclinal-margin deposition extending into the Neoproterozoic. An even older inheritance is suggested by the work of Rogers (1996), who proposed the assembly of major Proterozoic continents on the basis of the "oldest laterally extensive supracrustal sequences that lie on igneous-metamorphic basement over an area of 10,000 square miles" and on the basis of isotopic data that indicate that the youngest juvenile crust (mantle-derived) was created shortly before the deposition of supracrustal rock. Rogers (1996) proposed that east Gondwana was assembled at about 3,000 Ma, that a joined South America and Africa was assembled at about 2,000 Ma, that North America and Greenland were assembled from 2,500 Ma to 1,000 Ma, and that Europe and Asia were also assembled at 2,500 Ma. In particular, the configuration of the continent consisting of South America and Africa (joined at 2,000 Ma) implies continental margins similar, at least in places, to those described here, indicating an inheritance of the trends of Neoproterozoic continental-margin rocks from trends in older margins.

Stripped and covered margins

Many of the presumed continental margins described here are not characterized by sedimentary or igneous rocks (the defining features of many known Neoproterozoic margins). Most of these now-bereft margins were once the sites of deposition of Neoproterozoic sedimentary and igneous rocks, but these deposits have been stripped away at some time since their deposition, either by subaerial erosion, subduction erosion (Bourgois and others, 1996; Clift and Vannucchi, 2004; Sage and others, 2006; von Huene and Scholl, 1991), rifting, or extensional tectonic denudation or were covered by thrust nappes or by younger rocks. D.W. Scholl (oral commun., 2005) proposes that inward removal of crust by subduction erosion can take place at a rate of 2 to 3 km per million years.

Wilson cycle

J. Tuzo Wilson (1966) originally suggested that North America and Baltica were separated in the Paleozoic by the proto-North Atlantic Ocean. The continents then drifted together, closing the ocean, and finally the ocean reopened when the continents again drifted apart. Such a cycle of opening and closing of an ocean basin is referred to as a

Wilson cycle. The concept can be applied to a middle Neoproterozoic opening of the proto-North Atlantic Ocean (Dewey, 1974; Soper, 1994; Strachan and Holdsworth, 2000a, b; Winchester, 1988) and its closure possibly in the late Neoproterozoic and again in the Paleozoic. A concept of a late Neoproterozoic closure is related to the presence of magmatic-arc rocks along the eastern margin of North America (which, in this interpretation, would represent areas of subduction and ocean closure). The general view, however, is that these magmatic-arc rocks are far traveled and not indigenous to North America and that ocean closure occurred in the mid-Paleozoic (Dewey, 1974). As indicated by Zhao and others (2002), an important Neoproterozoic Wilson cycle occurs between South America and west Africa. This case, and other Neoproterozoic Wilson cycles, will also be discussed later in this report.

Middle and Upper Neoproterozoic Continents and Continental Margins

Understanding the distribution of continental-margin deposits is a vital tool in outlining the shape of continents and in determining the tectonic history of these margins. The subject was approached by preparing two world maps (Plates 1, 2). One of these maps (Plate 1) shows the distribution and lithologic character of the middle and upper Neoproterozoic (ca. 870 to 540 Ma—note that this does not include rocks older than Middle Neoproterozoic or younger than Upper Proterozoic) deposits, and the other (Plate 2) shows the location of continents, of continental margins of associated structures, and of Grenville-age (ca. 1,200 to 900 Ma) rocks considered to lie along Mesoproterozoic continental margins and to be the precursors of Neoproterozoic margins. These maps document Neoproterozoic features related to the breakup of Rodinia that are pertinent to any proposed Mesoproterozoic assembly.

The middle and late Neoproterozoic time interval is important in Proterozoic history, because it represents the time of the breakup of the hypothetical supercontinent Rodinia. If the concept of Rodinia is correct, then understanding the distribution of continents and their margins is key to deciphering the pattern and history of breakup of the supercontinent. A continent-by-continent description of the Neoproterozoic continents and continental margins is presented here.

Laurentia

Laurentia is the largest Neoproterozoic continent and the easiest to outline. As recognized by Stewart (1976), discontinuous Neoproterozoic rocks circumscribe the continent. The most continuous of these deposits extend along the western margin of North America, from northern Mexico to central and eastern Alaska (Brabb and Churkin, 1969; Christie-Blick and Levy, 1989; Link and others, 1993; Patton and others, 1994; Rainbird and others, 1996; Ross, 1991; Ross and others, 1995; Stewart, 1972; Stewart, 1976; Stewart and others, 2002). The deposits consist mostly of shallow-water outward-thickening miogeoclinal-margin deposits of carbonate rocks, shale, and siltstone. Glaciogenic deposits, including diamictite that is generally considered to consist of tillite, are associated with the continental shelf deposits and follow the same trend as the these deposits. Older parts of the Neoproterozoic succession are generally of about 740 to 780 Ma age (Ross and others, 1995; Stewart, 1991) and the upper part grades into the Lower Cambrian. Although the 740 to 780 Ma rifting is important, the continental margin of western North America appears to have undergone multiple rifting events. Rifting began at about 1,470 Ma (Evans and others, 2000) and formed a presumed continental margin along which the Mesoproterozoic Belt and Purcell Supergroups were deposited. An aulacogen formed an extension of Belt and Purcell Supergroups into the continent (Burchfiel and others, 1992; Burke and Dewey, 1973). Another major rift occurred along most of the North American Cordillera at about 750 Ma (Gabrielse and Campbell, 1992; Link and others, 1993; Stewart, 1972, 1978, 1991), and a somewhat older event (770 Ma to ~750 Ma) has been proposed in central and northern Utah (Dehler and others, 2005). Finally, a possible rifting event at about 600 to 650 Ma may be indicated by thermally driven subsidence (Levy and Christie-Blick, 1991b; Link and others, 1993). However, the 600 to 650 Ma rifting event has been proposed on the basis of subsidence models that have been challenged, because the time of proposed rifting corresponds in age to the well-defined passive-margin miogeocline deposits, which are presumably older than the age of rifting based on subsidence models. An alternative view presented here holds that the subsidence models indicate a time of rifting that may correspond to a time of eruption of scattered basalt flows in central Utah near the "Wasatch line," a zone of significant westward increase in the thickness of Neoproterozoic strata, perhaps related to rifting.

The fragments of the circumscribed continental margin of Laurentia are also recognized on Ellesmere Island in the Canadian Arctic (Frisch and Trettin, 1991; Trettin, 1991), in Greenland (Fairchild and Hambrey, 1995; Sonderholm and Jepsen, 1991; Sonderholm and Tirsgaard, 1993; Surlyk, 1991; Tirsgaard and Sonderholm, 1997; Watt and Thrane, 2001; Winchester, 1988), in Svalbard (Gee and Teben'kov, 2004), and in the foreland of the Caledonian orogen of Scotland (Dalziel and Soper, 2001; Duff and Smith, 1992; Soper, 1994; Strachan and Holdsworth, 2000a, b).

A long belt of Neoproterozoic-margin deposits extends along the eastern margin of North America from Newfoundland to the southern United States (Rankin and others, 1989). These deposits are primarily shelf clastics composed of conglomerate, fine to coarse sandstone, siltstone, mudstone, and glaciogenic diamictite. The miogeoclinal margin developed by rifting that started at about 760 Ma in the southern Appalachia and has been proposed to be as young as 550 to 620 Ma in the northern Appalachia (Cawood and others, 2001 and references therein; Hibbard and others, 2005; Su and others, 1994). The disparate ages between these two regions may indicate multiple occasions of rifting along the eastern margin of North America, much like the western margin of North America, where several intervals of rifting are proposed, ranging from about 1,470 to 600 Ma. Thomas (2006) has described the inheritance of continental margins in eastern North America through several Wilson cycles. These cycles consist of (1) the assembly of Rodinia as recorded in the Grenville orogeny, (2) the breakup of Rodinia, (3) the opening of the Iapetus Ocean, (4) the assembly of Pangaea as recorded in the Appalachian orogen, and (5) the breakup of Pangaea with the opening of the Iapetus Ocean.

Outboard of the miogeoclinal-margin deposits in eastern North America are accreted terranes composed primarily of magmatic-arc rocks (see further discussion under Cadomian and Avalonian magmatic arcs, below), including ophiolites (Dennis and Shervais, 1996; Hibbard and others, 2005; Keppie and others, 1991; Murphy and others, 1999; O'Brien and others, 1996; O'Driscoll and others, 2001) that generally range in age from 600 to 500 Ma. Some of these terranes may be exotic to North America, but others could have been produced by magmatic arcs outboard of the eastern North America miogeoclinal margin. In either case, these rocks formed along, or were accreted to, an already existing continental margin.

The southern margin of Laurentia is poorly defined but is considered to extend across the southern United States as a rift-transform fault system (Poole and others, 2005; Thomas, 1989, 1991, 2006). The Oklahoma aulacogen is a Precambrian and Lower Cambrian structure (Ham and others, 1964) that extends northwestward from the southern margin of Laurentia and probably represents a failed arm of a three-arm rift system. If so, the second arm extended to the southwest and the third arm to the east, a pattern compatible with a generally southward-rifting southern margin of Laurentia. The southern margin of Laurentia does not appear to extend south of latitude 28º in northern Mexico (Stewart, 1988).

South America

The Neoproterozoic margins of South America are difficult to define, because Neoproterozoic rocks are largely absent along much of the presumed margin. The problem is compounded by the presence in the southwestern part of Brazil of ocean-floor rocks within interior parts of the continent. These rocks generally have been considered to mark the boundary of major continental blocks that subdivide the continent into three major independent cratons (Weil and others, 1998). However, continental-margin rocks along parts of the South American continent and an inherited continental margin along the western side of South America suggest that the entire continent may have had some coherence in the Neoproterozoic. In this case, the boundaries between the Amazonia, Saõ Francisco, and Rio de La Plata blocks may be sites of relatively minor continental separation, perhaps involving Wilson cycles of opening and closing of ocean basins.

Traversing the continental margin in a clockwise direction, starting in northern Brazil, Neoproterozoic rift to miogeoclinal-margin deposits are widespread in northern Brazil and consist, in the rift stage, of sandstone, conglomerate, and pelite and, in the miogeoclinal-margin stage, of a varied assemblage of sandstone and sandy pelitic debris flows and turbidity currents, iron formation, and diamictite associated with carbonate rocks (Pedrosa-Soares and others, 2001). Although difficult to date, these rocks may have formed after a rifting event at about 800 Ma (Pedrosa-Soares and others, 2001). Miogeoclinal-margin deposits may be older than 686 Ma (D'Agrella-Filho and others, 2000; Evans, 2000; Kaufman and others, 1997) or as young as 670 to 600 Ma (Misi and Veizer, 1998). The miogeoclinal-margin deposits contain diamictite, which is characteristic of continental-margin deposits such as those that circumscribe Laurentia and are similar to continental-margin deposits along western Africa.

South of northern Brazil, a continuous band of Neoproterozoic rocks extends southward along the eastern margin of South America (Babinski and others, 1996; da Silva and others, 2005; Gaucher and others, 2004; Heilbron and Machado, 2003; Pedrosa-Soares and others, 2001). The oldest of these rocks may have formed after rifting at about 875 Ma, but much of the belt is a magmatic arc that formed from 560 to 500 Ma. The belt also contains small outcrops of ultramafic and mafic rocks. The 875 Ma rocks, perhaps widespread, may have been overprinted by the 560 to 500 Ma magmatic rocks.

The problematic rocks in the Tocantins Province (Pimentel and Fuck, 1992; Pimentel and others, 1999) in central Brazil contain diamictite-bearing sedimentary rocks that may be similar in age to those in northern Brazil (800 to 586 Ma) and in the Congo belt of western Africa (Tack and others, 2001; Trompette, 1994). The Province also contains 800 to 700 Ma syn-collisional granitoids, 900 to 630 Ma arc-related granitoids, and 590 to 480 Ma bimodal rocks (Pimentel and others, 1999). Ultramafic and mafic bodies associated with 590 to 485 Ma granites are also present. Pimentel and others (1999) indicate collisional and extensional events and the opening of a large ocean basin west of the Saõ Francisco block. An alternative idea is that the Tocantins Province (Plate 2) is the site of a mantle plume or bolide impact, as suggested by the circular pattern of Neoproterozoic rocks or perhaps radiating trends of Neoproterozoic rocks (Plate 2) similar to star-shaped rifts, including the three-armed rifts of aulacogens described previously (Sengor and Natal'in, 2001).

Further outcrops that may indicate the position of the continental margin are in Argentina, Chile, and Bolivia, where rocks are assigned to the Puncoviscana Formation, a poorly dated assemblage of fine and coarse turbidites and pelites (Keppie and Bahlburg, 1999; Omarini and others, 1999). Trace fossils in the Puncoviscana Formation indicate a Cambrian and Neoproterozoic age (Acenolaza, 2004).

The Puncoviscana Formation has commonly been considered to be a miogeoclinal-margin deposit along the western edge of the South American continent. However, Omarini and others (1999) and Keppie and Bahbug (1999) indicate that the formation may be a foreland basin deposit related to a coeval Cambrian magmatic arc, whereas Escayola and others (2007) consider the formation to be a back-arc deposit. Arguing against these proposals is the indication that the Puncoviscana Formation does not contain volcanic detritus (Keppie and Bahlburg, 1999), as would be expected in sediments near a magmatic arc. In addition, some rocks in the formation are considered as Neoproterozoic (Acenolaza, 2004), an age older than the dated Cambrian age of the magmatic-arc rocks. In any of these three proposals, the Puncoviscana Formation lies near the western margin of South America, either as continental-margin deposits or as foreland or back-arc basin deposits.

North of the outcrops of the Puncoviscano Formation, only three localities of Neoproterozoic, or possibly Neoproterozoic, rocks are known in South America. These are (1) the diamictite-bearing sedimentary rocks in the Arequipa Massif of coastal Peru (Shackleton and others, 1979), which are dated as Neoproterozoic on the basis of C isotope studies (F.A. Corsetti, written comm., 2002), (2) greenschist and granitoids of Cambrian or possible Neoproterozoic age in the Cordillera of Mérida in northwestern Venezuela (Case and others, 1990), and (3) possible Neoproterozoic rocks in the subsurface in northern Venezuela (Feo-Codecido and others, 1984) that trend east-west and appear to lie along the northern margin of Neoproterozoic South America.

These scattered outcrops perhaps outline the general trend of the continental margin in western and northern South America, but not with any certainty. A better idea for the location of the Neoproterozoic margin along the western margin of South America is the position of Mesoproterozoic metamorphic and igneous rocks that are generally related to Grenville-age rocks (1,200-900 Ma). These rocks form a fairly continuous belt following the western margin of South America from central Chile to northern Colombia (Ramos and Aleman, 2000). The Grenville-age rocks are generally considered to have formed by continental collision or, perhaps in part, by subduction-related processes. In either case, they appear to have formed along a continental margin. The Neoproterozoic margin may, therefore, have been inherited from the Grenville-age margin in the same manner as Neoproterozoic rocks in the eastern United States and eastern Canada (Thomas, 2006).

Africa

Africa is generally divided into six Neoproterozoic blocks: West Africa, Trans-Sahara, Congo, Kalahari, East African orogenic oelt, and East African craton (including Madagascar), all described below.

The East African orogenic belt is the site of an important continental margin that developed at about 850 Ma during a time of extension and presumable continental separation (Husseini and Husseini, 1990; Kroener and Stern, 2005; Kroener and others, 1987; Kusky and Matsah, 2003; Meert and Torsvik, 2003; Mosley, 1993; Muhongo and others, 2001; Pinna and others, 1993; Shackleton and others, 1979; Stern, 1994, 2002; Vail, 1985, 1987; Willis and others, 1988). Poorly dated sedimentary successions that may be miogeoclinal-margin deposits related to this margin are described in Sudan and Kenya (Kroener and others, 1987; Stern, 1994 and references therein). The margin marks the boundary between continental crust on the west and ocean-floor and magmatic-arc rocks to the east, which in turn are flanked by continental rocks farther to the east (Stern, 2002). The ocean-floor and magmatic-arc rocks formed during rifting and subsequent closure of an ocean basin between about 750 and 500 Ma (Stern, 1994).

Outcrops of juvenile rocks are widest in northeastern Africa and adjoining Saudi Arabia and Yemen, narrower to the south, and disappear in Tanzania and Zambia. However, the East African orogenic belt, which is presumably a boundary between continental blocks, extends south of the southern limit of juvenile rocks. Rocks of the East African orogenic belt are correlated with tectonic events in Antarctica (Jacobs and others, 1998; Jacobs and Thomas, 2004; Porada, 1985).

The East African orogenic belt is bounded on the east by various continental blocks including basement rocks in Saudi Arabia, Yemen, easternmost Africa, and Madagascar (reconstructed to its probable original position relative to Africa). On the west, the belt is bounded by the Trans-Sahara craton, the Congo craton, and the Kalahari craton. These cratons likely originated from the breakup of a larger block and, therefore, must have been more or less in place relative to each other by about 870 Ma, the time of their fragmentation. They reassembled during the Pan-African orogeny at about 870 to 550 Ma (Kroener and others, 1987). The Trans-Sahara craton is a poorly outlined block extending across northern Africa from the west side of the East African orogenic belt to the north of the Congo craton, to the east of northwestern Africa, and far to the south of fragmentary outcrops of magmatic-arc rocks in the northern Mediterranean region. The Trans-Sahara craton has also been called the Saharan metacraton by Abdelsalam and others (2002), who believe that the craton has been "remobilized during an orogenic event but is still recognizable dominantly through its rheological, geochronologic and isotopic characteristics." Most of the craton consists of medium- to high-grade gneisses, metasedimentary rocks, migmatites, and granulites that were produced by remobilization of pre-Neoproterozoic rocks. Low-grade metamorphic volcano-sedimentary rocks, also present, are intruded by granitoids ranging in age between 750 and 550 Ma.

The eastern boundary of the Trans-Sahara craton, as mentioned above, is along the East African orogenic belt. The southern boundary is along the Qubanguide fold belt (Trompette, 1994; Unrug, 1996), to the south of which is the Congo craton, containing along its northern margin large areas of Neoproterozoic sedimentary rocks, presumably continental margin rocks, locally containing diamictite.

The Congo craton is characterized by a ring of discontinuous sedimentary rocks, many containing diamictite (Plate 2). This pattern is similar to the circumscribing continental-margin rocks containing diamictite-bearing glaciogenic sedimentary rocks in Laurentia. But, the boundaries of the Congo craton are complex (De Waele and others, 2008). On the southeastern boundary is the Zambezi belt in Zambia, Mozambique, and Zimbabwe (Dirks and others, 1998; Goscombe and others, 2000; Hanson, 2003; Hanson and others, 1994; Johnson and Vail, 1965; Porada and Berhorst, 2000). The Zambezi belt is a major east-west-trending sedimentary and structural belt extending inland at a right angle to the north-south-trending Mozambique belt (part of the East African orogenic belt), which is presumed to track the eastern boundary of the continental margin in central Africa.

The Zambezi belt is considered to be an aulacogen or rift (Olade, 1980) extending inland to the west from a north-south-trending continental margin along the trend of the Mozambique belt. If so, the 850 Ma age of the initial deposits in the aulacogen dates the time of breakup of a major tectonic block. Inland, the Zambezi belt joins the north-south-trending Lufilian belt (Cailteux and others, 1994; Jackson and others, 2003; Porada and Berhorst, 2000; Unrug, 1983; Wendorff, 2005), which is characterized by thick sequences of sedimentary rocks containing evaporates deposited starting at about 880 Ma (Porada and Berhorst, 2000). The relationship of the Zambezi and Lufilian belts is not clear, but the two belts together with the Damara belt in west Africa appear to have originated as an intracontinental three-armed rift system related to the breakup of a major tectonic block. The intracraton rifting, however, is interpreted to be succeeded in the Zambezi belt by the opening of an ocean basin, so interpreted on the basis of the presence of eclogites (John and others, 2003; Vrana and others, 1975) in the western part of the Zambezi belt. John and others (2003), on the basis of geochemical studies, indicate that the eclogites are a mid-ocean-ridge type, and the ocean basin was over 1,000 km wide.

West of the Zambezi belt, the Lufilian belt near the southern margin of the Congo craton is generally considered to join with the northeast-trending Damara belt in west Africa. The Damara belt (Germs, 1995; Hanson, 2003; Jung and others, 2001; Porada, 1985; Trompette, 1994) is considered here to be an aulacogen extending inland from a north-south-trending continental margin. North of the Damara belt, the western margin of the Congo craton contains the Kaoko belt (Duerr and Dingeldey, 1996; Seth and others, 1998; Trompette, 1994) and the west Congo belt (Tack and others, 2001; Trompette, 1994). Both belts contain diamictite-bearing sedimentary rocks considered to be upper Neoproterozoic, in part about 760 to 750 Ma (Hoffman and others, 1998) and with minimum ages of 700 to 620 Ma (Evans, 2000). The west Congo belt contains older rocks (1,000-910 Ma) that are indicative of rifting. These rocks suggest two ages of rifting in the west Congo belt: one ranging in age from 1,000 to 910 Ma and one in the late Neoproterozoic, somewhat older than 760 to 750. The Otavi Platform contains diamictite-bearing sedimentary rocks and evidence of rifting from about 760 to 750 (Hoffman and others, 1998) and probably an older event from 800 to 750 Ma (Hoffmann and others, 2004).

The Kalahari craton, as mentioned above, is bounded on the north by aulacogens. The western boundary is characterized by diamictite-bearing sedimentary rocks of the

Gariep belt (Frimmel and Foelling, 2004; Frimmel and Frank, 1998; Frimmel and others, 1996a; Frimmel and others, 1996b; Hanson, 2003; Jacobs and others, 2008). Frimmel and Fölling (2004) indicate that the minimum age of continental rifting in the Gariep belt is about 740 Ma (Cordani and others, 2003a, b).

West Africa is a circular craton roughly outlined by discontinuous outcrops of diamictite-bearing sedimentary rocks and locally by mafic and ultramafic ocean-floor rocks. The diamictite-bearing sedimentary rocks are poorly dated, but many of the dates range from 700 and 600 Ma (Clauer and Deynoux, 1987; Evans, 2000). The sedimentary rocks appear to have been deposited along a continental margin that was produced by rifting at about 700 Ma or earlier (Hefferan and others, 2000; Villeneuve and others, 1993). The younger Neoproterozoic history of the West Africa continental margin is complex and includes collisional events, development of magmatic arcs, and (along the east side of West Africa and adjacent areas) extensive terranes of high-grade gneiss and igneous complexes of the Pan-African orogeny (Attoh and others, 1997; Black and others, 1994; Black and Liegeois, 1993; Caby and others, 1989; Hefferan and others, 2000; Inglis and others, 2005a; Leblanc and Moussine-Pouchkine, 1994; Liegeois and others, 1994; Samson and others, 2004; Trompette, 1994; Villeneuve and others, 1993; Villeneuve and Dallmeyer, 1987).

Antarctica

Neoproterozoic and Early Cambrian rocks in the Trans-Antarctic Mountains (Goodge, 2002; Goodge and others, 2002, 2004; Laird, 1991; Schmidt and others, 1965; Stump, 1982) are combined on Plate 1 because rocks of this age are difficult to separate. Definite Neoproterozoic rocks are recognized in the Beardmore Group of the central Trans-Antarctic Mountains (Goodge and others, 2004), but elsewhere recognition of Neoproterozoic rocks is questionable and most of the rocks shown on Plate 1 in the Trans-Antarctic Mountains may be Early Cambrian and even Middle Cambrian in age (Rowell and others, 2001). Nevertheless, the Neoproterozoic and Early Cambrian rocks form a tectonic package interpreted to be a 3,600-km-long, continental-margin deposit bordering East Antarctica. The presence of diamictite in the Beardmore Group (Goodge and others, 2004) and Nimrod Glacier area (Stump and others, 1988) is consistent with the characteristic presence of these deposits along continental margins elsewhere in the Neoproterozoic. Goodge and others (2004) indicate that the continental margin is formed by rifting but note that the age of this rifting is in doubt. Some geologists consider it to be about 750 Ma (Goodge and others, 2004 and references therein). A miogeoclinal margin is considered to have existed from 670 to 580 Ma and transitionally into younger rocks. A magmatic arc began to form by at least 515 Ma (Goodge and others, 2004). Neoproterozoic rocks situated mainly on the opposite side of Antarctica from the Trans-Antarctic Mountains are high-grade metamorphic rocks and associated voluminous granitoids related to the Pan-African tectonothermal event (Boger and others, 2002; Carson and others, 1995; Fitzsimons, 2000b; Kamenev, 1993; Rajesh and others, 1996; Shiraishi and others, 1994; Stuewe and Sandiford, 1993). The Lambert Glacier-Prydz Bay structure (Mishra and others, 1999; Stagg, 1985), likely of Neoproterozoic age, is considered to be the failed rift arm of a triple junction (aulacogen). If so, the margin of Antarctica, at least in the vicinity of Lambert Glacier and Prydz Bay, was likely a rift margin related to the two active arms of this triple junction. Although much smaller, the Lützow-Holm Bay structure may be a second aulacogen.

Australia

Except for ophiolitic rocks in eastern Australia (Bruce and others, 2000), Neoproterozoic rocks in Australia lie west of the Tasman line that marks the eastern limit of Proterozoic rocks in central and northern Australia (Direen and Crawford, 2003; Preiss, 2000; Preiss and Forbes, 1981; Walter and Veevers, 1997). The line is interpreted as a Rodinian breakup boundary (Direen and Crawford, 2003).

The Adelaide "geosyncline" (Plumb, 1985; Preiss, 1987, 2000; Preiss and others, 1993; Preiss and Forbes, 1981; Veevers and others, 1997) in southern Australia contains, in its lower part, thick Neoproterozoic sedimentary rocks and mafic igneous rocks that are ascribed to initial rifting that led to the formation of a continental boundary. The sedimentary rocks in the Adelaide geosyncline thicken eastward along the Torrens Hinge Line (Powell and others, 1994), and the Adelaide rocks resemble miogeoclinal deposits found elsewhere in the world. No comparable miogeoclinal deposits are known north of the Adelaide area along the Tasman line; thus if the Tasman line is indeed a breakup boundary, miogeoclinal sediments along it either were destroyed or are covered by younger rocks. A further complication is the presence of the Curnamona craton east of the Adelaide belt. Either this craton is a microcontinental block marking the eastern depositional boundary of the Adelaide sedimentary basin, or it is an accreted microcontinent. Initial extension in the Adelaide basin leading to continental separation is dated at about 700 to 760 Ma (Powell and others, 1994), whereas extension and continental breakup in Tasmania are dated at 579 Ma (Meffre and others, 2004), indicating multiple times of rifting in Australia and Tasmania.

In central Australia, there are several west-northwest intracontinental sedimentary basins (Walter and Veevers, 1997), including the Officer-Savory basin (as much as 8,000 m of sediment), the Amadeus basin (as much as 5,500 m of sediment), the Ngalla basin (as much as 2,000 m of sediment), and the Georgina basin (as much as 6,000 m of sediment). These basins are considered here to be major structural features (aulacogens) extending inward at a high angle to the presumed Neoproterozoic continental margin along the Tasman line. In southern Australia, the Gairdner dyke swarm (Barovich and Foden, 2000; Park and others, 1995; Wingate and others, 1998; Zhao and others, 1994) and the Polda trough (Preiss, 2000; Preiss and others, 1993) has a north-northwest trend at a high angle to the presumed north-south trend of the Neoproterozoic margin and may be related to rifting similar to that which formed the deep structural troughs (aulacogens or intracontinental rifts) of central Australia.

Except for eastern Australia, the position, or presumed position, of the Neoproterozoic margin is uncertain. In southern Australia and western Australia, Grenville-age rocks (Pinjerra and Albany-Fraser orogenic belts) fringe, or are near, the continental margin (Dawson and others, 2003; Fitzsimons, 2003; Myers and others, 1996). These belts mark the site of a Grenville-age continental margin that probably was reestablished in Neoproterozoic time.

India and related areas

The Indian continental block as described here includes the present country of India, as well as Nepal, Bhutan, and parts of Pakistan and Afghanistan. In northern India, Bhutan, and Nepal, Neoproterozoic rocks are widely exposed in the Lesser Himalaya and, to a lesser extent, in the High Himalaya. These Neoproterozoic rocks extend for about 2,500 km along the southern margin of the Himalaya and contain a thick, northward-thickening miogeoclinal accumulation of mostly clastic sedimentary rocks and metasedimentary rocks (Brookfield, 1993; Jiang and others, 2003; Paliwal, 1998; Srikantia and Sharma, 1972; Valdiya, 1995; Virdi, 1998). This accumulation marks the northern continental boundary of the Indian plate formed by rifting and the formation of a miogeocline. The miogeoclinal rocks in the western part of the Lesser Himalaya contain glaciogenic diamictite deposits (Hambrey and Harland, 1981; Virdi, 1998), in places associated with evaporites (Srikantia and Sharma, 1972; Virdi, 1998). Successions containing diamictite and associated evaporites also occur to the south and southwest of the west end of the Lesser Himalaya, in the western part of the Rajasthan of India, and in the Salt Range of Pakistan. Neoproterozoic strata in the High Himalaya may represent outboard strata in the miogeocline or deposition in a separate basin (Brookfield, 1993). The Pakistan Himalaya, as described by Brookfield (1993), shows distinct stratigraphic and structural differences from the Indian Himalaya (Brookfield, 1993), suggesting the possibility that the Pakistan Himalaya is an outboard block along the western boundary of the Indian block. But, the boundary between the Indian block and the Iran and Arabian Peninsula block is ill defined, as described below, leaving open the possibility that the Indian block and the Arabian block were connected.

The main part of the Indian block is characterized by scattered basins of Neoproterozoic rocks and by Neoproterozoic rifts (aulacogens) extending inward from interpreted Neoproterozoic continental margins. The Eastern Ghats Belt of the eastern margin of southern India consists of high-grade metamorphic rocks of Grenville age (Paliwal, 1998; Rickers and others, 2001; Yoshida and others, 1996). This belt is considered to have formed along a continental margin, probably by continental collision. A Neoproterozoic margin is inferred to have developed along a preexisting Grenville margin. Such a margin is interpreted from the presence of two major northwest-trending intracontinental grabens (aulacogens) that extend inland at high angle to this inherited Neoproterozoic margin (Biswas, 2003; Chaudhuri and others, 2002; Krishna Brahman and Negi, 1973; Pandey and Agrawal, 1999; Raval and Veeraswamy, 2003). Neoproterozoic rocks are also present in intracontinental basins outside of these grabens and in other intracratonal rifts (Jiang and others, 2003) and have also been recognized in southern India (Krishna Brahman and Negi, 1973) and below the rocks of the Deccan Traps in western India (Krishna Brahman and Negi, 1973).

The largest area of Neoproterozoic sedimentary rocks, including some parts dated as Paleoproterozoic and Mesoproterozoic, extends west-northwest in large, but scattered, outcrops across central India (Bose and others, 2001; Goodwin, 1991; Rasmussen and others, 2002; Ray and others, 2002; Valdiya, 1995). These intracratonal rocks, assigned to the Vindhyan Supergroup, consist mostly of sandstone, shale, limestone, and local diamictite (Hambrey and Harland, 1981) as thick as 4,259 m. A major basin lies along the eastern side of the main area of outcrop of the Vindhayan Supergroup. This basin is within a major east-northeast structural zone for which several names have been applied [Central Indian Tectonic zone (Chaudhuri and others, 2002); Son-Narmado-Tapti graben (Pandey and Agrawal, 1999); Satpura mobile belts (Raval and Veeraswamy, 2003)]. This structural zone, which extends southwest to the western side of India, is perhaps a rift zone or aulacogen along which rocks of the Vindhayan Supergroup were concentrated. Other areas of the Vindhayan Supergroup do not appear to be related to major rift features, indicating that deposition of some Neoproterozoic rocks in India took place within a stable continent.

The southernmost part of India and adjacent Sri Lanka contain Neoproterozoic high-grade metamorphic rocks and associated granitoids that, in places, have reworked older terranes (Bartlett and others, 1994; Miller and others, 1996; Yoshida and others, 1996; Yoshida and Vitanage, 1993). These rocks resemble metamorphic and granitoid terranes formed elsewhere in the Pan-African tectonothermal event.

Arabia and related areas

The Arabian continent was a relatively stable block in the Neoproterozoic, although much of the area is now broken by faults and divided into separate blocks. In Iran, Neoproterozoic stratigraphic units can be traced from range to range, indicating a stratigraphic coherence across the now fragmented region (Berberian and King, 1981; Stoecklin, 1968). The Neoproterozoic rocks thicken in northern Iran, near the Caspian Sea, perhaps indicating that this margin is a miogeocline. Such a miogeocline trends east-west and is aligned with the east-west trend of Neoproterozoic rocks in the Lesser Himalaya, indicating a speculative tie between Iran and India. Such a tie is supported by the east-west trend of sedimentary deposits and faults across northern Iran (Kopet Dagh Fault) and a similar trend in Afghanistan (Herat Fault). The Herat Fault is indicated by Wensink (1991) to mark the boundary of Asian rocks on the north and Gondwana rocks on the south, although little of the area considered contains Neoproterozoic, or possible Neoproterozoic, rocks. The north-south-trending Caman-Moqui fault of Pakistan (Wensink, 1991) may constitute the break between the Peninsular Arabia-Iran plate and the Indian plate, or (as suggested above) these plates were originally joined, or at least were close together.

The idea that the Arabian Peninsula-India region is a somewhat coherent block is suggested by the presence of Neoproterozoic evaporite deposits across the region, including the Arabian Peninsula (Edgell, 1991; Husseini and Husseini, 1990; Mattes and others, 1990), Iran (Edgell, 1991; Mattes and others, 1990; Srikantia and Sharma, 1972), Pakistan (Virdi, 1998), and western India (Srikantia and Sharma, 1972; Virdi, 1998). These areas contain the most widespread Neoproterozoic evaporite deposits in the world. In detail, surface and subsurface outcrops show a definite continuation of evaporite deposits from the Arabian Peninsula into western Iran (Edgell, 1991).

Sedimentary rocks including diamictite and evaporite deposits occur along the southeastern coast of the Arabian Peninsula in Oman (Brasier and others, 2000; Gass and others, 1990; Gorin and others, 1982; Leather and others, 2002; Mattes and others, 1990). The presence of diamictite suggests that these deposits in Oman represent a continental margin, perhaps the remnants of deposits that originally lay between Peninsular Arabia and India, but, as described above, an alternative idea is that Peninsular Arabia and India were joined together in the Neoproterozoic.

South China

In South China, outcrops of Neoproterozoic sedimentary rocks and lesser amounts of volcanic and volcanoclastic rocks occur in a roughly circular area outlined, in part, by small bodies of irregularly distributed mafic, ultramafic, and magmatic-arc rocks. Sedimentary rocks of the Sinian System are the dominant sedimentary rock types (Wang and others, 2003; Yang and others, 1986) and consist of mudstone, shale, muddy siltstone, sandstone, conglomerate, carbonate rock, glacial diamictite, and volcanic-volcanoclastic rocks, which are more abundant in the lower Sinian. The succession thickens to the southeast and is over 5,000 m thick locally. Shallow-water or alluvial deposits are dominant in the eastern areas of the succession, and shallow to moderately deep-water deposits are dominant in eastern outcrops. Deep sub-basins interrupt this pattern in places. Sinian System rocks are interpreted to lie in two grabens, the Nanhua in eastern South China, and the Kangdian in western South China. Rather than considering these to be grabens, the interpretation here is that the sedimentary and volcanic rocks in these two areas are largely continental-margin deposits: the Nanhua related to a margin along the southeast side of South China and the Kangdian to a margin along the west side of South China. Overall, the Sinian System, at least in the southeastern part of South China, appears to be an oceanward-thickening continental shelf deposit (miogeocline) similar to those described elsewhere in the world (West Africa, Ural Mountains, western and eastern North America, and Siberia). Wang and others (2003) consider the sedimentary sequence to consist of four sequence-sets, representing four phases of rifting: one at about 820 Ma after bimodal magmatism; the second at about 800 Ma; the third, a major rift phase, at about 780 to 750 Ma; and the fourth recording the rift-drift transition at about 750 to 690 Ma. Outboard of the central area of largely sedimentary rocks in China are relatively small outcrops of mafic-ultramafic rocks and granitoids generally about 820 Ma in age (Chen and others, 1991; Li, 1998, 1999; Li and others, 2001, 2003b; Ling and others, 2003; Meng and Zhang, 2000; Zhang and others, 2003; Zhou and others, 2002). These rocks may be, in part, oceanic rocks formed during the breakup of the Rodinian supercontinent. Rocks along the Qinling belt (Zhou and others, 2002) in northern South China consist of volcanics and mafic intrusives that have been interpreted as being due to either continental collision, subduction, or rifting. Isotopic studies (Zhou and others, 2002) indicate that at least some of these rocks probably formed as a magmatic arc along a subduction zone, before rifting and widespread deposition of Neoproterozoic sedimentary rocks. Rocks in the Kangdian area have also been considered to be related to subduction, but rifting is suggested by the presence of bimodal igneous rocks (Li and others, 2001) that may be related to the breakup of the Rodinian supercontinent. Regardless of the origin of these rocks, all appear to have formed along a continental margin and are used to define the Neoproterozoic outline of the South China and North China continents.

The circular area of Neoproterozoic sedimentary rocks and associated granitoids and mafic-ultramafic rocks in South China are interpreted to be the result of a mantle plume at about 825 Ma that was responsible for the initial breakup of the Rodinian supercontinent (Li and others, 1995, 2003a, b; Li, 1998, 1999). Alternatively, the breakup is not related specifically to a plume but rather to broader-scale tectonics during the apparent worldwide breakup of the Rodinian supercontinent from about 870 to 750 Ma.

North China

North China and South China are juxtaposed along the Qinling belt in central China. This belt, as described above under "South China," is characterized by mafic-ultramafic, volcanic, and volcanoclastic rocks. The Qinling belt is clearly a major Neoproterozoic boundary, regardless of whether it is considered to have been formed by continental collision, subduction, or rifting. In the Neoproterozoic, the belt marks the northern boundary of South China, as well as the southern boundary of North China (Lu and others, 2008b). In areas outside of the Qinling belt, the boundary of Neoproterozoic North China is not well defined. The eastern margin may lie outboard of a northeast-trending group of outcrops on the eastern margin of North China that may extend into Korea (Lee and others, 1998; Rogers and Santosh, 2003), and the western margin may lie along an even less-well-defined group of outcrops of Neoproterozoic rocks (Plate 1). The northern boundary of North China lies south of the Central Asian Mobile Belt, described below.

Tarim

The Tarim block, of northeast China, is a relatively small block compared to such major continents as Siberia or Laurentia, and its Neoproterozoic outline is only vaguely defined on the basis of the sparse Neoproterozoic rocks that crudely circumscribe the block (Carroll and others, 2001; Lu and others, 2008a; Xu and others, 2005). Bimodal igneous rocks in the northern part of the Tarim block indicate a rifting event at 755 Ma. Phanerozoic tectonism may have significantly modified the shape of the block.

Central Asian Mobile Belt

The east-west-trending Central Asian Mobile Belt, also called the Central Asiatic fold belt or Asian fold belt (Yakubchuk, 2004; Zonenshain, 1973; Zonenshain and others, 1990), consists of a complex assembly of moderate-sized plates, microcontinents, magmatic-arc systems, and ophiolitic rocks (Stern, R.J., written commun., 2006; Kovalenko and others, 2004; Li and others, 2003b; Safonova and others, 2004; Yakubchuk, 2004). Though the boundary of this belt is poorly defined, the belt is clearly tectonically significant. The belt has a complex distribution of continents and microcontinents, but the most distinctive characteristic of the belt is scattered ophiolitic rocks that are more widespread than in any other cratonal part of the Neoproterozoic world. As defined by most workers (Stern, R.J., written commun., 2006; Li and others, 2003b), the Central Asian Mobile Belt includes (1) Kazakhstan on the west, (2) a broad region between the Tarim and North China blocks on the south, and the Siberia continent on the north, and, (3) east of there, a broad region of microcontinental blocks north of North China. Kazakhstan, which contains fewer ophiolitic rocks, may be more structurally coherent than most other parts of the Central Asian Mobile Belt.

Kazakhstan, is along the western margin of the Central Asian Mobile Belt and is a structurally complex block composed of Precambrian-Paleozoic rocks assembled along sutures and associated with magmatic belts (Avdeyev, 1984). In the Neoproterozoic, however, it may have been structurally coherent, because Precambrian basement rocks have been proposed to extend across the region (Khain and others, 2003). Neoproterozoic sedimentary rocks, including diamictite, are abundant in a band along the southern margin of the Kazakhstan block (Plate 1) and may represent a continental margin. The abundance of Neoproterozoic sedimentary rocks along the east side of the Kazakhstan block also suggests a continental margin in that area. Cook and others (1994) indicate that during the late Proterozoic, Kazakhstan underwent rifting and separation into smaller continents and microcontinents. Avdeyev (1984) reported Neoproterozoic bimodal igneous rocks associated with rifts.

The broad area of the Central Asian Mobile Belt east of Kazakhstan, between North China-Tarim on the south and Siberia to the north (mostly Russia and Mongolia) is characterized by multiple microcontinents, magmatic arcs, and abundant scattered ophiolites (Buchan and others, 2002; Buslov and others, 2002; Khain and others, 2003; Khomentovsky and Gibsher, 1996; Kovalenko and others, 2004; Kuzmichev and others, 2001; Li and others, 2003b; Mossakovsky and others, 1994; Pfaender and others, 2002; Windley and others, 2007; Yakubchuk, 2004; Yue and others, 2001).

As mentioned above, the continuation of the Central Asian Mobile Belt to the east is not clearly defined. The microcontinental blocks north of the North China block and south of the Siberian block clearly appear to be part of the belt, but, east of there, either the relatively large Kingan-Bureya block (Zonenshain and others, 1990), also called the Songliano block (Zhang and others, 1984), is included in the belt (Li and others, 2003b) or the belt may continue along the northern margin of this block (Stern, R.J., written commun., 2006).

Siberia

Thick successions of miogeoclinal-margin, outward-thickening, miogeoclinal deposits of Mesoproterozoic and Neoproterozoic age are exposed on the margins of Siberia. These exposures are (1) in the Olenek uplift and Kharaulakh Mountains, in northeastern Siberia (Pelechaty, 1996, 1998; Pelechaty and others, 1996; Pisarevsky and Natapov, 2003), (2) in the Yodoma-Maya area of southeastern Siberia (Khomentovsky, 1986; Khudoley and others, 2001; Pelechaty, 1996; Pisarevsky and Natapov, 2003; Pisarevsky and others, 2008; Rainbird and others, 1998), (3) in the Patom highland in southern Siberia (Pelechaty, 1998; Pisarevsky and Natapov, 2003), (4) in the Yenisey Ridge area, Turakhansk uplift, and Igarka uplifts in southwest Siberia (Pisarevsky and Natapov, 2003), and (5) in the Taymyr area in northwestern Siberia (Vernikovsky and others, 1998, 2004). These deposits are considered to have once circumscribed the Siberian continent (Chumakov and Semikhatov, 1981; Khomentovsky, 1986; Pelechaty and others, 1996; Pisarevsky and Natapov, 2003). Regional unconformities are recognized within these successions, but major orogenic events or magmatic arcs are not in evidence, except in the middle Neoproterozoic in southern, southwestern, and northwestern Siberia. The miogeoclinal-margin deposits of Siberia contain strata that are as old as 1,600 Ma and range upward in northeastern and southeastern Siberia into the latest Neoproterozoic, and even into the Cambrian. Such long-lived miogeoclinal-margin deposits are unusual but are similar to other such deposits in the Ural Mountains and possibly in China. Also, the western margin of Laurentia contains continental-margin deposits ranging in age from 1,400 Ma to latest Neoproterozoic; these deposits are not in a single miogeoclinal package but instead appear to have developed as a consequence of several rifting events on the same margin.

The Siberian miogeoclinal-margin deposits are considered to have been initiated by continental rifting in the Mesoproterozoic, perhaps, as in the Patom highlands, at about 1,600 Ma (Pisarevsky and Natapov, 2003). A 543 to 530 Ma rifting event, followed by an onset of regional thermal subsidence, has been proposed in northeastern Siberia (Pelechaty, 1996) and appears to indicate a second time of rifting on the miogeoclinal margin. The long-lived miogeoclinal margins of Siberia suggest that Siberia was an independent continent during much of the Mesoproterozoic and Neoproterozoic and not connected to the Rodinia supercontinent (Pisarevsky and Natapov, 2003).

Middle Neoproterozoic magmatic-arc rocks, granitic rocks, and associated mafic and ultramafic rocks occur in the Baikalia-Vitim highlands in southern Siberia, in the Yenisey Ridge in southwest Siberia (Vernikovsky and others, 2003, 2004; Volobuyev, 1994), and in the Taymyr area of northwest Siberia (Vernikovsky, 1995; Vernikovsky and Vernikovskaya, 2001; Vernikovsky and others, 1998). These rocks mark the end of the Mesoproterozoic and Neoproterozoic miogeoclinal margin in these areas, but in southeastern and northeastern Siberia the miogeoclinal margin persisted into latest Neoproterozoic and Cambrian time. Although outcrops are scattered, the magmatic-arc rocks and mafic and ultramafic rocks circumscribe the Siberian continent, as did the miogeoclinal-margin deposits (Khain and others, 1997).

In the Baikal-Vitim area, Vernikovsky and others (2004) describe fragmented island-arc rocks (dated at about 900 to 812 Ma) and ophiolites thrust over Neoproterozoic strata. These rocks are described earlier in an area referred to as the Patom highlands.

In the Yenisey Ridge area, Vernikosky and others (2003) have outlined a complex history of emplacement of granitic rocks from 880 to 860 Ma on a continent or microcontinent, perhaps outside of the Siberian craton. They have also described a collision of this continent or microcontinent with the Siberian continent and the emplacement of syn-collisional 760 to 720 Ma granitic rocks. Magmatic-arc rocks and associated ophiolites formed from 760 to 720 Ma and were thrust onto the Siberian continental margin.

The Taymyr area (Vernikovsky, 1995; Vernikovsky and Vernikovskaya, 2001; Vernikovsky and others, 1998) is an accretionary belt composed of 900 to 850 Ma granite-gneiss terranes, ophiolite dated at 740 to 720 Ma, carbonate rocks, and other sedimentary rocks, all accreted to the Neoproterozoic margin of the Siberian continent.

Kolyma

The relatively small Kolyma block lies in the far northeast of Russia. It consists of Lower and Middle Proterozoic basement rocks partly encircled by Neoproterozoic sedimentary rocks, as well as Paleozoic, Mesozoic, and Cenozoic sedimentary and volcanic rocks (Abramovich and others, 1999; Natapov and others, 1978; Zonenshain and others, 1988).

Baltica

The Neoproterozoic continental margins of Baltica vary significantly in character from region to region. The eastern border of Baltica in the Ural Mountains is characterized by thick miogeoclinal successions of Mesoproterozoic and Neoproterozoic sedimentary rocks (Glasmacher and others, 2001; Ivanov and others, 1986; Maslov, 2004; Nikishin and others, 1996; Puchkov, 1997; Willner and others, 2001). These strata form east-facing miogeoclinal-margin deposits extended along the length of the Ural Mountains (Maslov, 2004). In the southern Urals, these deposits are as old as 1,635 Ma and, although the succession is interrupted by unconformities, the margin appears to have been miogeoclinal from the Mesoproterozoic to the latest Neoproterozoic. The deposition of these long-lasting miogeoclinal deposits can plausibly be considered to have been initiated by rifting prior to 1,635 Ma. Northward, miogeoclinal deposits as old as 1,000 Ma are recognized in the middle Ural Mountains and as old as 640 Ma in the northern Ural Mountains. Apparently, the initiation of deposition was younger to the north.

Northward along the eastern margin of Baltica, the continental-margin deposits bifurcate into a western belt, the Taminian belt, and an eastern belt, which is a continuation of the subpolar and polar Urals. The bifurcation is unusual, but perhaps was produced by the juxtaposition and, in part, amalgamation of Baltica and Siberia. In this speculation, the subpolar and polar Urals and Novaya Zemlya (Korago and others, 2004) continue through Arctic Novaya Zemlya (Korago and others, 2004) into northwestern Siberia (Gee and Teben'kov, 2004), forming the western boundary of Siberia. The Ural Mountains and the Taminian belt, in this reconstruction, are the northern and eastern boundary of Baltica. In this configuration the central part of the Ural Mountains would be a join between the two continents, and Novaya Zemlya may represent continental margin deposits detached from the margin of Siberia and then deformed.

The northern and western margins of Baltica form orthogonal segments—an east-west-trending segment (the Taminian margin on the east) and a north-south-trending segment (the Baltoscandian margin) on the west. The orthogonal shape of the northern and western margins of Baltica is considered (see the discussion in Siedlecka and others, 2004) to be two arms of a triple junction, the third arm of which is in an oceanic domain.

The Taminian belt consists of Neoproterozoic miogeoclinal-margin deposits that extend from the northern margin of Baltica in Russia northwestward in fragmentary outcrops to the Varanger Peninsula in Norway (Bogolepova and Gee, 2004; Gee and Teben'kov, 2004; Maslov, 2004; Roberts and Siedlecka, 2002; Roberts and others, 2004; Siedlecka and others, 1989, 2004). The Taminian margin is divided into two successions (Siedlecka and others, 2004): a lower one, as thick as 9,000 m, composed of submarine-fan turbidites grading upward into deltaic, coastal, and fluvial deposits and an upper succession, as thick as 6,000 m, which consists of shallow marine and subordinate fluvial deposits. The Neoproterozoic rocks on the Taminian margin can in places be demonstrated to thicken outward from the craton (Hambrey and Harland, 1981; Siedlecka and others, 2004), suggesting that the margin is a north-facing miogeocline.

The age of the strata in the Taminian belt is not well defined. Much of the belt is considered to be Late Riphean to Vendian (1,000 to 540 Ma) in age (Siedlecka and others, 2004), but, on the Varanger Peninsula and environs in northwestern Baltica, glaciogenetic deposits indicate ages mostly in the range from 650 to 630 Ma (Evans, 2000; Siedlecka and others, 2004). The Taminian continental margin is considered to have formed by rifting, as is indicated by the thick, north-facing, miogeoclinal deposits (Siedlecka and others, 2004). The age of this rifting is not precisely known but is considered to be Middle Riphean (1,350 Ma to 1,000 Ma) or Upper Riphean (1,000 to 650 Ma). Rifting around Baltica is poorly constrained, due to the uncertainty of available age dates. The Ural Mountains experienced rifting as old as 1,635 Ma, whereas (as is explicated below) the Baltoscandian margin was rifted about 700 to 500 Ma. How the Taminian belt fits with these disparate ages of rifting is unclear. In Russia, northeast of the Taminian belt in the subsurface of the Pehora Basin, is a large area of continental magmatic-arc rocks that have been dated at 618 to 551 Ma (Dovzhikova and others, 2004; Pease and others, 2004).

In Baltoscandia, Neoproterozoic rocks occur in complex nappe structures and consist of detrital and carbonate rocks, as well as distinctive diamictites interpreted as tills (Bockelie and Nystuen, 1985; Foyn, 1985; Hambrey and Harland, 1981; Kumpulainen and Nystuen, 1985; Roberts and Siedlecka, 2002; Siedlecka and others, 2004; Stephens and Gee, 1985; Vidal and Moczydlowska, 1995; Winchester, 1988). Three types of basins formed during the progressive breakup of the western margin of Baltica that led to the formation of a miogeoclinal margin (Siedlecka and others, 2004). The most distinctive of these basins is the third type, which occurs in the more outboard parts of the Baltoscandian margin. This type contains voluminous magmatic rocks, including mafic-dike swarms and ultramafic rocks. A minimum age on a dike at one locality is 608 Ma (Svenningsen, 2001), which is interpreted to mark the onset of seafloor spreading in the Iapetus Ocean. Elsewhere, these magmatic rocks have been dated in the general range of 700 to 530 Ma (Siedlecka and others, 2004). Paulsson and Andréasson (2002) indicate an age of 850 Ma for the attempted breakup of Rodinia in Scandinavia, whereas Greiling and others (1999) alternatively noted a transition from continental rifting to ocean-floor formation at about 600 Ma.

Neoproterozoic rocks are well defined in areas of the North Atlantic region including Britain, Ireland, Scotland, Greenland, and Svalbard (Dalziel and Soper, 2001; Dewey and Shackleton, 1984; Fairchild and Hambrey, 1995; Hambrey and others, 1991; Harland, 1985; Harland and others, 1997; Kelling and others, 1985; McCay and others, 2006; Soper, 1994; Strachan and Holdsworth, 2000a, b; Winchester, 1988). These rocks consist of thick accumulations (commonly 5 km to as much as 25 km thick) of predominantly shallow-water clastic rocks. The successions all contain diamictite interpreted to be glaciogenic. The rocks have been interpreted as rift-basin deposits, although the thickness and the presence of diamictite is suggestive of continental margin deposits. The successions may originally have been part of the continental margins of Laurentia or Baltica, both of which contain diamictite deposits. Subsequently, these deposits appear to have been fragmented and transported into their present positions.

The east-southeast margin of Baltica is the Trans-European Suture Zone, which locally corresponds with the Tornquist line or Teisseyre-Tornquist line (Belka and others, 2002; Bula and others, 1997; Dadlez, 2000; Krolikowski, 2006; Moczydlowska, 1997; Pharaoh and others, 1997; Poprawa and others, 1999; Savov and others, 2001; Strauss and others, 1997; Winchester and others, 2002). The suture zone is a lithospheric boundary between thick, older lithosphere on the northeast and thin, younger lithosphere on the southwest (Pharaoh and others, 1997). Neoproterozoic rocks to the northeast of the Trans-European Suture Zone are mainly platformal sedimentary rocks, whereas the Neoproterozoic rocks southwest of the suture zone are magmatic-arc rocks generally consisting of Cadomian and Avalonian terranes. Although the location of the southwest boundary of the Neoproterozoic Baltica margin is well defined on a regional scale, the history of the margin is not clear. Most of the Neoproterozoic continents described above are circumscribed by miogeoclinal-margin deposits, but such deposits are not apparent along the southwest margin of Baltica. Possible continental-margin deposits occur in the Lublin slope (Strauss and others, 1997; Vidal and Moczydlowska, 1995), where volcanic and sedimentary rocks of Neoproterozoic age are present. These rocks are relatively thin, however, and are present only in a relatively small part of the margin. Strata in the late Neoproterozoic and Early Paleozoic Baltic Basin thicken toward the suture zone and along the suture zone, but the strata do not extend beyond the margins of the basin (Garetskiy, 1982; Poprawa and others, 1999). High rates of subsidence in the Baltic Basin in Late Vendian to earliest Cambrian time (about 580 to 540 Ma) suggest extension and possible rifting (Poprawa and others, 1999). On a broader scale, Bogdanova and others (2003) have speculated that Baltica was surrounded by an ocean at about 700 Ma. The trend of the Trans-European Suture Zone is linear or slightly curved, leading to speculation and controversy about strike-slip movement (Dadlez, 2000). Many aulacogens, identified in the subsurface of Baltica (Vidal and Moczydlowska, 1995), and associated sediments are poorly dated but considered to be 800 to 700 Ma in age (Lobkovsky and others, 1996; Shpunt, 1988; Vidal and Moczydlowska, 1995).

Cadomian, Avalonian, and related magmatic rocks

Magmatic-arc rocks, generally referred to as Cadomian or Avalonian, occur in scattered outcrops in a diffuse belt from Turkey on the east to the southeastern United States and Yucatan on the west (Plate 1) (Nance and Murphy, 1994; Strachan and Holdsworth, 2000a, b). What are now fragments in this belt have undergone closer packing in the Neoproterozoic (Murphy and others, 2002; Nance and Murphy, 1994; Nance and others, 2002) and appear to have wrapped around northern Africa and northern South America. In the Neoproterozoic, outcrops in the southeastern United States and Yucatan may have been a continuation of this belt and, if so, provide a longitudinal tie between Africa, South America, southwestern United States, and Yucatan. Alternately, the magmatic-arc rocks are far-traveled blocks that do not present a simple pattern.

Scattered areas of polymetamorphosed and structurally complex rocks in Turkey are considered to be arc-related rocks (Neubauer, 2002; Sengoer and others, 1984; Ustaömer and others, 2005) dated in northwestern Turkey at 576 to 565 Ma and in western Turkey at 660 to 520 Ma. Arc-related rocks are present in scattered outcrops in Europe, including Saxo-Thuringia, Germany, where they are dated at 570 to 540 Ma (Linnemann and Romer, 2002), in the Teplá-Barrand of the Czech Republic, dated at 609 to 522 Ma (Doerr and others, 2002), in the Helvetic and Penninic basement of the western Alps, dated at 546 to 500 Ma (Neubauer, 2002), in the Pennine basement of the eastern Alps, dated at 657 to 482 Ma (Neubauer, 2002), in the Austroalpine and Southalpine of the eastern Alps, dated at 609 to 477 Ma (Neubauer, 2002), in the southern Carpathians and Serbo-Macedonian massif, dated at 777 to 545 Ma (Neubauer, 2002), in the Ossa-Morena zone of Iberia, dated at 620 to 480 Ma (Bandres and others, 2002), in northwestern France, dated at 610 to 573 Ma (Inglis and others, 2005b), and in north Wales and southeastern Ireland, dated at 650 to 550 Ma (Strachan and Holdsworth, 2000b).

Methods of Meso-Neoproterozoic Reconstructions

Different methods of reconstructing Rodinia have been proposed, most involving the fit of continents based on (1) matching structural belts (for example, Moores, 1991; Karlstrom and others, 1999; Hoffman, 1991), (2) similar lithologic-stratigraphic character of presumed conjugate blocks (Dalziel, 1999; Sears and Price, 2004), and (3) matches of Proterozoic belts (Karlstrom and others, 1999; Burrett and Berry, 2000). Other models use paleomagnetic information combined with lithologic correlation (Pisaresky and others, 2003; Li and others, 2008). The approach here is different, although it has a basis in many proposed assemblages based on the fit of continents. The model stresses the similarity in shape of conjugate margins, similarities that for many continents are remarkable. A large part of the hypothetical supercontinent of Rodinia fits together much like a giant jigsaw puzzle.

Although the matches are not always tightly constrained, enough information is provided by the sparse fragments of Meso-Neoproterozoic rocks to evidence a proposed reconstruction. In places, Baltica and Siberia for example, the jigsaw approach is not viable, apparently because the margins of these continents have been modified by erosion and tectonic deformation that preclude a close fit. In these areas of uncertain fit, other types of geologic information (stratigraphy, structure, geologic history, paleomagnetics) are used to define the correlation between continents.

Jigsaw puzzle method

The jigsaw puzzle method is a highly useful procedure in determining the relative positions of continents. If the continents fit together tightly, then the relative positions of the pieces is apparent. However the usefulness of the method is limited. It only determines the final assembly pattern of the pieces of the puzzle, namely what the jigsaw puzzle looks like when it is finally assembled. The method provides neither a means for determining the positions of the puzzle pieces when they are dispersed nor their movement in space and time.

Southern continents assembly

The Southern Continents (South America, Africa, India, Arabia, Australia, Antarctica) illustrate the method used here for assembly of continents. The six continents form a large part of the supercontinent of Rodinia, as well as the younger supercontinent of Gondwana. The reconstruction follows the model (LS) of Lawver and Scotese (1987) and the model (SH) of Smith and Harlam (1989). The LS and SH models show a remarkably tight fit of the six continents (fig. 1). The reconstruction is so tight that it is almost certainly a breakup pattern, rather than a convergent pattern. In a convergent pattern the shapes of the constituent margins would have been modified before or during assembly and would not allow a tight jigsaw-puzzle reconstruction. The breakup pattern is similar to reconstructing a broken dinner plate. The pieces fit tightly together because the pieces were initially together. A reconstruction of the broken dinner plate uses the jigsaw-puzzle method, whereby the pieces are reassembled using the shapes of continents. In these procedures, the jigsaw method produces an assemblage (the LS and SH models) that shows a breakup pattern. In the model considered here, the LS and SH models show an actual situation (the initial shape of the continent), whereas the jigsaw-puzzle method is a human device used to determine the initial shape of the assembly of continents. The jigsaw puzzle does not show a concatenated series of events but only the reassembly of continents into their original form. The assembly is essentially a virtual reconstruction that is extremely useful in indicating what the original assembled continents looked like (the dinner plate) but does not provide information on the distribution of continents after breakup or information on the relative positions of continents after they disperse.

Age of southern-continent assembly

The age of assembly of the southern continents is critical in the interpretations presented here. A study by du Toit (1937) indicates a Paleozoic age based on matching the grain of Precambrian and Cambrian rocks. Both LS and SH indicate that the southern continents are part of Gondwana and, thus, imply that the reconstruction is Mesozoic in age. However the widespread distribution of Proterozoic rocks in the southern continents (Plate 2) indicates that the continents also include a history older than Mesozoic, as indicated by the presence of rocks of Proterozoic age (fig. 1). This older history raises the possibility that the reconstructions of LS and SH indicate a Proterozoic assembly in addition to a Mesozoic assembly. Such a dual interpretation is possible because the age of assembly itself cannot be determined except by the ages of fragmentary circumscribing rocks. Considering these relations, the age of the assembly shown by LS and SH as used in this report is Proterozoic, although I realize that it also has a Mesozoic history, as implied by LS and SH. The dual age of the assembly of the rocks of the southern continents is possible, because both assemblies are constructed from the same continents, which, as described later, maintain remarkably similar shapes through time and could, thus, generate similar map patterns if reassembled at different times.

Complexities

Complexities in the jigsaw-puzzle reconstructions are evident in Plate 2, which shows that South America is broken into three tectonic sub-blocks (Amazonia, São Francisco, and Rio de La Plata) and that Africa is broken into five tectonic sub-blocks (West Africa, Trans-Sahara, Congo, Kalahari, and east Africa). As evident on Plate 2, these sub-blocks in South America can be reassembled to form part of the major continents of South America and Africa (the continents as used in the LS-SH models). The sub-blocks indicate rifting of the assembly as used by LS-SH but not rifting severe enough to significantly change the outline of South America and Africa.

Similar shape of southern-continent margins through time

The six continents maintained remarkably consistent shapes throughout much of the Proterozoic and Phanerozoic. This relation is supported by the reconstructions of LS and SH (fig. 1) that use present-day outlines of continents to tightly reconstruct assemblies that are considered here to be Meso-Neoproterozoic or younger in age**.** In this proposal, the continents between the present and the Neoproterozoic had similar shapes. Such a relation would not be possible if these continents had undergone major deformation or erosion that altered the shape of the continents and, thus, precluded precise jigsaw-puzzle reconstructions. If the shape of continents remains similar through time, the jigsaw-puzzle reconstructions would always lead back to the LS-SH model. Thus, a jigsaw reconstruction using the six continents would look similar for any time frame from Mesoproterozoic to the present, because the shape of the continents remains similar in this time frame. This relation leads to the conclusion that major continents, such as those of the southern continents, once formed, are stable blocks that are difficult to modify, at least on a lithospheric scale.

Neoproterozoic Reconstructions

A wide variety of models have been proposed for the worldwide distribution of major tectonic plates during the Mesoproterozoic and Neoproterozoic. Most models show that Mesoproterozoic plates were joined to form a supercontinent (Bond and others, 1984; Condie, 2003, 2004; Rogers and Santosh, 2003, 2004; Unrug, 1997). Most geologists refer to this supercontinent as Rodinia, which is thought to have assembled in the Mesoproterozoic and to have broken up in the Neoproterozoic. This breakup is indicated by the predominance of Neoproterozoic extensional margins (ca. 870 to 740 Ma), described here and by Condie (2002). The extensional margins apparently require the fragmentation and drifting of continents away from a supercontinent. Other models propose a rather loose fit of the continents or even widely dispersed continental plates (Cordani and others, 2003a; Meert and Torsvik, 2003; Pesonen and others, 2003). Most

Mesoproterozoic and Neoproterozoic plate reconstructions use orogenic fold belts to guide the assembly of continents. That method differs from the one used here, which uses the distribution of sedimentary and igneous rocks to define margins. This approach appears to be appropriate, because the dating of orogenic belts is not always clear. In addition, much of the middle and late Proterozoic is a time of extension, and boundaries of continents are better defined by extension-related sedimentary rocks than by compressional orogenic belts.

The shape and history of continents and continental margins is critical in the model proposed here. The continents are the major tectonic blocks of the Neoproterozoic, and their margins reveal histories of continental separation, multiple times of rifting, development of miogeoclinal margins, development of aulacogens, magmatic-arc terranes related to subduction, and continental sutures. Ideally, the margins of a continent are recognized by the presence of sedimentary and associated igneous rocks circumscribing the continent. Recognition of continental margins is enhanced, where found, by the presence of glaciogenic diamictite, which characteristically follows the trend of the continental margin, perhaps because it is more likely to be preserved there than in the center of a continent. The continental margins typically are miogeoclinal and preserve shallow-water miogeoclinal-margin sedimentary rocks that thicken away from the continent. These Neoproterozoic margins, or reactivated margins, are most evident surrounding Laurentia, in western Baltica, circumscribing Siberia, in China, in the Himalayan Mountains, in Iran, in East Africa, in eastern South America, and from sparse information in western South America (Plates 1, 2).

Many presumed continental margins are not characterized by these circumscribing major outcrops of sedimentary or igneous rocks, which are a defining feature of many known Neoproterozoic margins. Most of these margins were probably the sites of deposition of Neoproterozoic sedimentary and igneous rocks, as described earlier (see "Stripped and Covered Margins" under "Recognition and Classification of Continental Margins"), but have been stripped away or buried at some time since their deposition by either subareal or subduction erosion.

In the model presented here (in the general timeframe of 1,200 to 850 Ma), only major continents are presented. In particular, the relatively small Kolyma or Tarim blocks of Siberia and East Asia are not shown, nor are probable widespread microcontinents, ribbon continents, or accreted terranes. These are important localities but difficult to place in our assembly and to use in major plate-tectonic reconstructions. The smaller blocks nevertheless provide insight into what likely lay between some of the major blocks.

Construction of the model presented here is based primarily on geologic information enhanced in places by the paleomagnetic information. Once constructed, the model was checked against the paleomagnetic data (described below) to determine if the geologic construction is compatible with paleomagnetic data. The proposed models are described and compared with other models. The model proposed here is shown in Figure 2 and described in detail below.

Reconstructing the southern continents region

The southern Neoproterozoic continents form a widespread megablock composed of the following continents: South America (including Amazonia, Rio de la Plata, and

Saõ Francisco subcontinents), Africa (including West Africa, Trans-Sahara, Congo, Kalahari, and East Africa subcontinents), India, China, Australia, and Antarctica. This assemblage is similar to that of Mesozoic Gondwana (Lawver and Scotese, 1987; Smith and Hallam, 1970). South America and Africa are complex continents seemingly somewhat cohesive, but that contain intracontinental cratons showing partially defined boundaries that could represent rift zones and potential ocean sites. The cohesion of the South American continent is suggested, as described previously, by the presence of Grenville-age rocks (1,200 to 900 Ma) that follow the western margin of South America (Plate 2). These rocks coincide with subsequent Neoproterozoic and younger margins (Plate 2), although the Neoproterozoic rocks have been largely stripped away. Along the eastern side of South America, Neoproterozoic miogeoclinal-margin sedimentary deposits (about 700 to 650 Ma), magmatic-arc rocks (900 to 780 Ma), and Pan-African metamorphic/granitoid rocks (about 750 to 500 Ma) define a Neoproterozoic margin and indicate cohesion of at least a part of the continent. On the western side of Africa extending from northern Angola to the Cape of Good Hope, are miogeoclinal-margin deposits (about 700 to 650 Ma) and an inland-extending aulacogen (Damara belt). The miogeoclinal-margin deposits resemble similar deposits along the west coast of Laurentia, which are considered to have formed along a rift margin, and the African rocks seem to have formed in a similar tectonic setting. The shape of the Neoproterozoic continental margins along the eastern margin of South America and along the western margin of Africa are remarkably similar, strongly indicating that the Neoproterozoic continents were once joined. A Wilson cycle of opening and closing of an ocean is evident (Plate 3).

The breakup of South America and Africa may have occurred before 700 to 635 Ma, the most likely age of glaciogenic diamictite in post-rift miogeoclinal-margin deposits of the Saõ Francisco craton of northeastern South America. The continents presumably drifted apart, but drift was followed by or overlapped in age with the convergence of Africa and South America along a zone of magmatic-arc rocks in eastern South America. Associated orogenic events are dated at 790, 730 to 700, 640 to 620, and 600 Ma (da Silva and others, 2005). The other southern continents (India, Antarctica, Australia) are assembled in a fashion similar to the Mesozoic reconstruction of Lawver and Scotese (1987) and Fitzsimons (2000b, 2003). This reconstruction, too, shows a close match in the shape of the continental margins, a match so regular that the pattern was most likely to have formed during plate breakup rather than by plate assembly, when margins would be expected to be more complex, owing to the juxtaposing of margins of at least somewhat different shapes. The model is also supported by the correlation of rocks from one continent to another (Fitzsimons, 2000b, 2003; Unrug, 1996), as indicated by the distribution of Precambrian basement rocks and the distribution of high-grade metamorphic rocks/granitoids of the Pan-African tectonothermal event. The grouping of these continents is consistent with the idea presented by Rogers (1996) that Australia, Antarctica, India, Madagascar, and southern Africa were all close together at about 3,000 Ma (fig. 3), a judgment based on the age of extensive supercrustal sequences.

The southern continents in the Meso-Neoproterozoic consist of various tectonic blocks. West Africa, which contains rocks of the Pan-African tectonothermal event, seems logically placed west of northern Africa. Arabia contains extensive magmatic-arc rocks in its southern part that are a continuation of the magmatic-arc rocks of the East

African orogenic belt. Arabia's position in the northern part of the southern-continents region seems justified. South China and North China appear to have been joined in the Neoproterozoic along the Quinling Belt and are placed along the east margin of the southern continents region.

The region referred to here as the southern-continents region is considered to have been assembled into a tight-fitting block during the Meso-Neoproterozoic. This reconstruction is based largely on the distribution of Grenville-age rocks (1,200 to 900 Ma), which suggests a link between Africa, India, Antarctica, and Australia (fig. 2). Unrug (1997), alternatively, has proposed that the southern continents were not assembled until the Neoproterozoic, largely because the Neoproterozoic rocks indicate collision and orogeny during the Pan-African orogen (Kroener and Stern, 2005). The reconstruction of the southern continents shown here is largely the same as that described by Unrug (1996, 1997) at 500 Ma, but his reconstruction at 1,000 to 700 Ma shows a quite different pattern, with blocks widely scattered around Laurentia. The interpretation presented here holds that southern Rodinia fragmented at 870 to 750 Ma and that it was reassembled at 500 Ma with, perhaps, a somewhat similar shape. The most important difference regarding the southern continents region during the Neoproterozoic and that of most others (among which those of Moores, 1991; Karlstrom and others, 1999; Burrett and Berry, 2000 figure prominently) is that certain blocks, particularly Antarctica, India, and Australia are part of southern Rodinia, whereas most others contraveningly place those blocks adjacent to the western United States.

Continental margins shown in this report mostly occur in conjugate pairs. An exception, as shown in Figure 2, is Antarctica and Australia. These continents may have originally been joined as part of the southern continents assemblage that was flanked on the south by now largely fragmented continental fragments dated as 1,000 to 1,500 Ma by Sm-Nd model ages and Re-Os systematics ages in the Antarctica Peninsula, Thurston Island, Marie Byrd Land, and New Zealand (Handler and others, 2003).

Reconstructing Laurasia

The main tectonic blocks in Laurasia are North America, Greenland, Baltica, Kazakhstan, Siberia, and China. Many microcontinents and ophiolitic rocks are present in the East Asia orogenic belt (Plate 2). The Laurasia blocks are separated in western areas from the southern-continents region by the Avalonian-Cadomian belt of magmatic-arc rocks.

In the Neoproterozoic, the current north Atlantic region is characterized by continent-fringing miogeoclinal-margin deposits, or detached deposits. These are evident in eastern North America (Bird and Dewey, 1970; Dewey, 1974; Dewey and Shackleton, 1984; Rankin and others, 1989), in Greenland (Winchester, 1988), in Svalbard (Harland, 1985; Harland and others, 1997; Soper, 1994), in Scandinavia (Bockelie and Nystuen, 1985; Kumpulainen and Nystuen, 1985; Stephens and Gee, 1985; Vidal and Moczydlowska, 1995), and in Britain and Ireland (Bird and Dewey, 1970; Dewey, 1974; Dewey and Shackleton, 1984; Kelling and others, 1985; Strachan and Holdsworth, 2000a, b). These sedimentary deposits range in age from about 740 to 580 Ma, a judgment based mainly on the identification of Sturtian (about 700 Ma), Marinoan (635 Ma), and Gaskiers (580 Ma) glacial deposits (Evans, 2000; McCay and others, 2006) and associated isotopically dated igneous rocks that are considered to have formed along Neoproterozoic rift margins during the opening of the Iapetus ocean. These relations suggest that Baltica and Laurentia were joined in pre-rift Mesoproterozoic time (Dewey, 1974). This is compatible with the interpretation that Mesoproterozoic igneous and metamorphic rocks were joined across the north Atlantic region (Romer, 1996).

A particularly important question is whether the Cadomian-Avalonian magmatic-arc rocks formed outboard of the depositional belts of Neoproterozoic rocks or after the development of these fringing miogeoclinal-margin deposits. Most interpretations consider the Cadomian-Avalonian rocks to be far-traveled and older than the fringing miogeoclinal-margin deposits, mainly because isotopic dates on the magmatic-arc rocks are older than those on the continental-margin rocks and, thus, could not have originally formed outboard of the younger miogeoclinal-margin deposits. On the other hand, the initiation of rifting in the southeastern United States is considered to be about 760 to 740 Ma (Cawood and others, 2001; Su and others, 1994) and, thus, is compatible with a magmatic arc outboard of an older fringing miogeoclinal margin. However, in the northeastern United States and western Canada, the igneous rocks interpreted to be related to rifting are about 500 to 600 Ma, an age close to, or older than, that indicated for the miogeoclinal-margin deposits. As described previously, continental margins commonly have a complex history involving several episodes of rifting. Thus the younger belt of igneous rocks in the northeastern United States and eastern Canada could reflect a reactivation of a margin that originally formed along a preexisting 760 to 700 Ma rift (Cawood and others, 2001), in which case the 600 to 500 Ma magmatic-arc rocks could have formed outboard of the rift margin and, thus, represent the closing of the Iapetus Ocean in the Neoproterozoic or early Cambrian.

The position of Siberia in Neoproterozoic Eurasia is debated. It has been considered to lie along the western margin of Laurentia (Sears and Price, 2000, 2003; Sears and others, 2004), along the northern margin of Baltica (Condie and Rosen, 1994), or along the eastern margin of Laurentia. As noted previously, Siberia may have been a separate block in the Neoproterozoic, perhaps rifted from neighbors in the Mesoproterozoic.

An alternative idea, presented here and described in the discussion of Baltica above, is that the Baltica and Siberian margins touched in the Neoproterozoic, that the margin of eastern and northern Baltica was along the Ural Mountains, extending into the Taminian belt, and that the margin of Siberia followed the Ural Mountains and the Novaya Zemlya belt into northern Siberia. If these speculations are correct, Siberia lay east of Baltica in the Neoproterozoic.

Relative positions of Laurasia and the southern continents region

The conjecture that Eurasia and the southern-continents region were both somewhat coherent megablocks in the Neoproterozoic after fragmentation of Rodinia is critical to the model reconstruction proposed here that is partly based on a jigsaw-puzzle assembly of the Rodinia continental fragments. In this report, Eurasia and the southern continents are joined in such a way that South America lies east of North America and the other continents are assembled as shown in Figure 2. Positioning Eurasia east of South America is based on the presence of Cambrian rocks in the Argentina precordillera, rocks that have faunal affinities to rocks in the southwestern United States (Astini, 1998; Keller, 1999). Although the Cambrian age of these rocks is outside of the Neoproterozoic time frame of this report, the uniqueness of this faunal tie appears to unite Argentina and the southwestern United States in the Cambrian and thus perhaps in the Neoproterozoic as well. In addition, Grenville-age rocks along the eastern margin of North America may have been contiguous with Grenville-age rocks along the western side of South America.

Composite Rodinian reconstruction

The composite model of Mesoproterozoic Rodinia and its Neoproterozoic fragmentation, shown in Figure 2, is based on the discussion of the geologic features presented in Plates 1 and 2 for both the southern-continents region and Laurentia. Major continents are considered to have assembled during Grenville-age convergence, continental collision, and igneous activity into a tight fitting Grenville-age supercontinent. The supercontinent then fragmented in the Neoproterozoic along boundaries that became Neoproterozoic continental margins. Extension allowed the opening of ocean basins between the continents, and contraction in the late Neoproterozoic produced subduction zones marked by magmatic arcs.

Comparisons with other models

A variety of models has been proposed for the configuration of the supercontinent of Rodinia. Eleven of these are shown in Figures 4 to 13. These models were selected to show diversity, as well as the most commonly accepted types of models. Some models indicate that Rodinia was not a coherent assemblage of continents but rather that the continents were dispersed (Cordani and others, 2003a; Meert and Torsvik, 2003; Pesonen and others, 2003).

This diversity of models illustrates the difficulties confronting the interpretation of Proterozoic geology and the scarcity of definitive tie points or other geologic information that can be used to assemble Rodinia. Paleomagnetic data are difficult to use because reliable data, or even only possibly reliable data, are scarce and difficult to use because of structural dislocations, uncertain age control, and remagnetization. Nevertheless, the paleomagnetic data (discussed here) do appear to be compatible with the general location of major continents in the proposed model for the 1,200 to 850 Ma time frame.

The approach used here is to define Neoproterozoic continental margins, margins considered to reflect the breakup pattern of Rodinia, and to use these continental fragments to reconstruct Mesoproterozoic Rodinia in much the manner of a jigsaw puzzle. The following discussion compares the model used here with those published by others.

The model proposed here (fig. 2) is similar to models proposed by Bond (1984) and Keppie (1992), which are shown here as Figures 4 and 5. The similarity of these models lies in their arranging the continents into a rather tight assemblage in which the relative positions of the continents are similar. In particular, South America is east of, and adjacent to, North America. The location of Siberia, however, varies. It has been placed north of the northern margin of North America (Condie and Rosen, 1994), along the east side of Baltica (Torsvik and others, 1996), or along the western margin of Laurentia (Sears and Price, 2000, 2003; Sears and others, 2004).

Piper's model (Piper, 2000), shown in Figure 6, shows some similarities to the model used here, in particular in the reconstruction of the southern continents. A major difference, however, is that in Piper's model South America is located far from the eastern margin of North America. The models of Dalziel (1992, 2000), Waggoner (1999), and Hoffman (1991) are similar to those shown here as Figures 7, 8, and 13 in placing

South America east of the eastern margin of North America but differ significantly in placing Antarctica and Australia west of the western margin of North America. This is a common reconstruction, described in detail by Moores (1991), Burrett and Berry (2000), and Karlstrom and others (1999). But, as described previously, the location of Antarctica and Australia adjacent to and west of western North America is not used in our reconstruction. Antarctica and Australia fit well together in the construction of the southern continents (Australia, Antarctica, India, Africa). In addition, South America appears to join with Africa along the western side of the southern-continents region (Plate 3). Thus Gondwana appears to be a unified block that is difficult to relocate into a position west of western North America. Rogers (1996) proposed that the continents of Australia, Antarctica, India, and Africa were close together by about 3,000 Ma, when continents first developed in these areas (fig. 3). This ancestry is difficult to reconcile with the idea that Antarctica and Australia, having assembled in the Southern Hemisphere, transported to a position off western North America and then presumably transported back to a position near the Gondwanan join of Australia, Antarctica, Africa, and India. The paleomagnetic reconstruction, although difficult to interpret, appears to indicate a Southern Hemisphere position for Australia, Antarctica, Africa, and India, in contrast to a Northern Hemisphere position, if these continents were adjacent to western North America. Finally, the reconstructions of Sears and Price (2000, 2003) and Sears and others (2004) place Siberia in a position along western North America, a position at odds with placing continents of southern Rodinia in this position.

The model proposed by Waggoner (1999), shown here as Figure 8, is unique in using Ediacaran paleogeography as a guide to the assembly of the continents. In his reconstruction, South America lies east of the eastern margin of North America, a position consistent with other models (described above) that place Australia and Antarctica west of western North America. Biotas of the “Australia-Baltica-north Laurentia-Siberia cluster“ show correlation of biotas that imply that Antarctica and Australia lay west of western North America, a concept argued against previously. Perhaps Waggoner's model is correct, in which case my argument for a Gondwana origin of these biotas is not valid. Alternatively, the correlation of the biotas might not be entirely correct or other locations of continents might also be compatible with the distribution of the biotas.

The reconstruction of Weil and others (1998), shown here as Figure 9, and the reconstruction of Dalziel and others (2000), shown here as Figure 7, are similar in the general distribution of many of the continents, including the position of South America east of North America and the position of East Antarctica and Australia west of western North America.

The model of Pisarevsky and others (2003), shown here as Figure 10, is both similar to and dissimilar to several other models. It is similar to the model proposed here (fig. 2) in placing South America east of North America and in the relative positions of the Laurasian continents Laurentia, Greenland, and Baltica. The model differs in placing Antarctica well south of Laurentia and Australia also south of Laurentia but touching in one small area along southern North America.

Unrug (1997) shows a model for the time frames 1,000 to 700 Ma and 700 to 500 Ma. (figs. 11, 12). His older time frame, shown here as Figure 12, places South America east of Laurentia, as is the case in most models of Rodinia, and the southern continents west of Laurentia, as is also common in other models but considered problematic here as discussed previously. The 700 to 500 Ma model of Unrug (fig. 12) is fairly similar to the model shown here with South America east of Laurentia, and India, Antarctica, and Australia east of Africa. But, the 700 to 500 Ma model accords with our interpretation roughly within the time frame for the breakup of Rodinia, not the time interval of assembly as implied by Unrug's model.

Li and others (2008) present a major synthesis describing a well-constrained configuration of Rodinia based on detailed histories of tectonic blocks combined with paleomagnetic interpretations. Their well-documented interpretations are different in major ways from the model proposed here, because different methods have been used in the reconstructions. Li and others focus mainly on paleomagnetic location of continents combined with matching the geology of continents, whereas the reconstruct of Rodinia used here is mainly by matching continental margins using a jigsaw method of assembly.

Paleomagnetic Studies

The proposed model of Rodinia presented here (fig. 2) is based primarily on geologic evidence of the distribution and shape of Neoproterozoic continents and on the assembly of these continents in such a way as to match the shape and history of conjugate margins. Using paleomagnetic data, an attempt was made to test the validity of the model. However, such tests are difficult because Precambrian rocks can be affected by a number of different problems. Geochronologic studies, for example, of sedimentary sequences must often rely heavily on faunal constraints due to difficulties in directly dating sediments with radiometric methods, and metamorphic rocks typically rely on radioisotopic ages reset by an interval of metamorphism that may not necessarily coincide with that age of remanence. Aside from tenuous age control, two other factors limit the usefulness of existing Precambrian paleomagnetic data in testing paleoreconstructions: the absence of local structural control to establish paleohorizontal and to identify the boundaries of coherent cratonic blocks and uncertainties associated with the timing and stability of remanence.

For these reasons, stringent standards for pole selection (Buchan and others, 2001; Buchan and Halls, 1990) has led to much more restricted, presumably more accurate, pole lists. In most studies, these criteria are based on a system described by Van der Voo (1990) that outlines seven factors by which the quality of poles is characterized. These can be grouped into three categories relating to magnetic remanence reliability, age reliability, and constraints on structure. Magnetic-remanence-reliability criteria assess different methods of remanence determination, tests for remanence stability, and tests controls on the timing of remanence. These require that a sufficient number of samples are used to ensure adequate statistical precision; reliable demagnetization and analysis techniques are applied; field tests are available to constrain the age of magnetization and to assure that the age coincides with the time of remanence acquisition (these include fold, conglomerate, and contact tests); reversals are present in the stratigraphy to assure enough time has elapsed to average secular variation of the field and to provide a reversal-test for establishing antipodality in order to preclude effects of unrecognized overprints; and poles show no resemblance to younger paleopoles, to rule-out remagnetization. Unfortunately, Precambrian results often do not incorporate many of these techniques due to a lack of sufficient exposed section, a lack of paleomagnetic data with which to compare, and large uncertainties associated with the age of remanence. Consequently, methods that are routinely applied to reconstructions for younger times are difficult to apply to the Precambrian.

A common practice in Phanerozoic reconstructions is to compare multiple poles that form well-defined Apparent Polar Wander Path (APWP) segments from two continents. Unlike individual paleomagnetic poles that yield only paleolatitude constraints, this method provides a unique reconstruction of the fit of fragments of a once-joined supercontinent. Due to the paucity of data and to uncertainties in the age and locations of existing poles for the Paleoproterozoic (Buchan and others, 2001; Meert, 2001; Powell and others, 1993), few continents (Laurasia and Baltica) yield substantially long APWP segments to apply this technique to Rodina reconstructions. Indeed, it is not clear if existing lists are complete enough (or consist of sufficient high-quality data) that they might, on their own, yield unique reconstructions for this time period. Pisarevsky and others (2003), for example, indicate that existing data for Mesoproterozoic and Neoproterozoic rocks are insufficient to provide robust reconstructions, except for Laurentia and Baltica. As a result, many Proterozoic reconstructions that have made use of paleomagnetic data to assess geologic-based reconstructions have used those data, in a null sense, simply to exclude certain assemblies.

It is likely, with the growing database of high-quality paleomagnetic poles, that paleomagnetic data will soon provide reliable, independent Neoproterozoic paleoassemblies. However, because there is no consensus as to which lists form the most complete and reliable compilations, we consider discerning between existing data lists to be beyond the scope of this report. As a result, the approach taken here differs from other approaches that rely on highly selected pole lists to fit continents into a paleomagnetically defined paleoreconstruction. Instead, four commonly cited pole lists are combined with the intent that a large dataset, spanning multiple compilations, may reveal broad trends and minimize the influence of outliers that can have a significant influence due to the general paucity of data for the Paleoproterozoic.

The paleomagnetic poles of Meert and Torsvik (2003), Weil and others (1998), Pisarevsky and others (2003), and D'Agrella-Filho and others (1998) are grouped to yield the composite pole list used in testing the model (Table 1). These data are divided into three age ranges (1,200 to 850 Ma, 850 to 760 Ma, and 760 to 500 Ma) corresponding roughly to (1) the interpreted time of the assemblage of Rodinia, (2) the breakup of Rodinia, and (3) plate convergence as indicated by the widespread occurrence of magmatic arcs. Within these large time frames, relative movement of tectonic plates has doubtless taken place, accounting for at least some of the ambiguity of the results. Nevertheless, the data appear roughly compatible with our Neoproterozoic assembly for the older time frame (1,200 to 850 Ma). The two younger time frames are ambiguous, however, suggesting that they correspond with the time of breakup of the supercontinent and dispersal of its cratonic fragments. Throughout the discussion presented below, the reader should keep in mind the sparsity of paleomagnetic data for Southern Hemisphere continents (Table 1) and, in many cases, the uncertain quality of these data.

Due to the lack of sufficiently long APWP paths for most Rodinia continents, a "shotgun approach" is used here that compares scattered paleopoles from one continent with scattered paleopoles from the reference path for roughly the same time period to qualitatively assess the feasibility of individual continental fits of the model. Because

Laurentia is the physically largest of the Rodinian continents and has also the largest number of Rodinia-age paleopoles, the Laurentian pole list has become the reference against which paleopoles from other cratons are compared. The continents and their paleomagnetic poles are, therefore, rotated into a present-day Laurentian reference frame. The Euler poles, and corresponding angles of rotation (about which continental blocks and their poles were rotated) used to construct our model are given in Table 2.

Paleomagnetic data for each of the Rodinia continents (Amazonia, Antarctica, Australia, Baltica, Congo, Kalahari, Siberia) for the period 1,200-850 Ma (excluding possibly unreliable paleopoles for India and Saõ Francisco) are plotted against the APWP for Laurentia (consisting of 72 paleopoles, Table 1, Figure 15) for the same time frame. These continents constitute a majority of the primary tectonic blocks in Rodinia, but the number of paleopoles for each continent varies and is generally small, ranging from one (Amazonia) to 21 (Baltica) (Table 1). An assessment of the reconstruction of the Rodinian continents based on paleomagnetic data is inhibited by these limited sets of paleopoles and by the uncertain reliability of data. Indeed, Meert and Torsvik (2003) note that the Neoproterozoic paleopoles for India are scattered and may not be usable in paleomagnetic reconstructions, as are the scattered paleopoles for Saõ Francisco. Furthermore, although existing data allow us to test the fit of most of the major Rodinian continents, many of the cratonic blocks that form our model (for example, Greenland, North China, South China, Siberia, Rio de la Plata, and west Africa for the timeframe 1,200-850 Ma) remain untested paleomagnetically.

Nevertheless, the available data do suggest that the Rodinia reconstruction proposed here is valid for many of the continents in the 1,200 Ma to 850 Ma time frame. The reconstruction for that interval shows that the Rodinia continents may be tightly packed and supports the general reconstruction presented here (fig. 2). An assessment of the 850 to 670 Ma and 670 to 500 Ma time frames, corresponding to times of continental breakup and presumed dispersal of Rodinian fragments, was not as successful as for the older time frame, partly due to the fact that most of the paleomagnetic data for the two younger time frames were not sufficient and (or) reliable enough to adequately constrain the locations of continental blocks.

Summary

This report identifies and describes middle and upper Neoproterozoic continents and continental margins with the purpose of obtaining a better understanding of Neoproterozoic geologic history, particularly the assembly and breakup of the Rodinian supercontinent. The margins include extensional rift margins, miogeoclinal margins, margins involving Wilson cycles of opening and closing of oceans, margins containing aulacogens (a failed arm of a three-armed rift), reactivated margins, active margins containing subduction-related magmatic rocks, and collisional margins related to plate sutures. These features are presented in two maps (Plates 1 and 2).

Some margins, which appear to have formed by rifting as early as 1,600 Ma, had a long history, including those of Siberia and western North America, which resulted from rifting as old as 1,450 Ma. These margins are considered to be fundamental tectonic boundaries at the interface of oceanic and continental crust and to have remained miogeoclinal, except for extension, at least into the Cambrian.

Neoproterozoic extensional and contractional events are, for the most part, worldwide. Extension occurred within the time frame of about 870 to 740 Ma, whereas contractional events associated with subduction and magmatic arcs extend mainly from about 600 to 500 Ma.

Aulacogens, failed arms of three-armed rift systems, are common in the Neoproterozoic. They consist of graben-like structures commonly containing great thicknesses of sedimentary rock and extend into continents at a high angle to the margin. The other two arms of a rift system are generally not exposed but imply the existence of a continental margin. Aulacogens occur in the western, southern, and northeastern United States, in the subsurface of Baltica, and in South America, Africa, India, Australia, and Antarctica. The abundance of aulacogens in the Neoproterozoic may be related to widespread extension during the breakup of the Rodinian supercontinent.

As shown in Plate 2, major Neoproterozoic continents are defined most directly by the distribution of miogeoclinal sedimentary rocks, by the distribution of continent-fringing magmatic-arc rocks, by the distribution of oceanic mafic and ultramafic rocks adjacent to continents, and by aulacogens. Glaciogenic diamictite, common in the fringing miogeoclinal sediments, is a further guide to the recognition of continental margins.

As thus defined, the continents are used as a starting point in plate-tectonic reconstructions. The continents were assembled much like a jigsaw puzzle into a fairly tight-fitting assemblage. Possible conjugate continental pairs of margins are suggested by their similar histories. The reconstruction of southern Rodinia, similar to the Mesoproterozoic assembly described by many geologists (Fitzsimons, 2000b, 2003; Lawver and Scotese, 1987), allows a rather tight fit between Africa and South America. The reconstruction also places the west side of South America adjacent to the east side of Laurentia. Baltica is placed east of Laurentia, and east of Baltica are Arabia, India, and China. The reconstruction is similar to several previous reconstructions in showing South America east of Laurentia (Bond and others, 1984; Dalziel, 1992; Dalziel and others, 2000; Pisarevsky and others, 2003; Unrug, 1997). Siberia has been placed in several different positions including adjacent to the western United States (Karlstrom and others, 1999; Sears and Price, 2000, 2003; Sears and others, 2004); north of North America (Condie and Rosen, 1994); and finally, possibly east of Baltica, as described above under the discussion of Siberia.

The reconstruction proposed here differs from other reconstructions, (for example, Li and others, 2008), most notably in that it places Australia and Antarctica in the present-day southern latitudes rather than near western North America as shown by Burrett and Berry (2000), Karlstrom and others (1999), and Weil and others (1998). The placement shown here of these continents in Rodinia is based on the assembly pattern of Neoproterozoic and Grenville-age rocks and the presence of 3,000 Ma rocks in the region of Australia, Antarctica, and Africa (Rogers, 1996), which suggests that these continents were assembled at this time.

Paleomagnetic data in the time frame from 1,200 to 850 Ma, used to check the validity of the proposed model, show a rough compatibility with the proposed model, but the number of paleopoles is small and the paleopoles from any one continent are moderately scattered.

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Figures

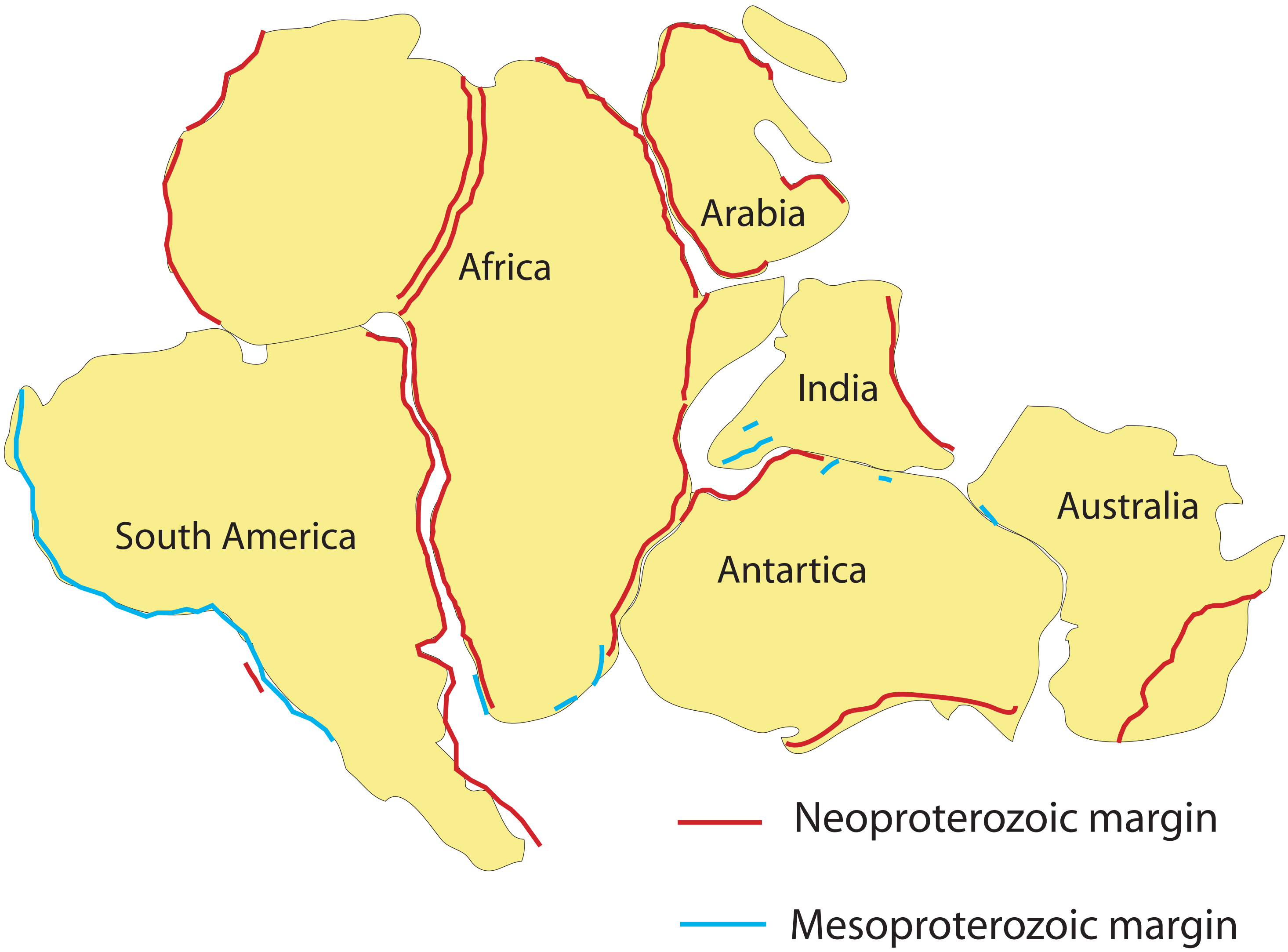


Figure 1. Tight fitting assembly of the southern continents (South America, Africa, India, Arabia, Antarctica, and Australia) based on Smith and Hallam, 1970 and Lawver and Scotesce, 1987. Assignments of ages of continental margins based on plate 2 of this report. The assembly has been considered by Smith and Hallam and Lawver and Scotesce to be Gondwana and thus Mesozoic in age in seeming contradiction to the proposed Meso-Neoproterozic age used here. As described in more detail under section entitled "Unchanged pattern of southern continents through time" the assembly patterns of the southern continents are generally similar in Rodinia, Gondwana, and Pangea, and thus a similarity in a Meso-Neoproterozoic assembly and a Mesozoic assembly is possible.

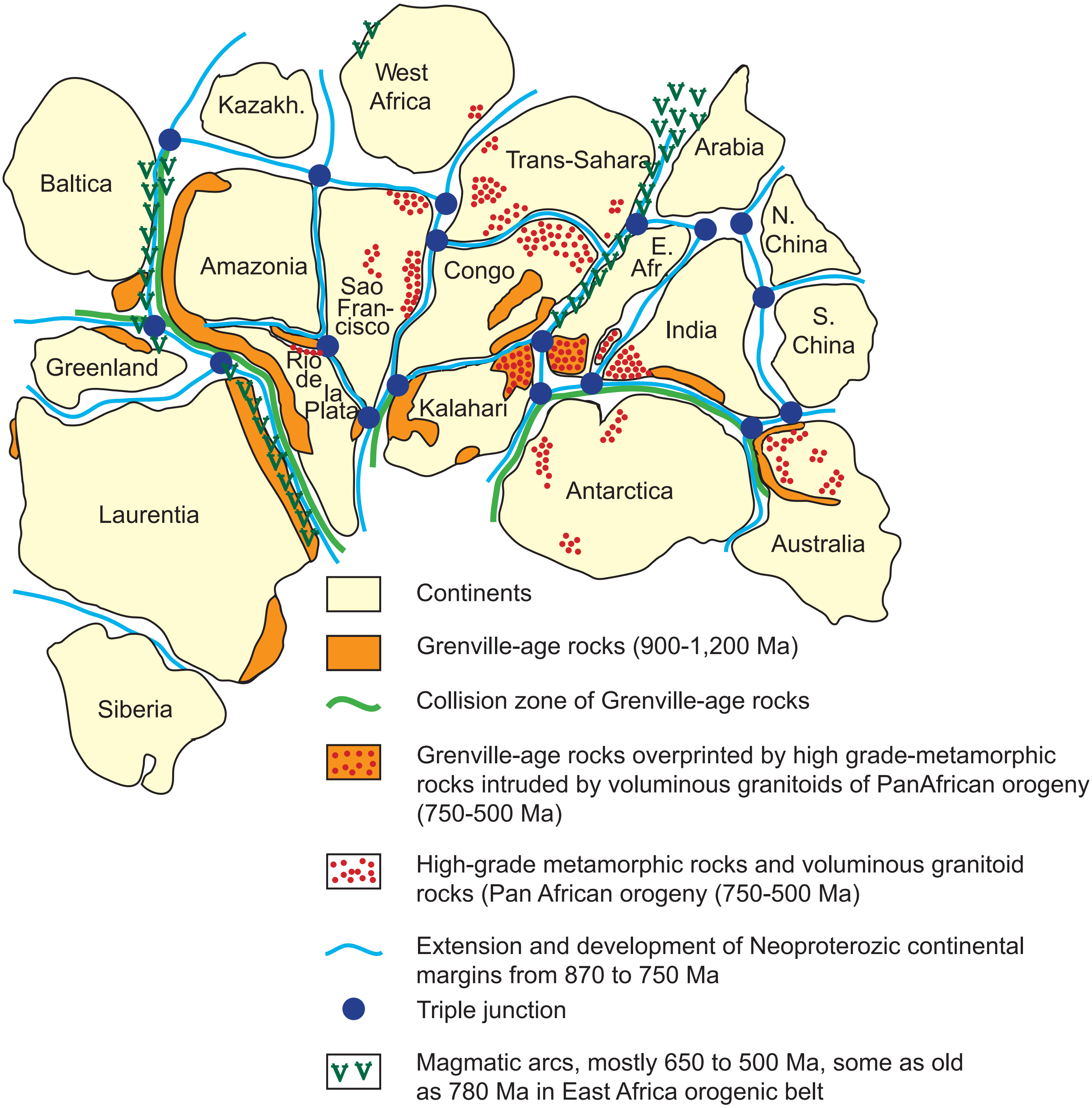


Figure 2. Proposed composite model of the assembly and breakup of Rodinia (1200-850 Ma). Yellow indicates continental blocks; orange indicates Grenville-age (1200 to 900 Ma) high-grade metamorphic rocks formed during the assembly of Rodinia; blue lines indicate the fragmentation pattern of Rodinia; dark - blue dots indicate proposed sites of triple junctions; green "V's" indicate the location of magmatic-arc rocks, mostly 600 to 500 Ma, except in the East African Orogenic Belt, where magmatic-arc rocks are as old as 750 Ma; red dots indicate high-grade metamorphic rocks formed during the Late Neoproterozoic Pan-African orogeny.

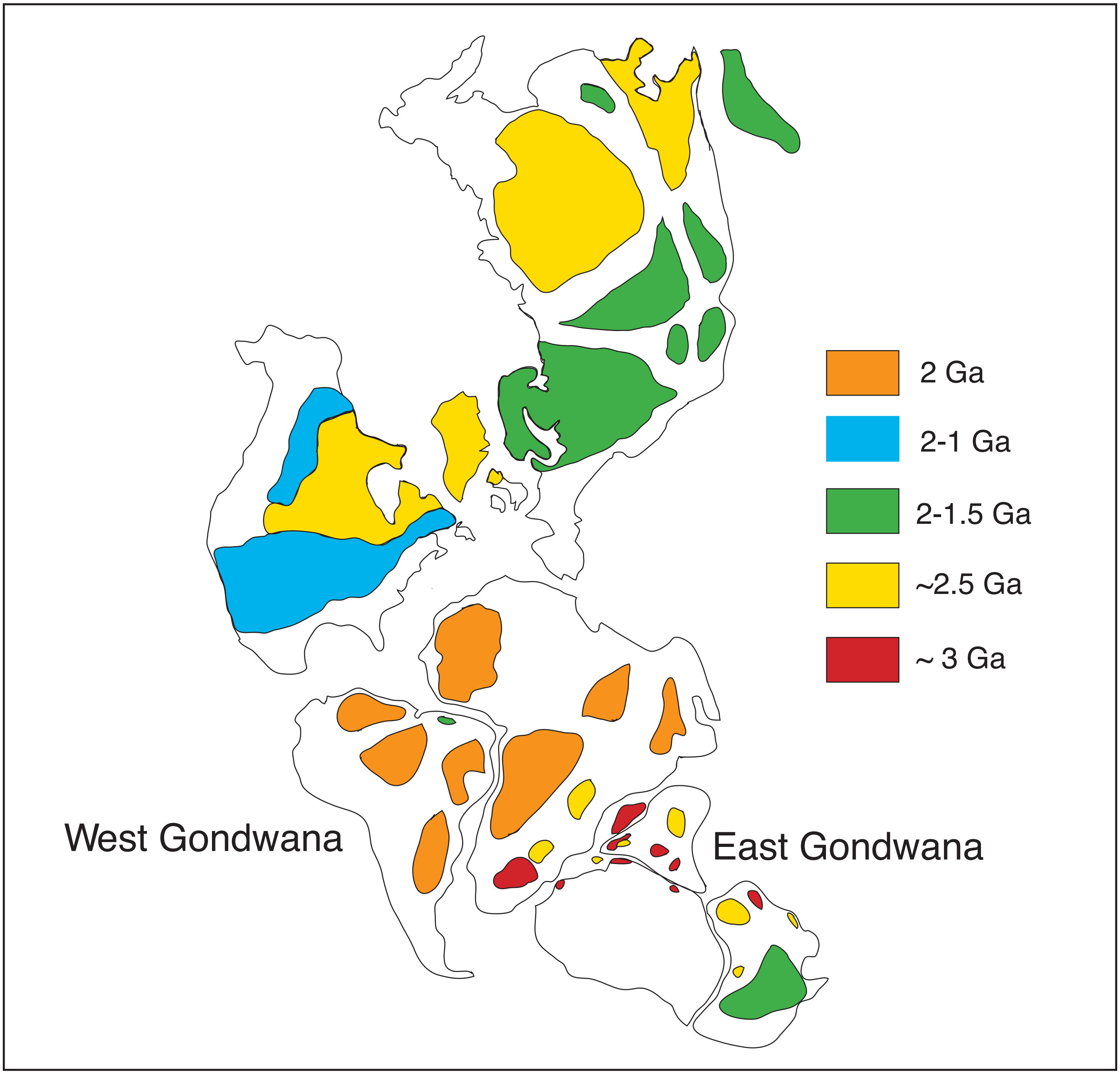
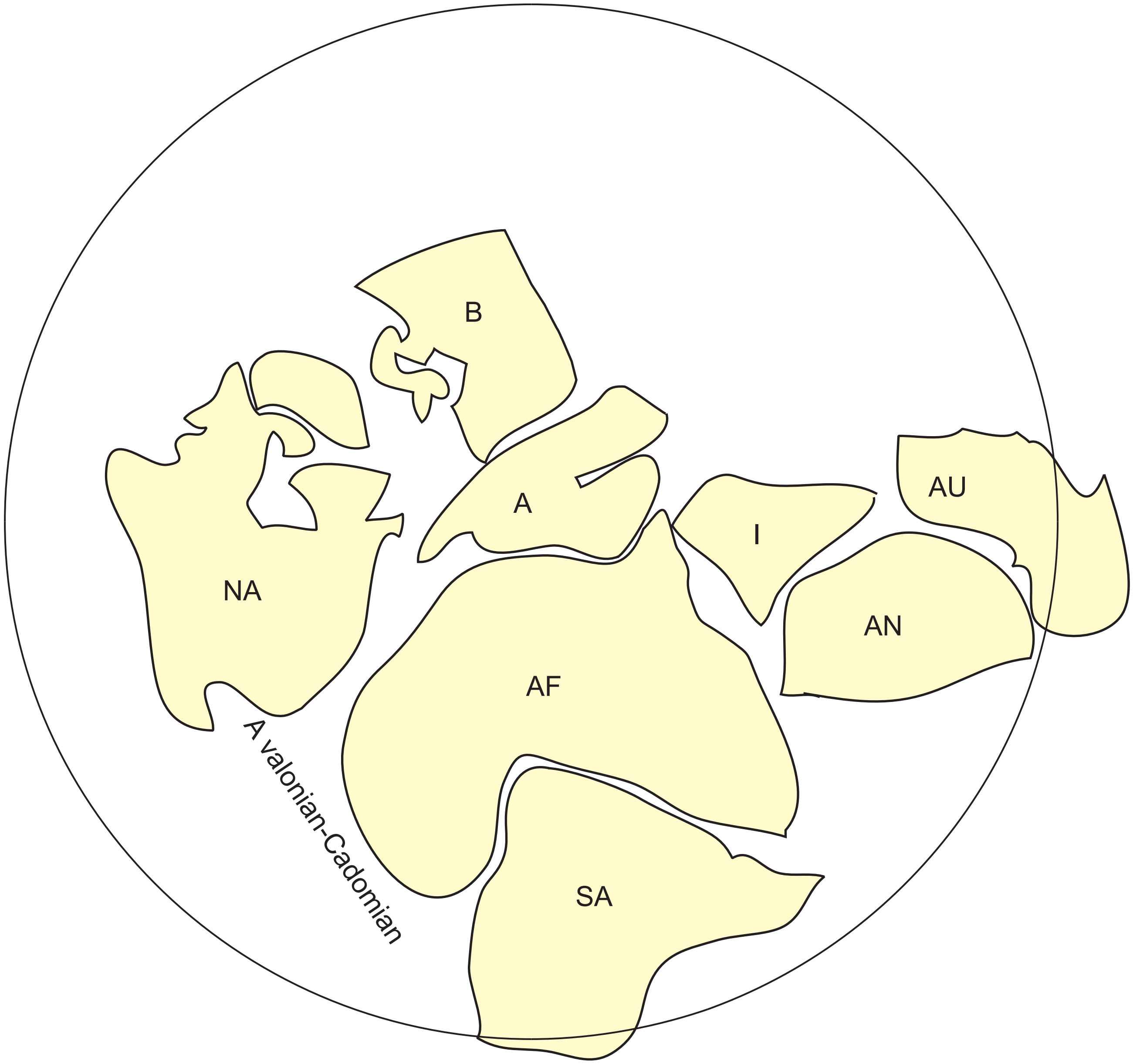


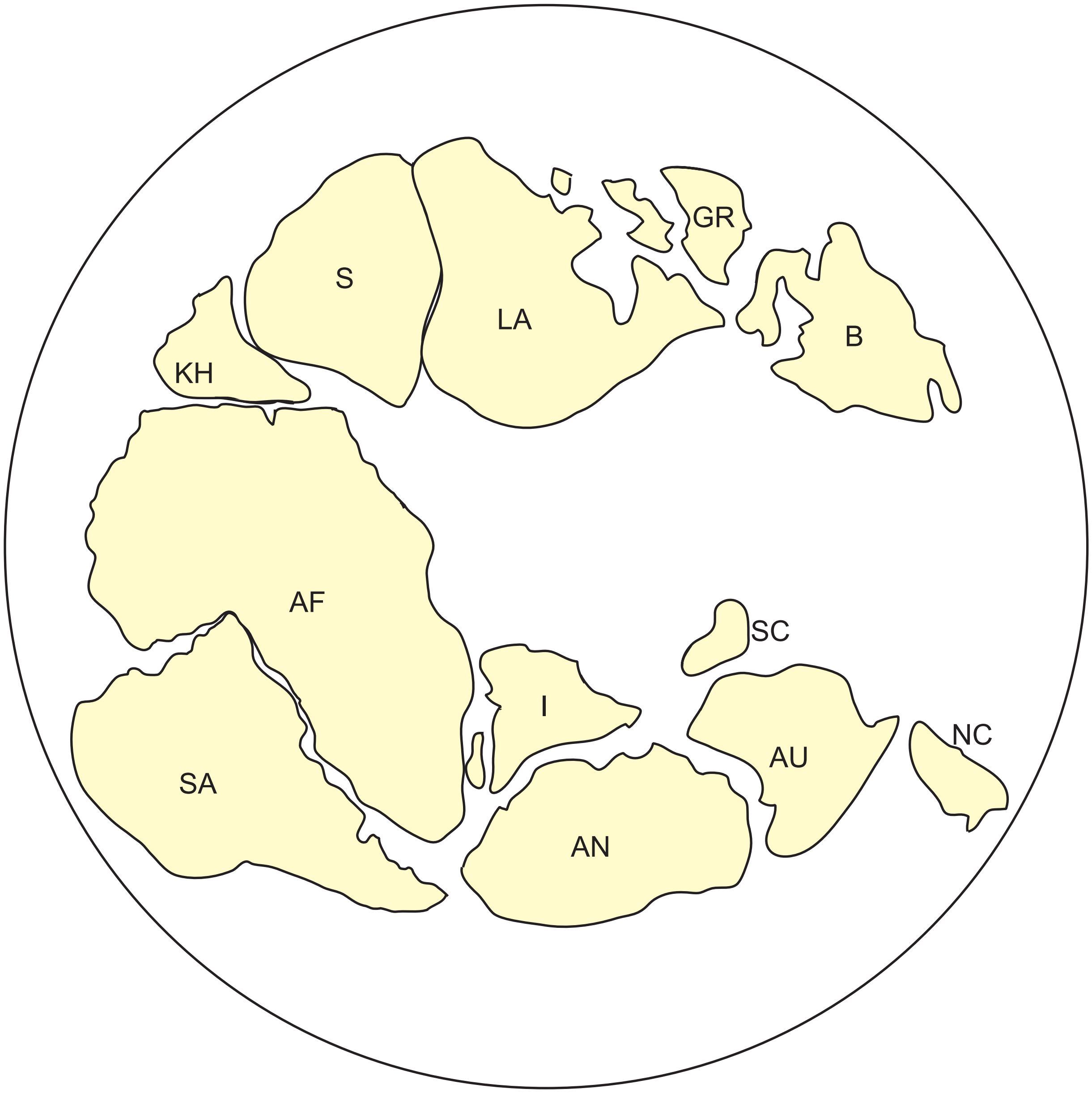
Figure 3. Ages of Precambrian cratons on a Pangea base (Rogers, 1996).

Figure 4. Rodinia reconstruction (ca. 600 Ma) of Bond and others (1984). Drafted using figure 3 of Nance and Murphy (1994).



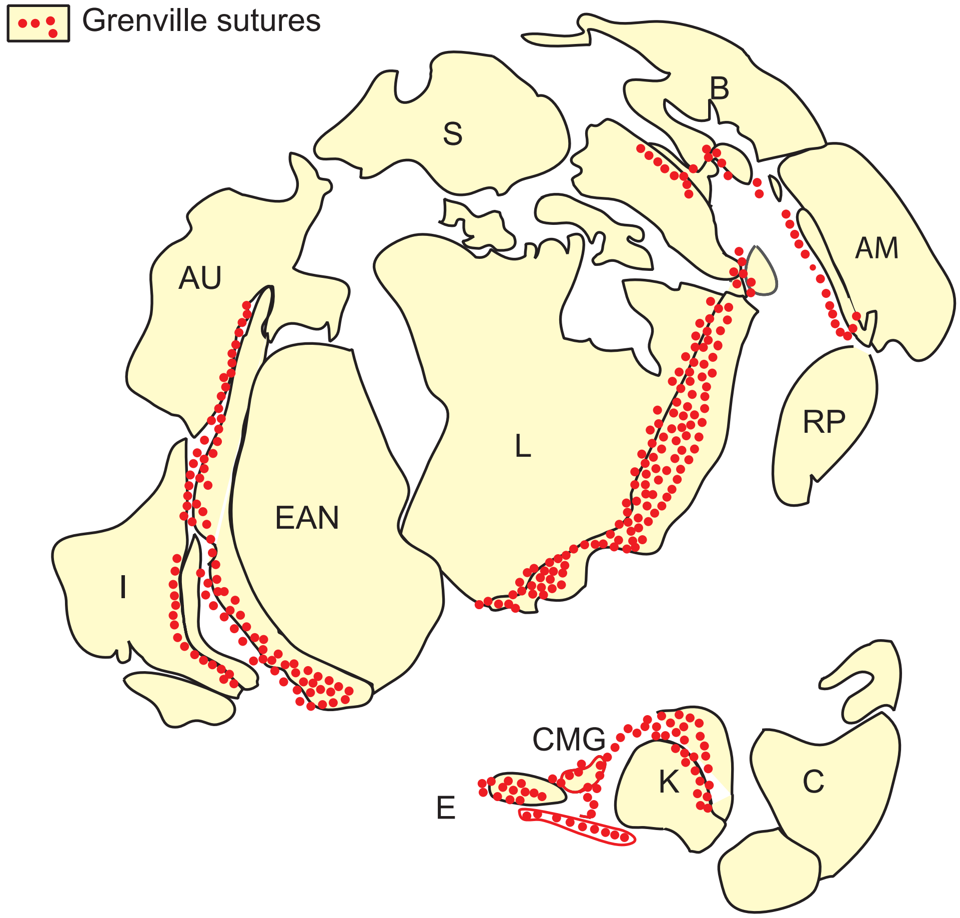
A,	Arabia	B,	Baltica
AF,	Africa	I,	India
AN,	Antarctica	NA,	North America
AU,	Australia	SA,	South America

Figure 5. Rodinia reconstruction (ca. 600 Ma) of Keppie (1992). Drafted using figure 4 of Nance and Murphy (1994).



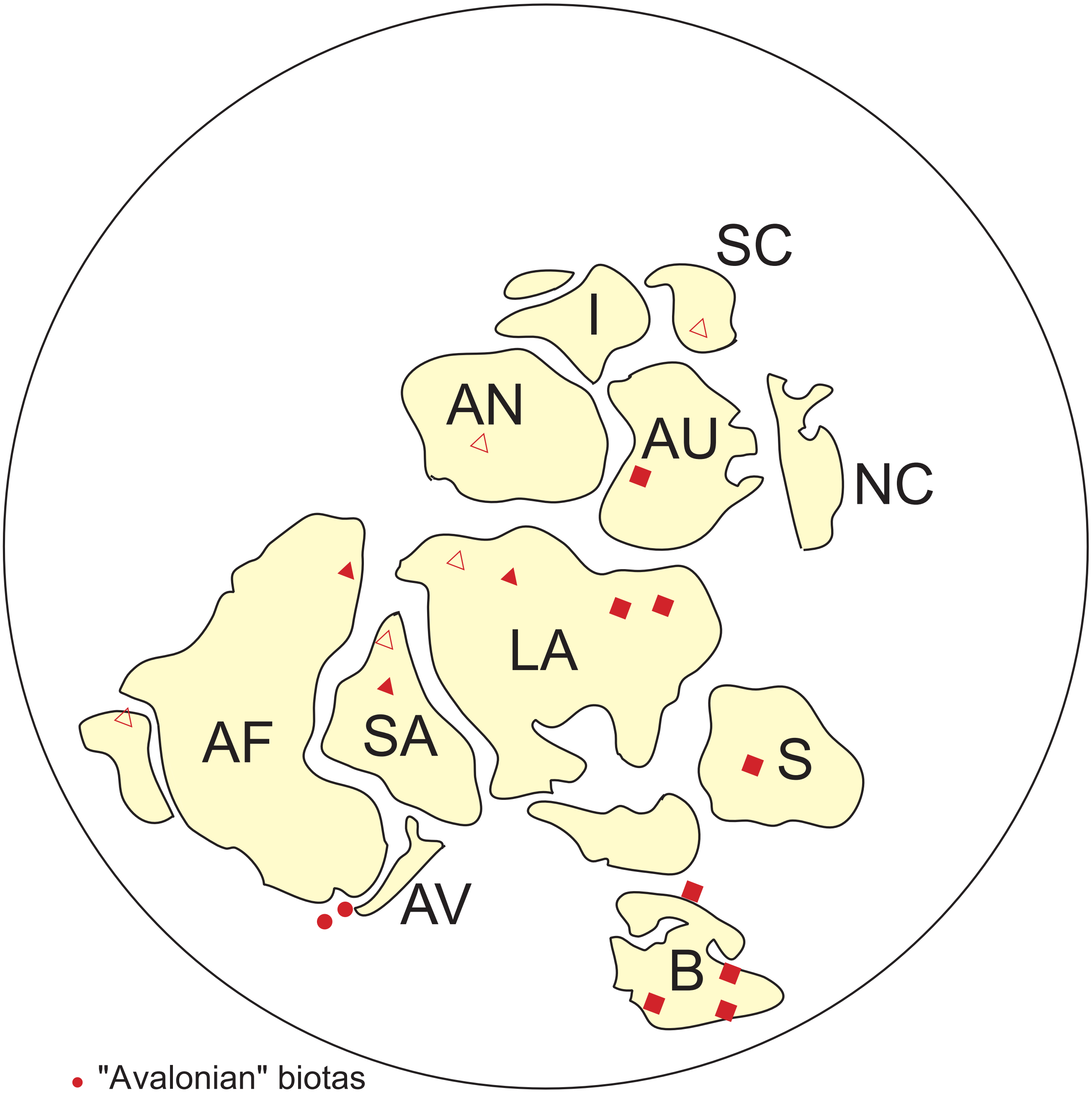
AF,	Africa	KH,	Kazakhstan
AN,	Antarctica	LA,	Laurentia
AU,	Australia	NC,	North China
B,	Baltica	S,	Siberia
GR,	Greenland	SA,	South America
I,	India	SC,	South China

Figure 6. Model of Paleopangaea (800-600 Ma) proposed by Piper (2000).



AM,	Amazonia	EAN,	East Antarctica
AU,	Australia	I,	India
B,	Baltica	K,	Kalahari
C,	Congo	L,	Laurentia
CMG,	Coates Land-Maudheim-Grunehogan	RP,	Rio de la Plata
E,	Ellsworth-Whitmore Mountains	S,	Siberia

Figure 7. Rodinia reconstruction of Dalziel and others (2000).



- ● "Avalonian" biotas
- ▲ Namibian-type biotas
- ■ Biotas of the Australia-Baltica-north Laurentia-Siberia cluster
- △ Cloudiniids but poor or no soft-bodied biotas

AF,	Africa	LA,	Laurentia
AN,	Antarctica	NC,	North China
AU,	Australia	S,	Siberia
AV,	Avalonia	SA,	South America
B,	Baltica	SC,	South China
I,	India		

Figure 8. Rodinia reconstruction of Waggoner (1999).

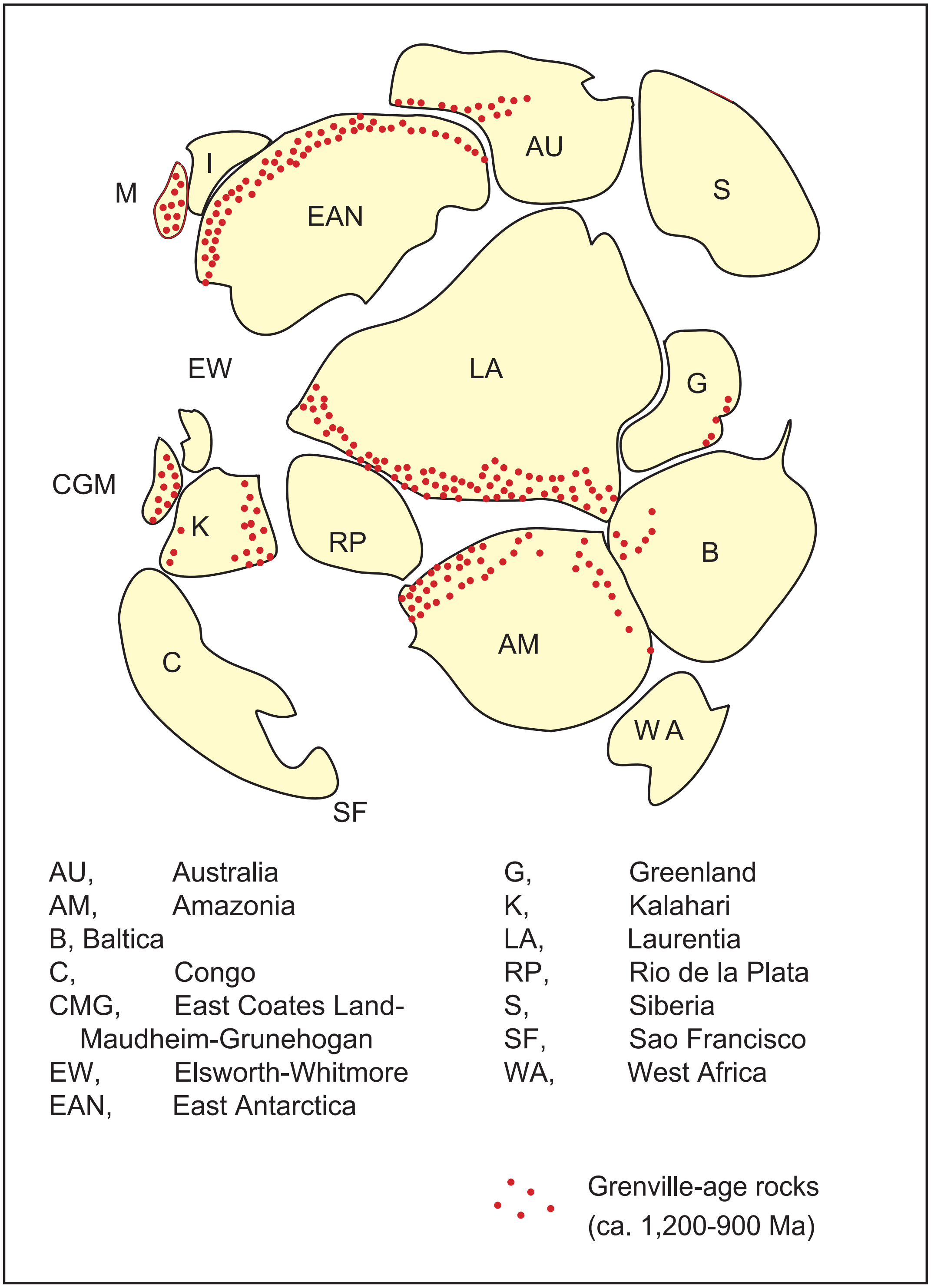


Figure 9. Rodinia reconstruction of Weil and others (1998).

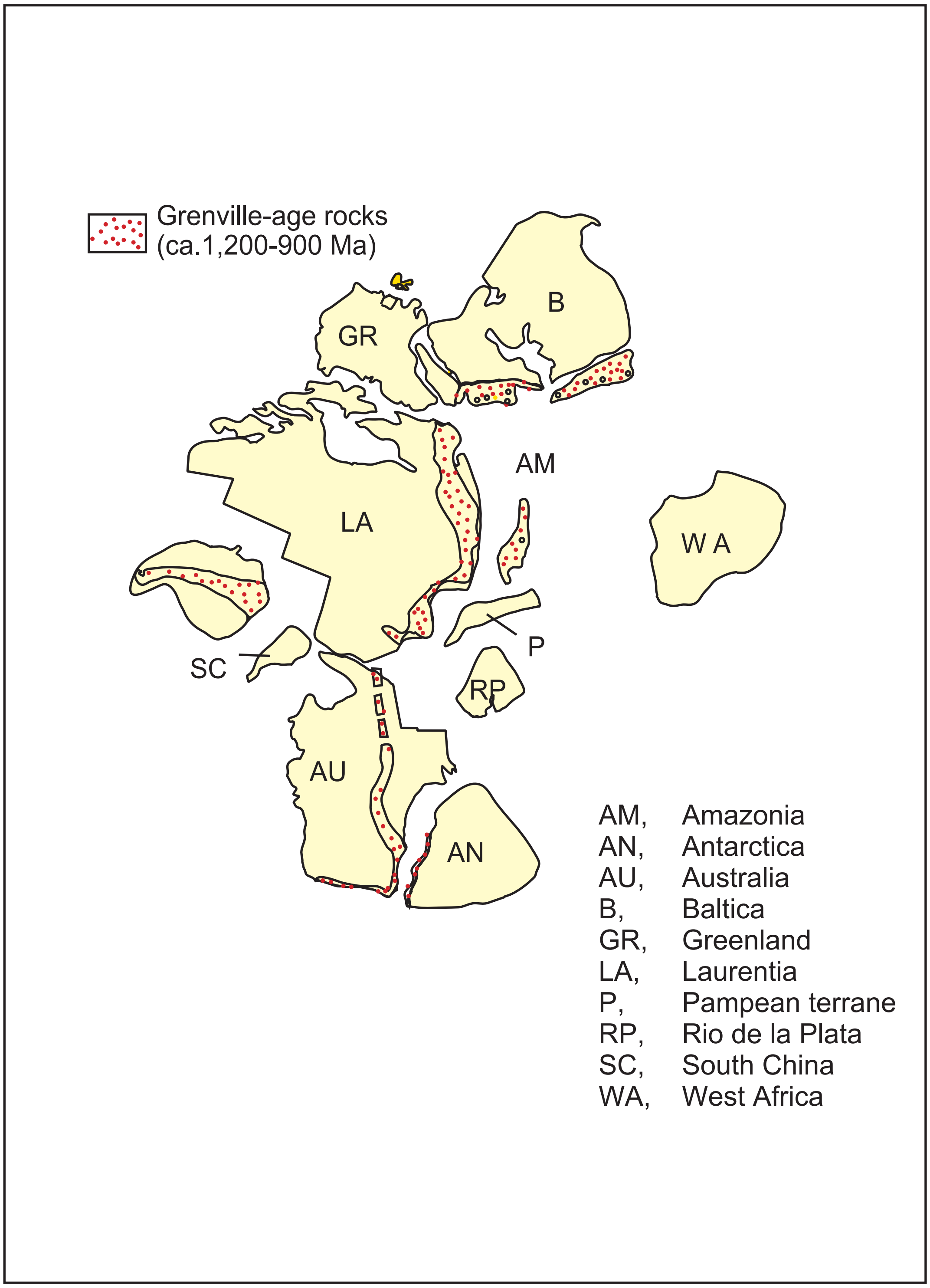


Figure 10. Rodinia reconstruction (ca. 990 Ma) of Pisarevsky and others (2003).

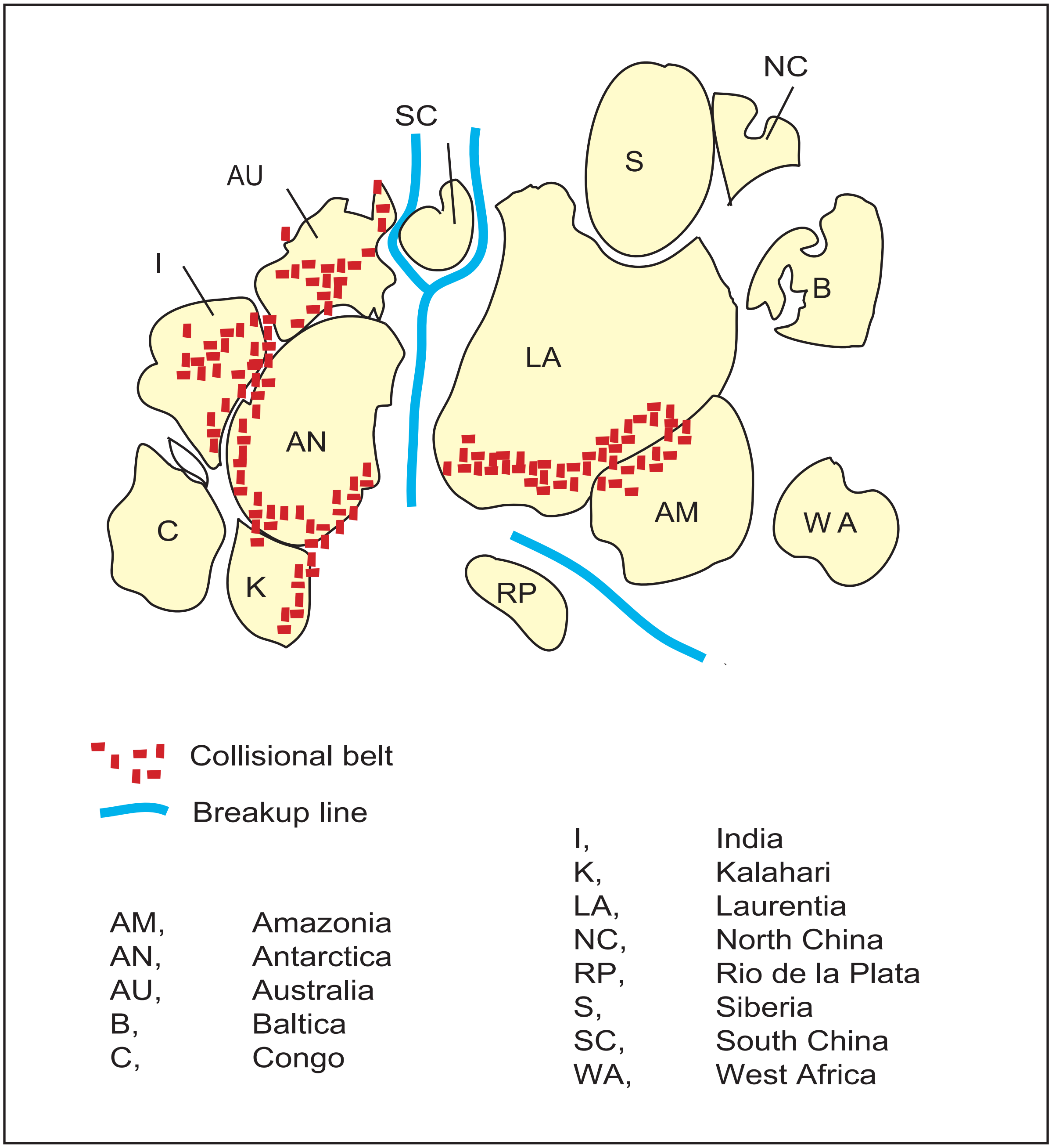


Figure 11. Rodinia reconstruction (ca. 1,000-700 Ma) of Unrug (1997, fig. 2).

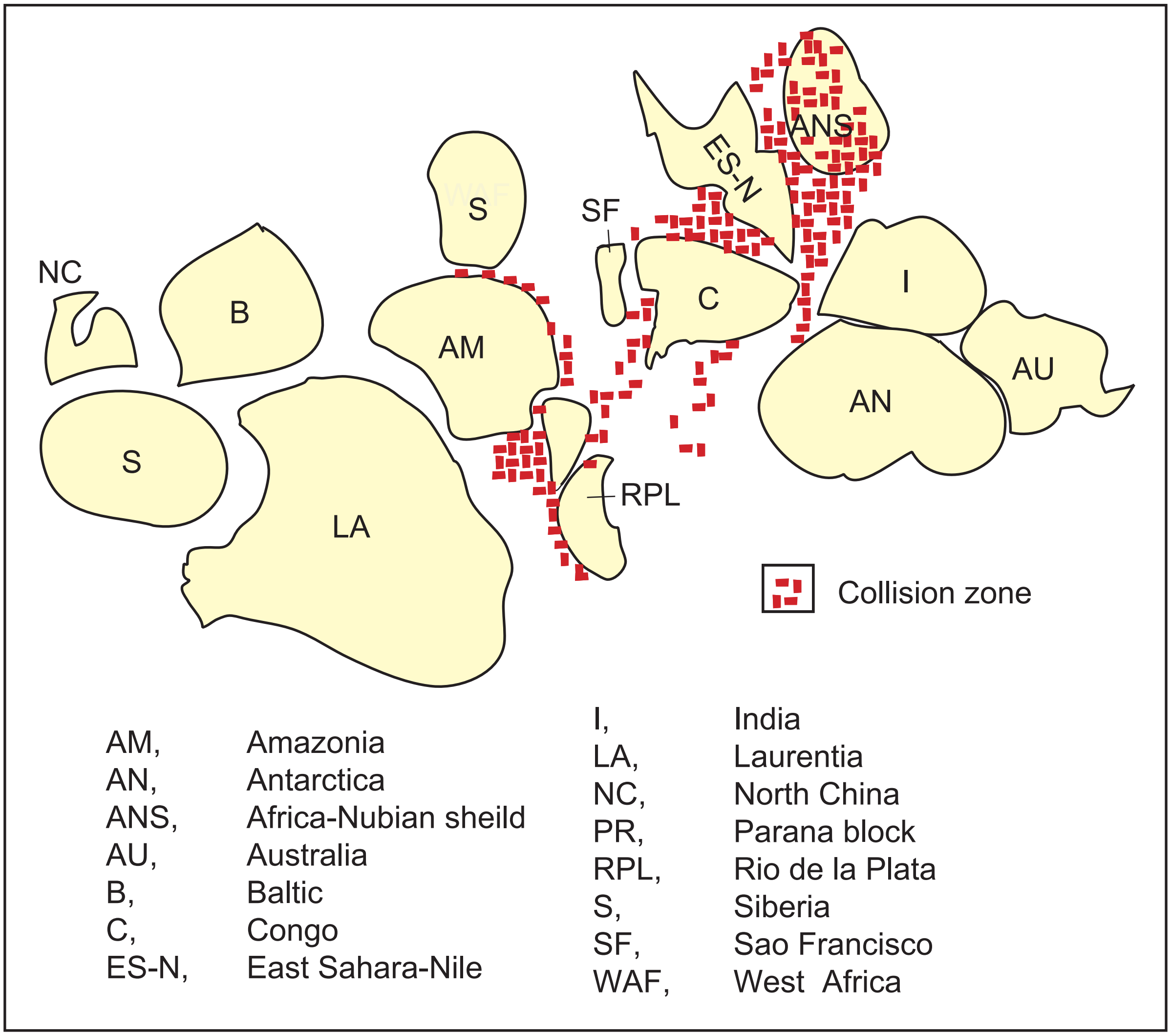


Figure 12. Rodinia reconstruction (ca. 700-500 Ma) of Unrug (1997, figure 3)

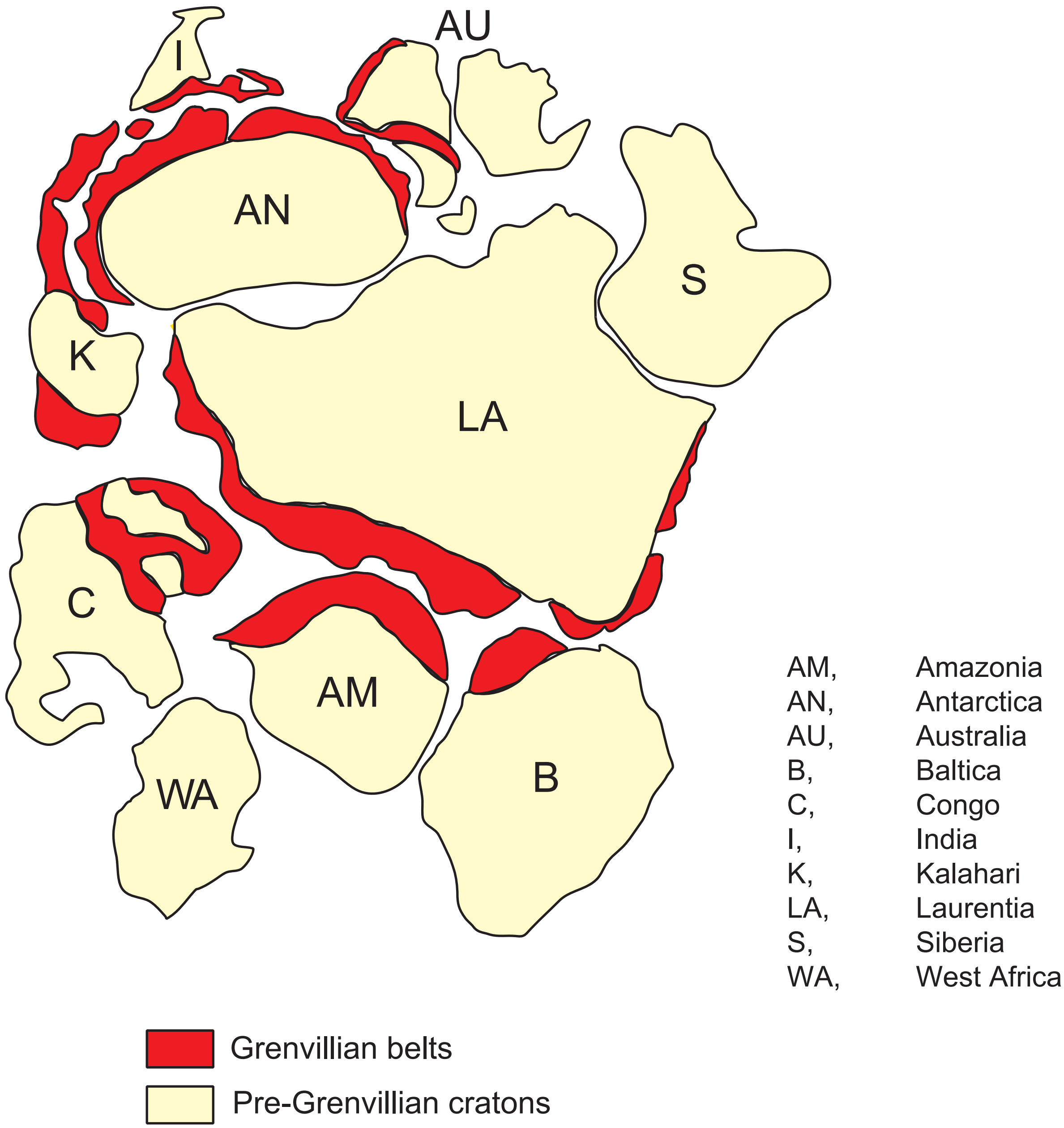
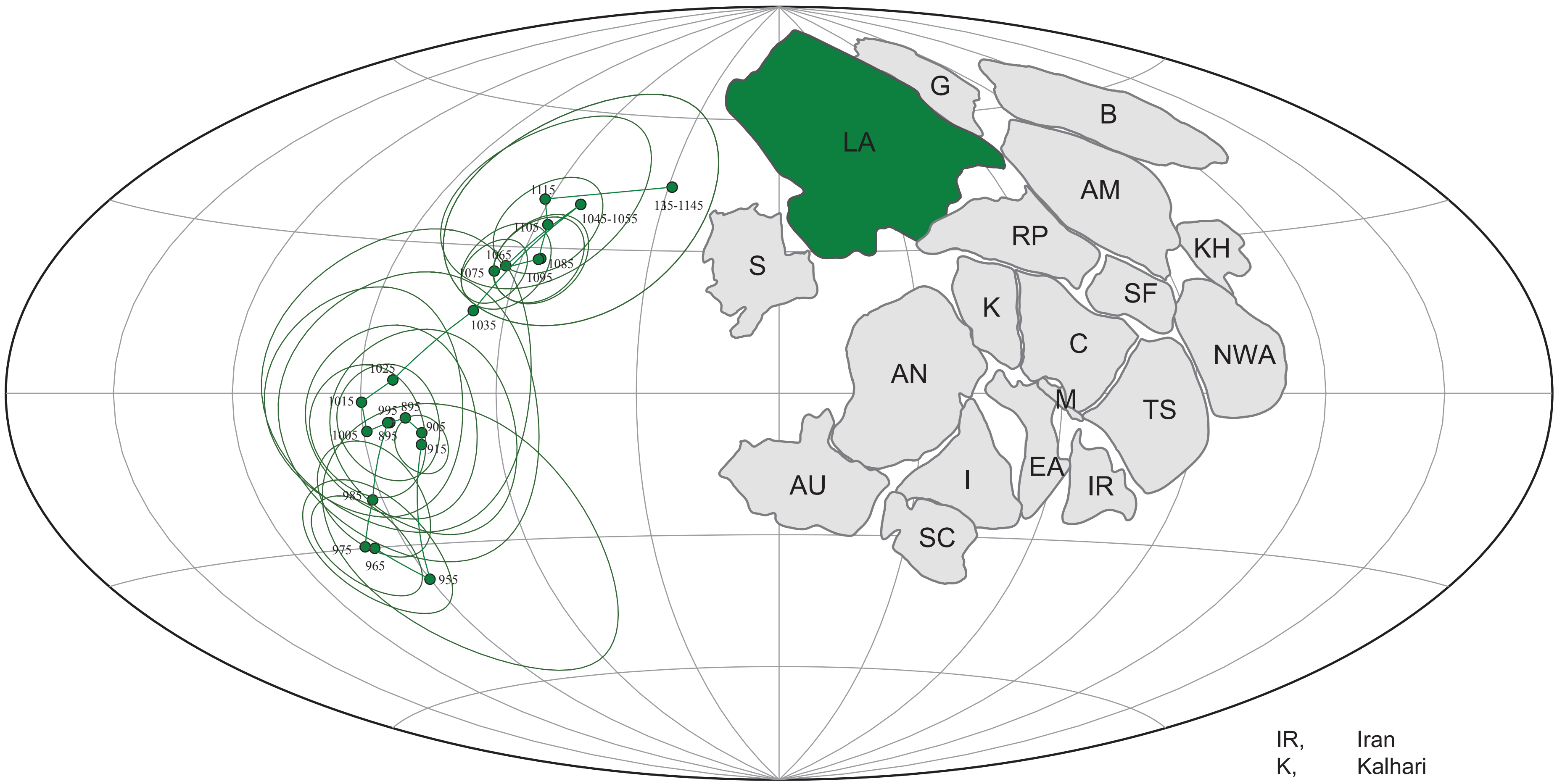


Figure 13. Rodinia reconstruction (ca. 700 Ma) of Hoffman (1991).



1,200-850 Ma

AM, Amazonia
AN, Antarctica
AU, Australia
B, Baltic
C, Congo
EA, East Africa
G, Greenland
I, India
IR, Iran
K, Kalhari
KH, Kazakhstan
LA, Laurentia
M, Madagascar
NWA, Northwest Africa
RP, Rio de la Plata
S, Siberia
SC, South China
SF, Sao Francisco

Figure 14. Paleomagnetic poles for Laurentia (green) for the time period 1,200-850 Ma, and an averaged apparent polar wander path (APWP) derived from these poles.

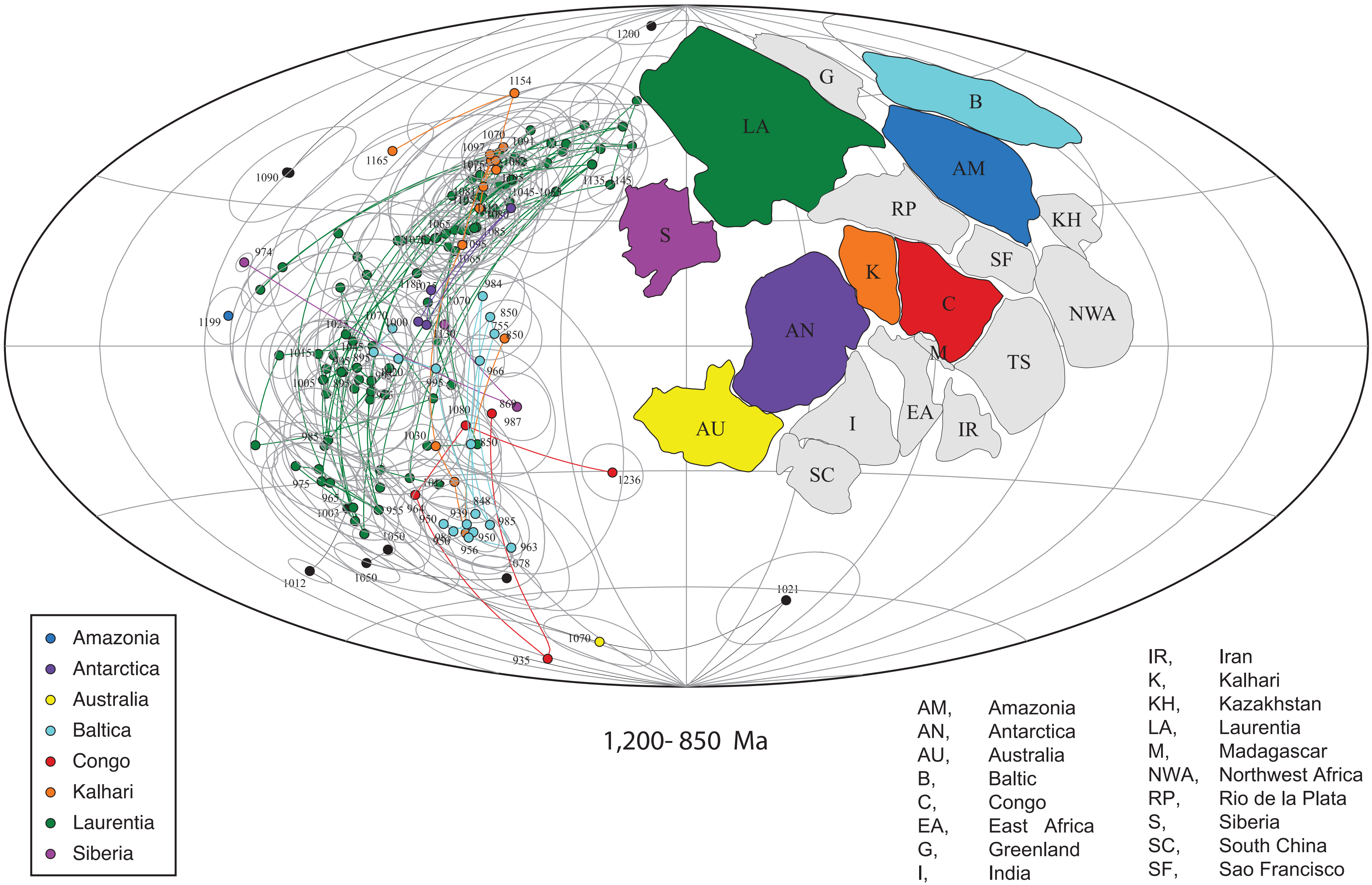


Figure 15. Paleomagnetic poles for Amazonia, Antarctica, Australia, Baltica, Congo, and Siberia for the time period 1,200-850 Ma. Also shown is the averaged APWP for Laurentia. Gray continents are those for which no paleomagnetic data were used. Poles have been rotated into a present-day Laurentian reference frame Paleomagnetic poles are from Table 1.

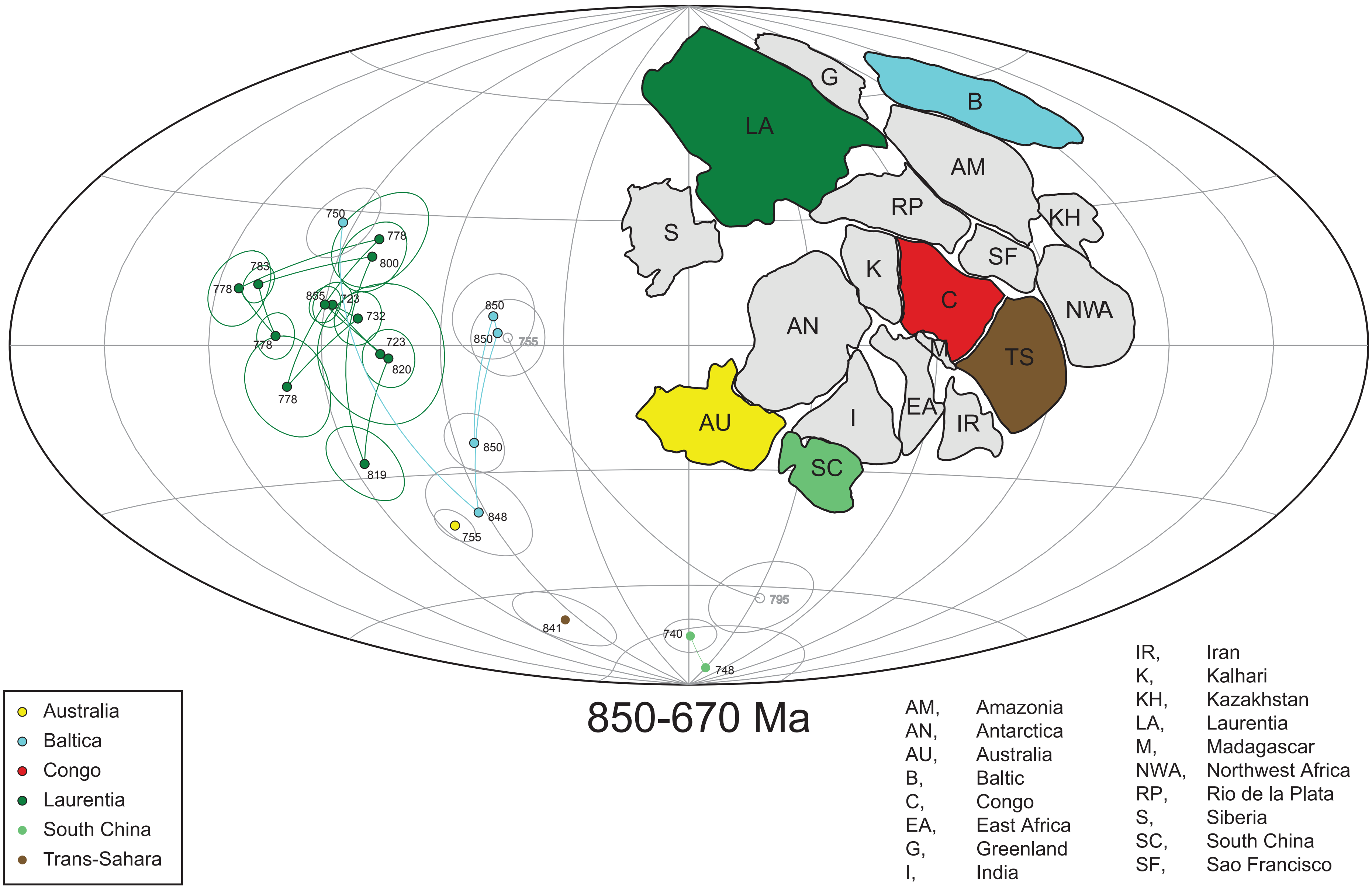


Figure 16. Paleomagnetic poles for Australia, Baltica, Congo, Laurentia Kalahari, South China, and Trans-Sahara for the time period 850-670 Ma.

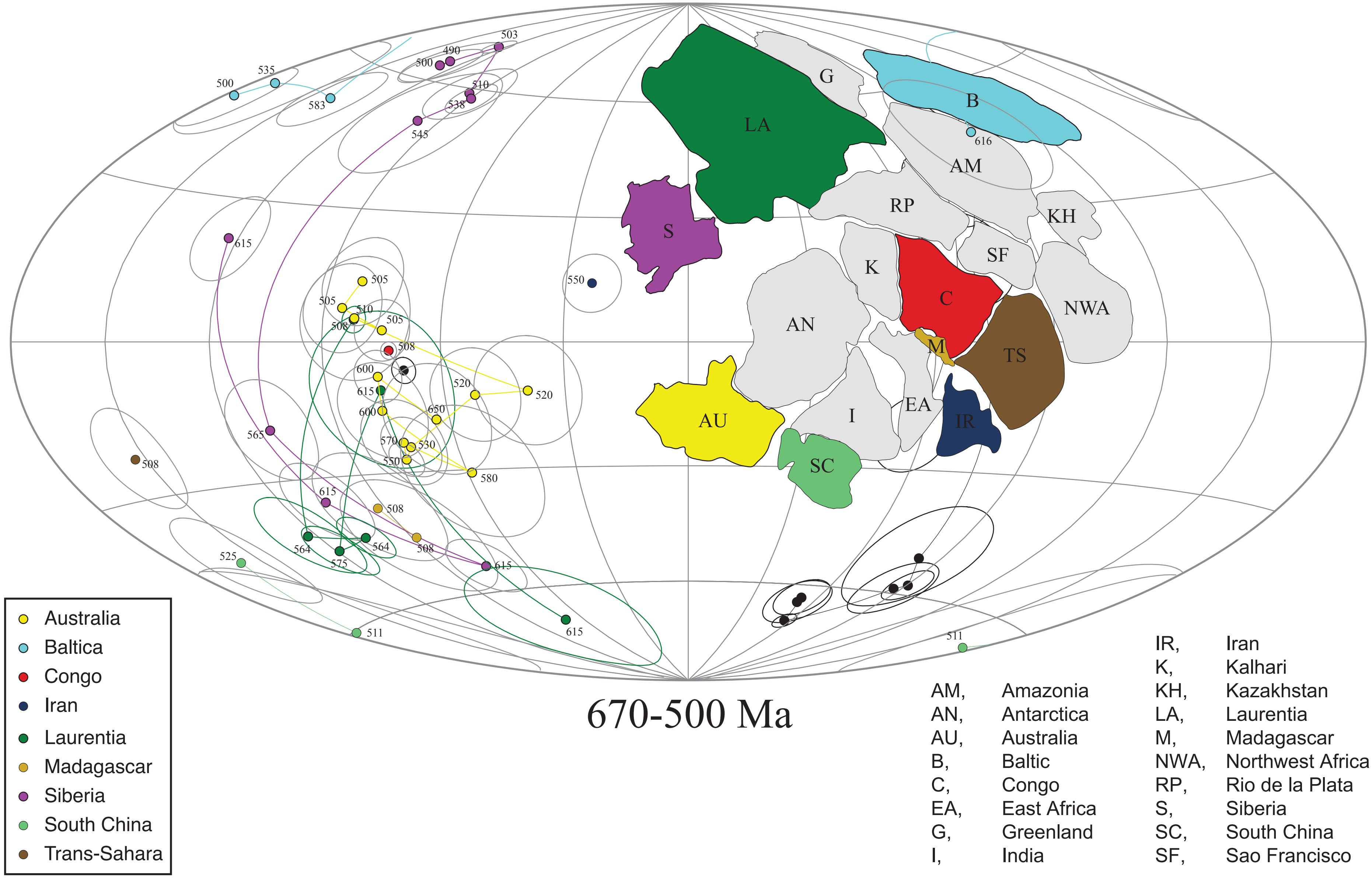


Figure 17. Paleomagnetic poles for Australia, Baltica, Congo, Iran, Laurentia, Madagascar, Siberia, South China and Trans-Sahara for the time period 670-500 Ma. Poles have been rotated into a present-day Laurentian reference frame. Paleomagnetic poles are from Table 1.

Tables

Table 1. Neoproterozoic paleomagnetic poles used in Figures 14–17. Poles are compiled from pole lists given by Meert and Torsvik (2003), Weil and others (1998), Pisarevsky and others (2003), and D'Agrella-Filho and others (1998). Table lists (from left to right): continent, site latitude (slat), site longitude (slon), pole name (name), mean age of pole in Ma based on range given in reference age), pole latitude (plat), pole longitude (plon), alpha 95 or dp/dm for pole given in reference (A95), reference, pole-list source (DAF= D'Agrella-Filho and others, 1998; MT= Meert and Torsvik, 2003; P=Pisarevsky and others, 2003; Weil and others, 1998). Note that Saõ Francisco and India data were not used for reasons discussed in text.

Continent	slat	slon	Name	Age	plat	plon	A95	Reference	Source[1]
Amazonia	-10.8	-63.7	Nova Floresta	1,199.0	25.0	165.0	6.0	Tohver et al., 2002	MT
Antarctica	-72.7	-2.5	Borgmassivet	1,000.0	-6.0	233.0	7.0	Hodgkinson, 1989; Moyes et al., 1995	DAF
Antarctica	-77.9	-30.3	Coats Land Nunataks	1,112.0	22.9	80.3	6.8	Gose et al., 1997	DAF,MT
Antarctica	-73.0	-10.0	Ritscherflya	1,130.0	-8.0	232.0	4\4	Jones et al., 1999	MT,P
Antarctica	-71.6	-2.0	Ahlmannryggen	1,183.0	-9.0	240.0	6.0	Peters, 1989	DAF
Australia	-31.1	139.0	Giles Creek Dolomite-lower	505.0	38.0	205.0	10.0	Klootwijk, 1980	MT
Australia	-31.1	139.0	Lake Frome - A	505.0	31.0	207.0	10.0	Klootwijk, 1980	MT
Australia	-31.1	139.0	Pertaoorta Group	505.0	33.0	192.0	7.0	Klootwijk, 1980	MT
Australia	-31.7	139.0	Billy Creek Aroona-Wirrealpa - A	510.0	37.0	200.0	14.0	Klootwijk, 1980	MT
Australia	-15.0	130.1	Antrim plateau volcanics	520.0	9.0	160.0	13.0	McElhinny and Luck, 1970	MT
Australia	-31.0	138.9	Hawker Group A	520.0	21.0	165.0	11.0	Klootwijk, 1980	MT
Australia	-23.6	134.5	Todd River	530.0	43.0	160.0	7.0	Kirschvink, 1978	MT
Australia	-23.6	134.5	Upper Arumbera SS	550.0	46.0	157.0	4.0	Kirschvink, 1978	MT
Australia	-23.6	134.5	Lower Arumbera/Pertataka Fm	570.0	44.0	162.0	10.0	Kirschvink, 1978	MT
Australia	-31.4	139.2	Brachina Fm	580.0	33.0	148.0	16.0	McWilliams and McElhinny, 1980	MT
Australia	-31.3	138.6	Elatina	600.0	40.0	182.0	6.0	Sohl et al., 1999	MT
Australia	-31.3	138.6	Yalipena Fm	600.0	44.0	173.0	11.0	Sohl et al., 1999	MT
Australia	-30.5	139.3	Angepena Fm	650.0	33.0	164.0	13.0	McWilliams and McElhinny, 1980	MT
Australia	-23.8	115.7	Mundine dikes	755.0	45.0	135.0	4.0	Wingate and Giddings, 2000	MT,P
Australia	-23.8	116.6	Bangemall Basin Sills	1,070.0	34.0	95.0	8.0	Wingate et al., 2002	MT,P
Baltica	56.0	10.9	Andarum limestone	500.0	52.0	111.0	7\10	Torsvik and Rehnstrom, 2001	MT
Baltica	68.9	19.5	Tornetrask Fm	535.0	56.0	116.0	12\15	Torsvik and Rehnstrom, 2001	MT
Baltica	59.1	9.1	Fen complex	583.0	56.0	150.0	7\10	Meert et al., 1998	MT
Baltica	58.4	6.2	Egersund dikes	616.0	48.0	20.0	14.0	Poorter, 1972; Torsvik et al., unpublished data (see Meert and Torsvik, 2003)	MT,P
Baltica	70.5	30.0	Mean pole	750.0	-28.0	17.0	8.0	Meert and Torsvik, 2003	MT
Baltica	55.5	6.7	Hunnedalen dikes	848.0	41.0	42.0	11\12	Walderrhaug et al., 1999	MT,P
Baltica	59	16.0	East of Protogine Zone	850.0	0.0	242.0	-	Pesonen et al., 1989	W
Baltica	59	13.0	West of Protogine Zone	850.0	4.0	241.0	10.0	Pesonen et al., 1989	W
Baltica	59	13.0	West of Protogine Zone	850.0	-25.0	231.0	7.0	Pesonen et al., 1989	W
Baltica	56.6	-17.0	Pyätteryd Amphibolite	939.0	-43.0	217.0	11.0	Pisarevsky and Bylund, 1998; Wang et al., 1996; Wang and Lindh, 1996	P
Baltica	59	16.0	East of Protogine Zone	950.0	-42.0	210.0	-	Pesonen et al., 1989	W
Baltica	59	13.0	West of Protogine Zone	950.0	-45.0	217.0	5.0	Pesonen et al., 1989	W
Baltica	59	15.0	Within Protogine Zone	950.0	-44.0	211.0	11.0	Pesonen et al., 1989	W
Baltica	57.1	-17.1	Gällared Amphibolite	956.0	-46.0	214.0	19.0	Pisarevsky and Bylund, 1998; Möller and Södderland, 1997	P
Baltica	56.5	-17.2	Känna Gneiss	963.0	-50.0	225.0	17.0	Pisarevsky and Bylund, 1998; Wang et al., 1996; Wang and Lindh, 1996	P
Baltica	61.0	16.0	Falun dolerite	966.0	6.0	58.0	6.0	Patchett and Bylund, 1977	MT
Baltica	57.7	15.0	Nilstorp dolerite	984.0	-9.0	59.0	10.0	Patchett and Bylund, 1977	MT
Baltica	57.1	-18.1	Gällared Granite Gneiss	985.0	-44.0	224.0	6.0	Pisarevsky and Bylund, 1998; Möller and Södderland, 1997	P

Continent	slat	slon	Name	Age	plat	plon	A95	Reference	Source[1]
Baltica	59.3	17.0	Arby dolerite	995.0	7.0	47.0	7.0	Patchett and Bylund, 1977	MT
Baltica	69.4	28.6	Laanila dyke swarm, Finland	1,020.0	-4.0	218.0	6.0	Pesonen et al., 1989	W
Baltica	69.4	28.6	Laanila Dolerite	1,045.0	-2.0	212.0	15.0	Mertanen et al., 1996	P
Baltica	58.8	12.1	Bamble intrusions (mean)	1,070.0	-3.0	37.0	15.0	Meert and Torsvik, 2003	MT
Congo	-15.3	35.2	Ntonya Ring structure (PA)	522.0	28.0	355.0	2.0	Briden et al., 1993	MT
Congo	-4.6	30.1	Gagwe lavas	795.0	-25.0	273.0	10.0	Meert et al., 1995	MT,P,W
Congo	-23.3	31.4	Bukoban intrusives, Tanzania	806.0	11.0	101.0	19.0	Piper, 1972	W
Congo	-18.0	15.0	Nosib group	869.0	28.0	323.0	15.0	Van der Voo and Meert, 1991	DAF
Congo	-3.5	30.0	Nyabikere massif	935.0	43.0	137.0	14.0	Meert et al., 1994a,b	MT,W
Congo	-40.6	34.7	Kisi lavas	964.0	-0.1	158.0	11.0	Onstott et al., 1986	DAF
Congo	-15.0	30.0	Chaela group	1,080.0	23.0	329.0	36.7	Renne et al., 1990; Jones et al., 1992	DAF
Congo	-4.0	30.0	Host-Kibaran intrusives	1,236.0	-17.0	113.0	7.0	Meert et al., 1994a,b	MT
East Africa	0.5	37.1	Sinyai dolerite (PA)	547.0	-28.0	319.0	5.0	Meert and Van der Voo, 1996	MT
India	25.5	78.0	Bhander-Rewa (IND)	750.0	-47.0	33.0	6\6	McElhinny et al., 1978	MT
India	-4.4	55.4	Mahe Dikes (SEY)	750.0	80.0	79.0	9.9\14.9	Torsvik et al., 2001b; Hargraves and Duncan, 1990	MT,P
India	-4.4	55.4	Mahe granites (SEY)	755.0	77.0	23.0	1.7\2.6	Torsvik et al., 2001b; Suwa et al., 1994	MT
India	26.0	72.7	Malani rhylites (IND)	761.0	75.0	71.0	10.0	Torsvik et al., 2001a	MT,P
India	12.6	77.5	Harohalli dikes (IND)	821.0	27.0	79.0	9.0	Radhakrishna and Hoseph, 1996	MT,P
India	15.1	77.3	Lattavaram Kimberlite (IND)	1,090.0	-45.0	238.0	11.0	Miller and Hargraves, 1994	MT
India	24.0	79.5	Majhgawan Kimberlite (IND)	1,116.0	39.0	217.0	31.0	Miller and Hargraves, 1994	MT
India	24.6	83.1	Kaimure series (IND)	1,200.0	82.0	286.0	6.0	Sahasrabudhe and Mishra, 1966	MT
Iran	17.2	54.5	Mirbat sandstone (ANS)	550.0	-32.0	134.0	7.0	Kempf et al., 2000	MT
Kalahari	-9.1	33.9	Mbozi complex (CC)	755.0	46.0	325.0	9.0	Meert et al., 1995	MT,P
Kalahari			Port Edward Charnockite	985.0	5.0	148.0	9.0	Onstott et al., 1986	W
Kalahari	-29.0	20.0	Central Namaqua metamorphics	1,015.0	8.0	330.0	10\10	Onstott et al., 1986	DAF,MT,P,W
Kalahari	17.9	-29.4	O'Okiep intrusives	1,030.0	-15.0	155.0	15.0	Piper, 1975	W
Kalahari	-28.5	21.7	Kalkpunt fm	1,065.0	57.0	3.0	7.0	Onstott et al., 1986; Briden et al., 1979	MT,P,DAF,W
Kalahari	-10.6	11.6	Umkondo lavas	1,080.0	-63.0	196.0	15.0	McElhinny, 1966	W
Kalahari	-10.6	11.6	Umkondo combined	1,081.0	-64.0	208.0	8.0	McElhinny, 1966	W
Kalahari	-10.6	11.6	Umkondo dolerites	1,082.0	-65.0	223.0	6.0	McElhinny and Opdyke, 1964	DAF,W
Kalahari	-20.5	27.5	Post-Waterberg diabase	1,091.0	65.0	51.0	8.0	Jones and McElhinny, 1966	DAF,MT,W
Kalahari	-24.0	31.5	Timbavati gabbro	1,097.0	63.0	47.0	3.0	Renne et al., 1990	DAF
Kalahari	-10.6	11.6	Umkondo Igneous Province	1,105.0	66.0	37.0	3\3	Hargraves et al., 1994; Powell et al., 2001	DAF,MT,P
Kalahari	-28.5	21.6	Ezelsfontein formation	1,154.0	55.0	77.0	17.0	Renne et al., 1990	DAF
Kalahari	-25.7	28.5	Premier kimberlites	1,165.0	41.0	55.0	7.0	Powell et al., 2001	MT
Madigascar	-18.8	48.7	Carion granite	509.0	-7.0	1.0	13\17	Meert et al, 2001	MT
Madigascar	-18.0	47.0	Stratoid granite remag	521.0	-7.0	353.0	14.0	Meert et al., 2003	MT
Mexico	17.1	-97.0	Oaxaca Anorthosite	950.0	47.0	267.0	23.0	Ballard et al., 1989	P
North America	48.5	-58.5	Steel Mountain anorthosite	451.0	22.5	138.0	8\14	Murthy and Rao, 1975	W
North America	36.4	-112.5	Tapeats sandstone	508.0	-5.0	338.0	3.0	Elston and Bressler, 1977	MT
North America	38.4	-78.0	Catoctin basalts	564.0	42.0	297.0	9.0	Meert et al., 1994a,b	MT
North America	50.7	-66.6	Sept-Lles complex B	564.0	44.0	315.0	5.0	Tanczyk et al., 1987	MT
North America	46.0	-80.1	Calendar complex	575.0	46.0	301.0	6\6	Symons and Chiasson, 1991	MT
North America	53.6	-53.5	Long Range dikes (a)	615.0	11.0	344.0	18.0	Murthy et al., 1992	MT
North America	53.6	-53.5	Long Range dikes (b)	615.0	69.0	350.0	15.0	Murthy et al., 1992	MT
North America	68.3	-121.6	Brook Inlier sills	723.0	2.0	345.0	16.0	Park, 1981a,b	MT
North America	72.4	-83.0	Franklin dikes	723.0	-9.0	332.0	5.0	Chrisite and Fahrig, 1983	MT,P
North America	72.0	-112.0	Natkusiak Formation	732.0	6.0	159.0	6.0	Palmer et al., 1983; Heaman et al., 1992	P
North America	68.2	-121.5	Little dal (a+b)	778.0	9.0	320.0	11.0	Park, 1981a,b	MT

Continent	slat	slon	Name	Age	plat	plon	A95	Reference	Source[1]
North America	63.5	-232.5	Top Little Dal	778.0	-24.0	339.0	11.0	Morris and Aitken, 1982	MT
North America	64.0	-128.0	Tsezotene fm	778.0	-12.0	326.0	8.0	Park and Aitken, 1986	DAF,MT,P
North America	64.0	-128.0	Tsezotene sills	778.0	2.0	138.0	5.0	Park, 1981a,b	P
North America	43.7	-110.8	Wyoming Dykes	783.0	13.0	131.0	4.0	Harlan et al., 1997	P
North America	44.8	-77.8	Thanet gabbro (B)	800.0	20.0	159.0	8\11	Buchan, 1978	DAF,W
North America	72.7	-80.0	Borden dikes	819.0	-26.7	153.3	8.6	Christie and Fahrig, 1983	DAF,W
North America	45.4	-79.9	Haliburton intrusions (C)	820.0	-3.0	167.0	6.6	Buchan and Dunlop, 1976	DAF,W
North America	64.5	-128.0	Katherine group	855.0	-9.0	330.0	4\8	Powell et al, 1993; Park and Aitken, 1996b	DAF
North America	48.5	-58.5	Indian Head anorthosite	885.0	-9.5	158.5	15\20	Murthy and Rao, 1975	DAF,W
North America	47.5	-30.3	St. Urbain anorthosite	890.0	-2.0	154.0	7.0	Robertson and Roy, 1979	DAF,W
North America	45.6	-75.6	Gatineau Hills metamorphics	900.0	-32.0	155.0	5.0	Irving et al., 1972	MT
North America	45.4	-79.9	Haliburton intrusions (B)	900.0	24.5	172.3	15.7	Buchan and Dunlop, 1976	DAF,W
North America	44.1	-76.1	Frontenac Axis dikes	910.0	-12.0	162.0	7.0	Park and Irving, 1972	DAF,W
North America	45.5	-77.8	Umfraville gabbro	911.0	-11.0	166.0	8.2	Symons, 1978	W
North America	45.5	-77.8	Umfraville gabbro	911.0	-7.5	160.0	9.0	Palmer et al., 1979	W
North America	47.9	-79.7	Granodiorites reset	960.0	-37.0	150.0	8.0	Hyodo et al., 1986	MT,W
North America	46.1	-80.7	French River anorthosite	975.0	-26.0	135.0	10.0	Stupavsky and Symons, 1982	DAF,W
North America	47.9	-79.7	Nippissing diabase remag	975.0	-27.0	141.0	8.0	Hyodo et al., 1986	MT,W
North America	45.4	-79.9	Haliburton intrusions (A)	980.0	-36.0	142.5	6.3	Buchan and Dunlop, 1976	W
North America	46.0	-78.0	Haliburton intrusives	980.0	-36.0	143.0	6.0	Hyodo and Dunlop, 1993	DAF,MT,P
North America	47.1	79.1	Archean Greenschist reset	990.0	-5.0	152.0	11.0	Hyodo et al., 1986	W
North America	45.5	-79.9	Whitestone diorite	995.0	22.0	326.0	8\10	Hyodo and Dunlop, 1993; Dalmeyer and Sutter, 1980	DAF
North America	45.5	-79.9	Whitestone massive anorthosite (Y)	995.0	-5.0	168.0	6\10	Ueno et al., 1975	W
North America	45.5	-79.9	Whitestone quartz diorite (Z)	995.0	-22.0	146.0	8\10	Ueno et al., 1975	W
North America	46.1	-80.7	French River anorthosite	1,000.0	13.0	154.0	2.0	Stupavsky and Symons, 1982	DAF,W
North America	45.3	-20.2	Grenville thermochron Zone A	1,000.0	1.0	159.0	6.0	McWilliams and Dunlop, 1978	W
North America	46.0	-74.5	Morin anorthosite (S)	1,000.0	0.0	164.0	10.0	Irving et al., 1974	DAF,W
North America	46.2	-78.2	Mattawa Tonalitic gneiss Mto	1,009.0	-2.0	140.0	5\5.6	Hyodo and Dunlop, 1993	W
North America	46.2	-78.2	Mattawa Tonalitic gneiss Mt1	1,009.0	-21.0	127.0	5\5.6	Hyodo and Dunlop, 1993	W
North America	46.8	-70.7	Chequamegon Sandstone	1,020.0	-12.0	178.0	5.0	McCabe and Van der Voo, 1983	P
North America	46.5	-91.0	Eileen sandstones	1,020.0	20.0	156.0	10.0	Watts, 1981	MT,W
North America	46.7	-92.3	K1 Fond du Lac sandstones	1,020.0	16.0	160.0	4.0	Watts, 1981	MT,W
North America	46.6	-91.8	Middle River sandstones	1,020.0	25.0	148.0	9.0	Watts, 1981	MT,W
North America	51.0	-63.0	Allard Lake anorthosite	1,025.0	-39.0	140.0	18.0	Hargraves and Bert, 1967	DAF,W
North America	48.4	-82.7	Shenango Complex normal	1,045.0	-45.0	6.0	9\14	Costanzo-Alvarez et al., 1993	DAF
North America	48.4	-82.7	Shenango Complex reverse	1,045.0	-34.0	11.0	11\17	Costanzo-Alvarez et al., 1993	DAF
North America	48.7	-86.5	Clay-Howells carbonotite	1,075.0	27.0	179.0	7.0	Lewchuk and Symons, 1990b	DAF,MT,W
North America	47.8	-85.9	Michipicoten Island volcanics	1,075.0	25.0	175.0	8.0	Palmer and Davis, 1987	DAF,W
North America	29.7	-110.5	Arizona diabases	1,085.0	-23.0	359.0	7.8	Harlan, 1993	DAF
North America	47.8	-85.9	Mamainse Point intrusive unit	1,085.0	24.0	166.0	31.0	Palmer and Davis, 1987	W
North America	47.2	-88.5	Copper Harbor	1,087.0	-22.0	1.0	6.5	Diehl and Haig, 1994	DAF
North America	46.5	-90.0	Freda sandstone	1,087.0	1.0	180.0	1.0	Henry et al., 1977	DAF,MT,P,W
North America	47.0	-89.5	Jacobsville sandstone mean	1,087.0	-9.0	183.0	3.0	Roy and Robertson, 1978	DAF,MT,P,W
North America	47.4	-87.7	Lakeshore traps	1,087.0	22.0	181.0	7.0	Diehl and Haig, 1994	MT,P
North America	46.5	-90.0	Nonesuch shale	1,087.0	10.0	177.0	3.0	Henry et al., 1977	DAF,MT,P,W
North America	49.9	-86.2	Chipman Lake carbonotites	1,090.0	38.0	186.0	8.0	Symons, 1992	DAF,W
North America	47.8	-85.9	Mamainse Point 1	1,090.0	-39.0	357.0	4\6	Costanzo-Alvarez et al., 1993	DAF
North America	47.8	-85.9	Mamainse Point 2	1,090.0	-31.0	8.0	6\10	Costanzo-Alvarez et al., 1993	DAF
North America	47.8	-85.9	Mamainse Point 3	1,090.0	-37.0	23.0	15\20	Costanzo-Alvarez et al., 1993	DAF

Continent	slat	slon	Name	Age	plat	plon	A95	Reference	Source[1]
North America	47.8	-85.9	Mamainse Point 4	1,090.0	-61.0	39.0	13\14	Costanzo-Alvarez et al., 1993	DAF
North America	47.8	-85.9	Mamainse Point volcanics	1,090.0	38.0	188.0	1.0	Palmer and Davis, 1987	W
North America	47.4	-87.7	Portage Lake lavas	1,095.0	27.0	181.0	3.0	Halls and Pesonen, 1982	DAF,MT,P,W
North America	46.3	-90.1	Powder Mill reverse	1,095.0	39.0	218.0	5.0	Palmer and Halls, 1986	MT
North America	47.8	-90.0	Upper North Shore Volcanics	1,097.0	32.0	184.0	5.0	Halls and Pesonen, 1982; Halls and Green, 1997	P
North America	47.8	-90.0	North Shore volcanics (N)	1,098.0	-32.0	8.0		Roy, 1983	DAF
North America	47.8	-90.0	North Shore volcanics (R)	1,098.0	-47.0	20.0		Roy, 1983	DAF
North America	48.6	-88.0	Upper Osler volcanics	1,098.0	34.0	178.0	10.0	Halls, 1974	W
North America	47.2	-88.5	Copper Harbor lavas	1,100.0	35.0	176.0	3.0	Halls and Palmer, 1981	MT,W
North America	52.9	-60.0	Mealy Mountain anorthosite (A)	1,100.0	-22.9	173.4	5.0	Park and Emslie, 1983	DAF,W
North America	47.0	-89.4	mean Logan dikes	1,100.0	35.0	181.0	10.0	Halls and Pesonen, 1982	MT,P,W
North America	48.4	-88.6	Thunder Bay (N)	1,100.0	-35.0	354.0	4\7	Costanzo-Alvarez et al., 1993	DAF
North America	48.4	-88.6	Thunder Bay (R)	1,100.0	-48.0	32.0	7\8	Costanzo-Alvarez et al., 1993	DAF
North America	47.4	-87.7	Keewanawan dikes	1,102.0	44.0	197.0	11\11	Green et al., 1987	MT
North America	45.3	-77.7	Cordova gabbro (A)	1,103.0	-10.5	151.0	5.5	Dunlop and Sterling, 1985	W
North America	45.3	-77.7	Cordova gabbro (B)	1,103.0	24.0	165.0	9.5	Dunlop and Sterling, 1985	W
North America	47.7	-83.1	Lackener Lake Complex	1,105.0	-54.0	23.0	7\8	Costanzo-Alvarez et al., 1993	DAF
North America	48.0	-83.1	Nemegosenda NM1	1,107.0	-43.0	2.0	15\24	Costanzo-Alvarez et al., 1993	DAF
North America	48.0	-83.1	Nemegosenda NM2	1,107.0	-52.0	5.0	13\18	Costanzo-Alvarez et al., 1993	DAF
North America	48.0	-60.6	Firesand River	1,107.5	-27.0	3.0	6\13	Costanzo-Alvarez et al., 1993	DAF
North America	48.8	-86.5	Coldwell complex	1,108.0	49.0	200.0	16.5	Lewchuk and Symons, 1990a	MT
North America	48.8	-86.5	Coldwell complex 1	1,108.0	-54.0	37.0	5\6	Costanzo-Alvarez et al., 1993	DAF
North America	48.8	-86.5	Coldwell complex 2	1,108.0	-49.0	15.0	7\9	Costanzo-Alvarez et al., 1993	DAF
North America	48.8	-86.5	Coldwell complex 3	1,108.0	-44.0	9.0	5\7	Costanzo-Alvarez et al., 1993	DAF
North America	49.0	-88.1	mean Logan sills	1,109.0	49.0	220.0	3.0	Halls and Pesonen, 1982	DAF,MT,W
North America	44.8	-77.6	Tudor gabbro Tu1	1,110.0	17.0	137.0	8.4	Palmer and Carmichael, 1973	W
North America	44.8	-77.6	Tudor gabbro Tu2	1,110.0	12.0	133.0	4.8	Dunlop et al., 1985	W
North America	47.0	-83.3	Seabrook Lake carbonotite	1,113.0	46.0	180.0	11.0	Symons, 1992	DAF,MT,W
North America	46.0	-74.5	Morin anorthosite (P)	1,124.0	-42.0	139.0	5.3	Irving et al., 1974	DAF,W
North America	48.9	-79.5	Abitibi dikes	1,141.0	44.0	211.0	15\12	Ernst and Buchan, 1993	MT,P
North America	44.8	-77.8	Thanet gabbro (A)	1,202.0	-30.0	165.0	3.6\6.8	Buchan, 1978	W
North America	44.8	-77.8	Thanet gabbro (A1)	1,202.0	28.0	338.0	3.6\3.8	Buchan et al. 1983, Dalmeyer and Sutter, 1980	DAF
North America	44.8	-77.8	Thanet gabbro (A2)	1,202.0	32.0	352.0	5.5\9.2	Buchan et al. 1983, Dalmeyer and Sutter, 1980	DAF
North America	48.5	-71.7	Lac St. Jean anorthosite (Normal)	1,451.0	-8.0	167.0	4.6\7.3	Buchan et al., 1983	W
North America	52.9	-60.0	Mealy Mountain anorthosite (E)	1,550.0	-38.0	178.0	9.0	Fahrig et al., 1974	DAF,W
North America	52.9	-60.0	Mealy Mountain anorthosite (NW)	1,550.0	8.5	181.0	12.0	Fahrig et al., 1974	DAF,W
North America	52.9	-60.0	Mealy Mountain anorthosite (B)	?	-34.2	147.9	11.0	Park and Emslie, 1983	DAF,W
North America	47.9	-89.6	Grand Portage dikes	none	-48.0	20.0	8\10	Costanzo-Alvarez et al., 1993	DAF
North America			Grenville dikes	none	3.0	331.0		Buchan et al., 1983	DAF
North America	45.4	-79.9	Grenville Front anorthosite	none	8.0	161.0	6.3	Palmer and Carmichael, 1973	DAF,W
North America	48.5	-71.7	Lac St. Jean anorthosite (Reverse)	none	-19.0	147.0	8.5\105	Buchan et al., 1983	W
North America	44.5	-78.3	Magnetawan metasediments	none	-24.0	130.0	4.0	McWilliams and Dunlop, 1975	DAF,W
North America	48.0	-83.1	Nemegosenda NM3	none	-49.0	14.0	21\29	Costanzo-Alvarez et al., 1993	DAF
North America	45.2	-76.1	Ottowa basic intrusions	none	-32.0	155.0	8.0	Irving et al., 1972	DAF,W
North America	48.1	-82.9	Shawmere western sites	none	-53.0	9.0	15\20	Costanzo-Alvarez et al., 1993	DAF
North America	45.5	-79.9	Whitestone anorthosite, oxide segregation (W)	none	-16.0	156.0	8\11	Ueno et al., 1975	W
North America	45.0	-78.2	Wilberforce pyroxenite	none	-14.5	148.0	6.0	Palmer and Carmichael, 1973	W

Continent	slat	slon	Name	Age	plat	plon	A95	Reference	Source[1]
North America	48.6	-88.0	Lower Osler volcanics		-49.0	23.0	12\14	Costanzo-Alvarez et al., 1993	DAF
North America	46.4	-87.4	Marquette dikes		-48.0	33.0	5\6	Costanzo-Alvarez et al., 1993	DAF
Sao Francisco	-19.6	-45.4	Mean Sao Francisco Pole (SFC)	520.0	19.0	330.0	13.0	D'Agrella-Filho et al., 2000	MT
Sao Francisco	-13.0	-38.5	Salvador dikes, normal	1,003.0	9.0	121.0	15.0	D'Agrella-Filho and Pacca et al., 1994	DAF
Sao Francisco	-15.4	-39.0	Ilheus dikes (SFC)	1,012.0	30.0	100.0	3.0	D'Agrella-Filho et al., 1990	W
Sao Francisco	-13.0	-38.5	Salvador dikes, reverse	1,021.0	18.0	228.0	13.0	D'Agrella-Filho and Pacca et al., 1994	DAF
Sao Francisco	-15.2	-39.7	Itaju do Colonia	1,050.0	8.0	111.0	6.0	D'Agrella-Filho et al., 1990	W
Sao Francisco	-15.2	-39.0	Olivenca dikes - N (SFC)	1,050.0	16.0	107.0	5.0	D'Agrella-Filho et al., 1990	W
Sao Francisco	-15.2	-39.0	Olivenca dikes - R (SFC)	1,078.0	-10.0	100.0	6.0	D'Agrella-Filho et al., 1990	W
Siberia	67.5	105.0	Moyero River Seds	490.0	-37.0	139.0	6.0	Gallet and Pavlov, 1996	MT
Siberia	71.0	122.5	Yuryakh Fm	500.0	-36.0	140.0	5.0	Pisarevsky et al., 1998	MT
Siberia	68.0	88.6	Kulumbe River	503.0	-42.0	136.0	2\3	Pavlov and Gallet, 2001	MT
Siberia	71.0	122.5	Ekreket Fm	510.0	-45.0	159.0	7.0	Pisarevsky et al., 1998	MT
Siberia	59.0	135.0	Inican	538.0	-46.0	162.0	4.0	Osipova, 1986	MT
Siberia	71.0	122.5	Kessyusa	545.0	-38.0	165.0	13.0	Pisarevsky et al., 1998	MT
Siberia	47.0	90.6	Tsagan-Olom	565.0	23.0	203.0	11\22	Kravchinsky et al., 2001	MT
Siberia	54.0	108.0	Cisbaikalia	615.0	-3.0	168.0	9.0	Pisarevsky et al., 2000	MT
Siberia	78.1	110.0	Minya Fm	615.0	34.0	217.0	9\15	Kravchinsky et al., 2001	MT
Siberia	52.1	103.8	Shaman Fm	615.0	32.0	251.0	7\14	Kravchinsky et al., 2001	MT
Siberia	58.9	136.0	Sette-Daban Sills/Kandyk Fm	974.0	-4.0	177.0	2.0	Pavlov et al., 1992	MT
Siberia	66.0	88.0	Turukhansk sediments	987.0	-15.0	256.0	8.0	Gallet et al., 2000	P
Siberia	68.0	89.0	Uchur-Maya (Malgin) sediments	1,070.0	-25.0	231.0	3.0	Gallet et al., 2000	P,MT
South China	30.2	119.7	Hetang Fm	511.0	-3.0	16.0	17.0	Lin et al., 1985	MT
South China	30.8	111.2	Tianheban Fm	511.0	-7.0	10.0	23.0	Lin et al., 1985	MT
South China	24.4	102.3	Meishucun Fm	525.0	9.0	31.0	10.0	Lin et al., 1985	MT
South China	22.5	105.0	Nantuo Fm	740.0	0.0	331.0	5.0	Rui and Piper, 1997	MT
South China	30.5	111.1	Liantuo	748.0	-4.0	341.0	13.0	Evans et al., 2000	P
Trans-Sarara	26.2	33.5	Dokhan volcanics (ANS)	593.0	-43.0	36.0	10.0	Davies et al., 1980; Naim et al., 1987	MT
Trans-Sarara	19.0	37.0	Suakin gabbros (CC)	841.0	25.0	134.0	8.0	Reischmann et al., 1992	MT

Table 2. List of Euler poles and angles of rotation (with respect to Laurentia) for continents shown in the Rodinia reconstruction proposed here for the time frame 1,200-850 Ma (fig. 14). Table lists: continent, Euler pole latitude (elat), Euler pole longitude (elon), and rotation about the Euler pole (rot). Rotations are given as clockwise (negative) or counterclockwise (positive). Shaded continents are ones for which no paleomagnetic data exists (in the compilation used here). Light shading indicates continents (East Africa, Northwest Africa, Trans-Sahara) which were rotated with Congo.

Continent	elat	elon	rot
Amazonia	3	334	-92
Antarctica	-23	199	110
Australia	-11	7	-121
Baltica	-84	180	48
Congo	–16.2	166	145
East Africa	–16.2	166	145
Greenland	68	-118	-14
India	2	359	174
Iran	-11	178	178
Kalahari	–16.2	166	145
Khazakstan	–59	191	140
Madagascar	–23.7	180	127
Northwest Africa	–16.2	166	145
Rio Plata.	3	334	-92
Sao Francisco	3	334	-92
Siberia	-65	204	155
South China	-44	162	-174
Trans-Sahara	–16.2	166	145