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Earth-Science Reviews 67 (2004) 91–123

EARTH-SCIENCE REVIEWS

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A Paleo-Mesoproterozoic supercontinent: assembly, growth and breakup

Guochun Zhao[a,*], Min Sun[a], Simon A. Wilde[b], Sanzhong Li[c]

[a] Department of Earth Sciences, The University of Hong Kong, Pokfulam Road, James Lee Science Building, Hong Kong
[b] Department of Applied Geology, Curtin University of Technology, GPO Box U1987, Perth 6845, WA, Australia
[c] College of Earth Sciences, Ocean University of China, Qingdao 266603, China

Received 8 July 2003; accepted 23 February 2004

Abstract

Geological and paleomagnetic data support the hypothesis that a Paleo-Mesoproterozoic supercontinent, referred to as Columbia, existed before the formation of Rodinia. This pre-Rodinia supercontinent was assembled along global-scale 2.1–1.8 Ga collisional orogens and contained almost all of Earth's continental blocks. Following its final assembly at ~ 1.8 Ga, the supercontinent Columbia underwent long-lived (1.8–1.3 Ga), subduction-related growth via accretion at key continental margins, forming a 1.8–1.3 Ga large magmatic accretionary belt along the present-day southern margin of North America, Greenland and Baltica. It includes the 1.8–1.7 Ga Yavapai, Central Plains and Makkovikian Belts, 1.7–1.6 Ga Mazatzal and Labradorian Belts, 1.5–1.3 Ga St. Francois and Spavinaw Belts and 1.3–1.2 Ga Elzevirian Belt in North America; the 1.8–1.7 Ga Ketilidian Belt in Greenland; and the 1.8–1.7-Transscandinavian Igneous Belt, 1.7–1.6 Ga Kongsberggian–Gothian Belt, and 1.5–1.3 Ga Southwest Sweden Granitoid Belt in Baltica. Other cratonic blocks also underwent marginal outgrowth at about the same time. In South America, a 1.8–1.3 Ga accretionary zone occurs along the western margin of the Amazonia Craton, represented by the Rio Negro, Juruena and Rondonian Belts. In Australia, 1.8–1.5 Ga accretionary magmatic belts, including the Arunta, Mt. Isa, Georgetown, Coen and Broken Hill Belts, occur surrounding the southern and eastern margins of the North Australia Craton and the eastern margin of the Gawler Craton. In China, a 1.8–1.4 Ga accretionary magmatic zone, called the Xiong'er belt (Group), extends along the southern margin of the North China Craton. Fragmentation of this supercontinent began about 1.6 Ga ago, associated with continental rifting along the western margin of Laurentia (Belt–Purcell Supergroup), southern margin of Baltica (Telemark Supergroup), southeastern margin of Siberia (Riphean aulacogens), northwestern margin of South Africa (Kalahari Copper Belt), and northern margin of North China (Zhaertai–Bayan Obo Belt). The fragmentation corresponded with widespread anorogenic magmatic activity, forming anorthosite–mangerite–charnockite–granite (AMCG) suites in North America, Baltica, Amazonia and North China, and continued until the final breakup of the supercontinent at about 1.3–1.2 Ga, marked by the emplacement of the 1.27 Ga MacKenzie and 1.24 Ga Sudbury mafic dike swarms in North America.

Keywords: supercontinent; Paleo-Mesoproterozoic; assembly; accretion; breakup; reconstruction

* Corresponding author. Tel.: +86-852-28578203; fax: +86-852-25176912.
E-mail address: gzhao@hkucc.hku.hk (G. Zhao).

0012-8252/$ - see front matter
doi:10.1016/j.earscirev.2004.02.003

1. Introduction

In 1912, Alfred Wegener, a German meteorologist, proposed the hypothesis of continental drift to describe the movement of major landmasses across the surface of the Earth (Wegener, 1912). He explained the circum-Atlantic continents as megablocks derived from fragmentation of a much large unit referred to as the supercontinent Pangea (meaning "all Earth"). Initially, Wegener's hypothesis was ridiculed, but half a century later it became, in a modified form, fully acknowledged and now has been incorporated into the current theory of plate tectonics. By the 1970s, most earth scientists had accepted Wegener's concept of Pangea, which was considered to have resulted from the assembly of Earth's continents in the time range of 300–250 Ma, and which consisted of Gondwana (Australia, India, Sri Lanka, Madagascar, East Antarctica, South America and Africa) as its southern half, and Laurasia (North America, Greenland and Eurasia) as its northern half (Fig. 1a). Later, it was perceived that similar concentrations of continental blocks might also have occurred during other periods in Earth's history (Anderson, 1982; Piper, 1982). Nance et al. (1986, 1988) even proposed that at intervals of 400–500 Ma, the Earth underwent periods of continental assembly and dispersion as a result of plate convergence and separation. Since the 1980s, the notion of Proterozoic supercontinents has attracted much attention. Piper (1982, 1987) produced paleomagnetic evidence for the existence of a long-lived Proterozoic supercontinent. Hoffman (1989a) provided geological evidence for a supercontinent that was assembled in the period 2.0–1.8 Ga. In 1990, McMenamin and MacMenamin outlined growing evidence for a Neoproterozoic supercontinent, named Rodinia, from a Russian word meaning "to beget", and 1 year later, Dalziel (1991): Hoffman (1991) and Moores (1991) nearly at the same time proposed the configuration of Rodinia in which Laurentia (North America and Greenland) forms the core of the supercontinent (Fig. 1b). Since then, Rodinia reconstruction has dominated discussion of supercontinents. There is a broad agreement that the assembly of Rodinia was completed by the global-scale Grenvillian Orogeny or its age-equivalent collisional orogens at approximately 1.0 Ga (Powell et al., 1993, 2001; Rogers, 1993; Dalziel, 1992, 1995, 1997; Dalziel and

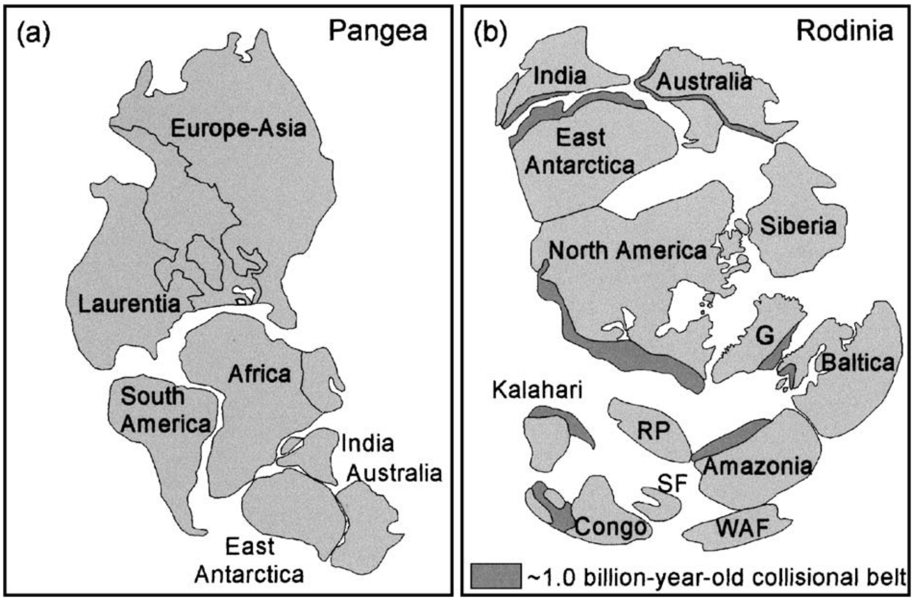

Fig. 1. Two supercontinents in Earth's history: (a) Pangea formed 300–250 million years ago (Rogers et al., 1995), and (b) Rodinia formed ∼ 1.0 billion years ago (Dalziel, 1997). Abbreviations: G—Greenland; RP—Rio de la Plata; SF—Sao Francisco; WAF—West Africa.

Gahagan, 1996; Hoffman, 1999; Dalziel et al., 2000; Meert and Powell, 2001; Nast, 2002).

On the other hand, geologists have noted that most Rodinia reconstructions are established on the basis of Paleo- to Mesoproterozoic connections between continental blocks (e.g. Laurentia vs. Siberia, Laurentia vs. Australia, Laurentia vs. Antarctica, Amazonia vs. Baltica, West Africa vs. Amazonia). Moreover, most Rodinian fragments contain abundant evidence that they are a collage of earlier collision events, mostly occurring between 2.1 and 1.8 Ga (Rogers et al., 1995; Nast, 1997; Zhao and Cawood, 1999). These have led some geologists to reconsider Hoffman's (1989a) early speculation that a Paleo-Mesoproterozoic supercontinent may have existed before Rodinia (e.g. Windley, 1995; Rogers, 1996; Condie, 1998, 2000; Zhao and Cawood, 1999; Zhai et al., 2000; Luepke and Lyons, 2001). Although the exact details about the history of this proposed supercontinent still remain shadowy, some important advances have been made in understanding its initial amalgamation, subsequent outgrowth and final fragmentation (Condie, 2002; Rogers and Santosh, 2002, 2003; Hartmann, 2002; Meert, 2002; Sears and Price, 2002; Rao and Reddy, 2002; Wilde et al., 2002; Zhao et al., 2002a, 2003a,b). The purpose of this paper is to summarize and analyze critical evidence for the assembly, subsequent growth and final breakup of such a pre-Rodinia supercontinent, and discuss some key issues related to its full reconstruction.

2. Paleo-Mesoproterozoic supercontinent—Columbia

In his analysis of Proterozoic Laurentia, Hoffman (1989a) concluded that in the period 2.0–1.8 Ga, seven microcontinents in North America and Greenland were assembled to form a Laurentia protocraton, which possibly included Baltica and other accreted terranes, possibly forming an extremely large landmass that was named "Nena" by Gower et al. (1990). Zhao et al. (2002a) extended this hypothesis to other cratonic blocks and proposed that in the period 2.1–1.8 Ga, the Earth's continents existing at that time began to collide and finally coalesce into a pre-Rodinian supercontinent. Rogers and Santosh (2002) named this supercontinent "Columbia" because they thought some of the best evidence for its existence came from matching patterns of two pairs of rifts in the Columbia River region of western North America and eastern India. In North America, these are called the Belt–Purcell and the Uinta rifts, and in eastern India, they are known as the Mahanadi and the Godavari rifts. In both countries, the two pairs of rifts are about 500 km apart. Moreover, geochronological data suggest that these rifts are the same age; all formed about 1.5 Ga ago. These similarities led Rogers and Santosh (2002) to propose that eastern India was once connected to the western side of North America as part of Columbia. Fig. 2a is a preliminary configuration of Columbia as proposed by Rogers and Santosh (2002), in which South Africa, Madagascar, India, Australia and attached parts of Antarctica are placed adjacent to the western margin of North America, whereas Greenland, Baltica (Northern Europe) and Siberia are positioned adjacent to the northern margin of North America, and South America is placed against West Africa.

Most recently, two rhyolitic volcanic horizons from the Vindhyan Supergroup in eastern India have revealed U–Pb zircon ages of 1631 $\pm$ 5 and 1631 $\pm$ 1 Ma (Ray et al., 2002), older than the supposed correlative Belt Supergroup (1470 Ma). In addition, if Australia was not connected to the western margin of Laurentia as in the Rogers and Santosh configuration, the 1.6 Ga-age detritus that dominates the fill of the Belt–Purcell basins in North America would have no known sources. Detritus of this age can be easily obtained from central Australia but not from eastern India so far as is known. Therefore, the eastern India–western North America connection still needs a further geological and geochronological testing.

Alternatively, Zhao et al. (2002a) proposed another configuration of Columbia (Fig. 2b), in which the fits of Baltica and Siberia with Laurentia and the fit of South America with West Africa are similar to those of the Rogers and Santosh (2002) configuration, whereas the fits of India, East Antarctica and Australia with Laurentia are similar to their corresponding fits in the configuration of Rodinia (e.g. Dalziel, 1997). This configuration is based on the available geological reconstructions of 2.1–1.8 Ga orogens and related Archean cratonic blocks, especially on those reconstructions between South

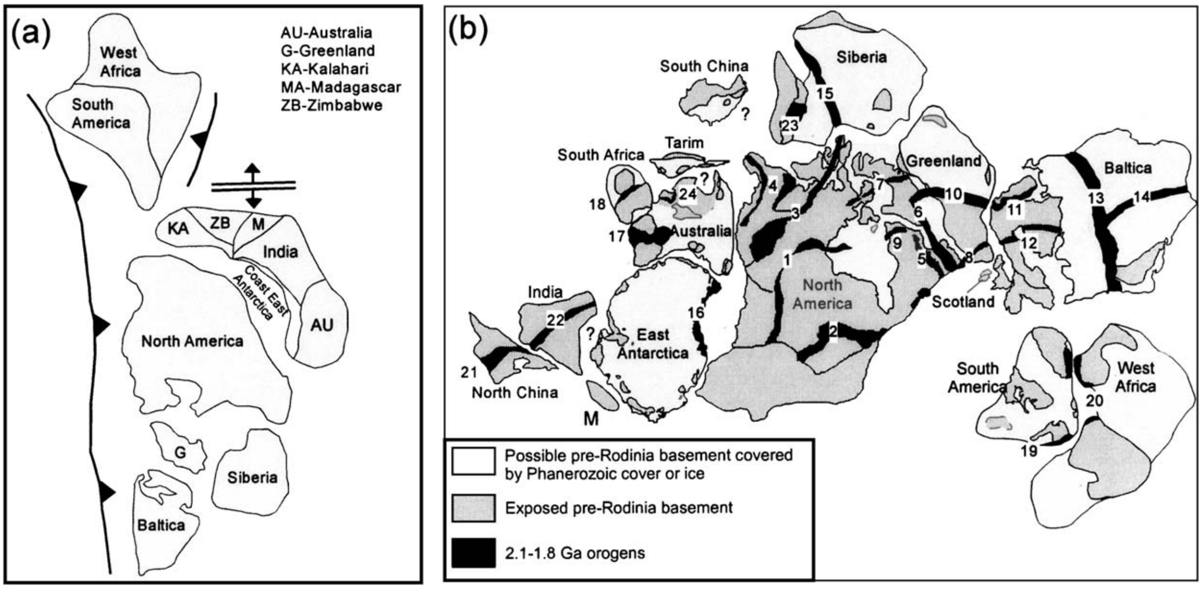

Fig. 2. Possible alternatives for the reconstruction of a Paleo-Mesoproterozoic supercontinent, named "Columbia". (a) Reconstruction of Rogers and Santosh (2002); (b) Reconstruction of Zhao et al. (2002a). Symbols: (1) Trans-Hudson; (2) Penokean; (3) Taltson–Thelon; (4) Wopmay; (5) New Quebec; (6) Torngat; (7) Foxe; (8) Makkovik–Ketilidian; (9) Ungava; (10) Nugssugtoqidian; (11) Kola–Karelian; (12) Svecofennian; (13) Volhyn–Central Russian Orogen; (14) Pachelma; (15) Akitkan; (16) Transantarctic Orogen; (17) Capricorn; (18) Limpopo Belt; (19) Transamazonian; (20) Eburnean; (21) Trans-North China Orogen; (22) Central Indian Tectonic Zone; (23) Central Aldan Orogen; (24) Halls Creek Orogen.

America and West Africa (Bullard et al., 1965; Onstott and Hargraves, 1981; Bertrand and Jardim de Sá, 1990; Ledru et al., 1994; Rogers, 1996; Zhao et al., 2002b); Western Australia and South Africa (Button, 1976; Cheney, 1996; Aspler and Chiarenzelli, 1998; Zegers et al., 1998); Laurentia and Baltica (Hoffman, 1988, 1989b, 1991; Winchester, 1988; Gower et al., 1990; Van Kranendonk et al., 1993; Park, 1994, 1995; Buchan et al., 2000); Siberia and Laurentia (e.g. Sears and Price, 1978, 2000, 2002; Condie and Rosen, 1994; Frost et al., 1998); Laurentia and Central Australia (e.g. Dalziel, 1991; Young, 1992; Burrett and Berry, 2000, 2002; Karlstrom et al., 1999, 2001; Wingate et al., 2002); East Antarctica and Laurentia (Dalziel, 1991; Moores, 1991), North China and India (Zhao et al., 2003a). A detailed review on these geological reconstructions has been given by Zhao et al. (2002a) and is not repeated here. Of these reconstructions, the fits of Baltica and Siberia with Laurentia, South America with West Africa, and Southern Africa with Western Australia are also consistent with paleomagnetic data.

3. Paleomagnetic evidence

In the late 1970s and early 1980s, the consensus of opinion by most paleomagnetists was that the West African and Congo Cratons in West Africa and the Amazonia and São Francisco Cratons in South America were assembled by at least 2.0 Ga (McElhinny and McWilliams, 1977; Onstott and Hargraves, 1981; Piper, 1982; Onstott et al., 1984). This view was reached on the basis of the paleomagnetic data of Onstott and Hargraves (1981) and Onstott et al. (1984), which show that coeval rocks between 2.1 and 1.5 Ga in the Amazonia and West African Cratons record similar paleomagnetic polar wander paths if a ~ 1000-km right-lateral strike-slip movement is assumed to have occurred between the two cratons. Furthermore, Onstott et al. (1984) showed that the

2.0–1.9 Ga paleomagnetic poles of the Amazonia and West African Cratons are distinctly different from the poles of the Kalahari Craton at similar times, suggesting that relative motion has occurred between the West African and Kalahari Cratons since that time, or else the adjoining Amazonia–West African Craton and the Kalahari Craton were not contiguous at that time. Although the early paleomagnetic reconstruction of the Amazonia and West African Cratons was once questioned because of the poor quality of the data, recent paleomagnetic data from single cratons and terranes still support the conclusion that the two cratons belonged to a single continent at ∼2.0 Ga. For example, D'Agrella et al. (1996) obtained a paleomagnetic pole from the ∼2.0 Ga rocks in the Ogooué Formation of the Congo Craton in West Africa, and a comparison of this pole position with the paleomagnetic pole of similar age obtained for the granulites from the Jequie complex of the Sao Francisco Craton allows a close affiliation between the Sao Francisco and Congo Cratons, suggesting that both cratons belonged to the same landmass at that time.

Reviews of early paleomagnetic data from Laurentia and Baltica indicate that in the period 1.9–1.2 Ga, most cratonic blocks in these continents showed largely consistent apparent polar wandering paths (APWPs), which were mostly restricted to a ±30° paleolatitude range (Poorter, 1976; Patchett et al., 1978; Irving, 1979; Piper, 1982, 1987), suggesting that these cratonic blocks may have existed as a single continental landmass during this period. By comparing the Baltica and Laurentia apparent polar wandering paths (APWPs) of Svecofennian/Hudsonian to Jotnian/Mackenzie age, Poorter (1976) proposed a Mesoproterozoic reconstruction with northern Baltica adjacent to southwestern Greenland and southern Labrador. Patchett et al. (1978) obtained a somewhat similar fit by using 1.26–1.19 Ga Jotnian palaeopoles of Baltica and Mackenzie-aged paleopoles of Laurentia. Most of these early paleomagnetic poles were not precisely constrained by reliable geochronology, and thus, these early paleomagnetic reconstructions have required significant modification by later good-quality paleomagnetic data. However, the basic conclusion that Laurentia and Baltica have fairly consistent APWPs during much of the Mesoproterozic is still supported by recent data (Symons, 1991; Elming, 1994; Smethurst et al., 1998; Gala et al., 1998; Buchan et al., 2000, 2001; Elming and Mattsson, 2001). For example, Buchan et al. (2000, 2001) show that the key paleopole at 1265 Ma from Jotnian sills and dykes of the Central Scandinavian Dolerite complex matches the key 1267 Ma Mackenzie dyke from Laurentia, which is consistent with reconstructions in which Baltica is adjacent to present-day east Greenland. Based on these new paleomagnetic constraints combined with geological considerations, Buchan et al. (2000) proposed a Laurentia–Baltica reconstruction, as shown in Fig. 3, whereby the eastern margin of the Gothian belt of Baltica strikes towards its "natural" continuation through the northern British Isles and Rockall Bank, and along the margin of Labradorian and Ketilidian belts of Laurentia. In addition, recent paleomagnetic data also indicate that collisions between the Archean components of Laurentia may have occurred during the Paleoproterozoic (Symons, 1991, 1994). For example, Dunsmore and Symons (1990), using paleomagnetic data, noted that the minimum 'north–south' width of the Manikewan Ocean, which is assumed to have intervened between the Superior Craton and the Slave–Hearne–Rae Cratons, was at least 4800 km before 1.85 Ga. This implies that these Archean blocks were not amalgamated to form a coherent craton until about 1.85 Ga ago (Symons, 1991, 1994).

Paleomagnetic data indicate that during the Mesoproterozoic, Siberia was also at low latitudes restricted to a ±30° paleolatitude range (Piper, 1982; Smethurst et al., 1998; Ernst et al., 2000), broadly similar to latitudes determined for Laurentia and Baltica (Irving, 1979; Piper, 1982; Elming, 1994; Gala et al., 1998; Buchan et al., 2000). Paleomagnetic data cannot constrain longitude, and thus divergent paleomagnetic reconstruction models for relative locations of Siberia and Laurentia or Baltica have been postulated (e.g. Poorter, 1981; Piper, 1982; Scotese and McKerrow, 1990; Smethurst et al., 1998; Ernst et al., 2000). For example, Piper (1982) located Siberia adjacent to southwestern Laurentia, similar to the geological reconstruction by Sears and Price (1978), Scotese and McKerrow (1990) suggested that Siberia lay near eastern Greenland; and Smethurst et al. (1998) placed Siberia close to eastern Baltica rather than Laurentia. More recently, Ernst et al. (2000) obtained good-quality paleomagnetic data from the 1503 Ma Kuonanmka and 1384 Ma Chieress swarms.

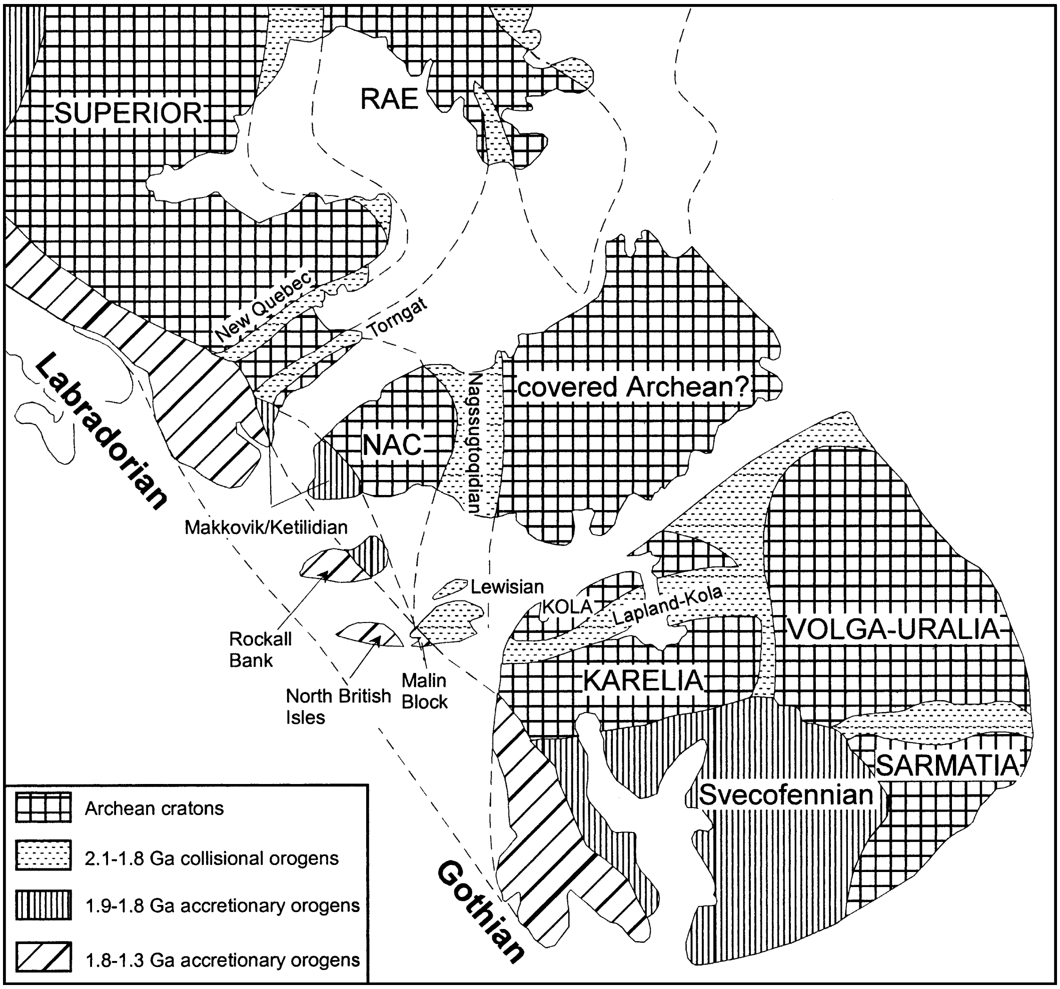

Fig. 3. Reconstruction of Laurentia and Baltica at 1267 Ma based on geological criteria within the paleomagnetic constraints (Buchan et al., 2000). NAC is North Atlantic Craton.

These data locate the Anabar Shield, and perhaps the whole of Siberia, at low latitude during the early Mesoproterozoic (Fig. 4a), whereas well-constrained paleomagnetic data for Laurentia also place North America and Greenland at low latitudes at 1460–1420, 1320–1290, and 1267 Ma (Fig. 4b; Symons, 1991; Elming, 1994; Smethurst et al., 1998; Gala et al., 1998; Buchan et al., 2000). Thus, these new data further confirm the early paleomagnetic conclusion that Laurentia and Siberia drifted together throughout the Mesoproterozoic. However, these paleomagnetic data do not provide a reliable early Mesoproterozoic reconstruction or discriminate between the different models for Siberia and Laurentia, as proposed by several studies using geological criteria (Poorter, 1981; Scotese and McKerrow, 1990; Smethurst et al., 1998; Ernst et al., 2000).

A late Archean to Paleoproterozoic (2.8–2.1 Ga) link between South Africa and Western Australia was first proposed by Button (1976) and then by Cheney (1996), based on comparisons of sequence stratigraphy. As reviewed by Cheney (1996), the Kaapvaal and Pilbara Cratons show a three-fold lithostratigraphic similarity between 2.8 and 2.1 Ga. The lowest sequence is 2.77–2.71 Ga arkosic sandstones and basalts that comprise the Pniel/Venters-

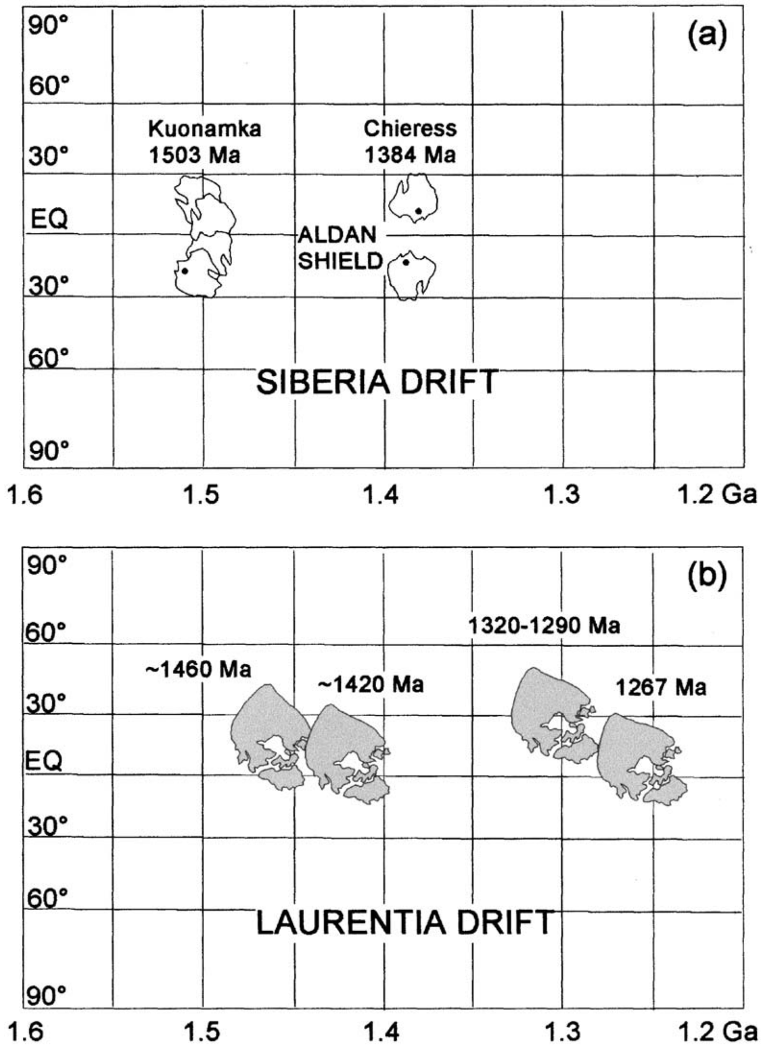

Fig. 4. Comparison of latitudinal drift and rotation of (a) Siberia and (b) Laurentia in the early Mesoproterozoic based on paleomagnetic data (Ernst et al., 2000). See text for explanation.

dorp Supergroup in Kaapvaal and the Fortescue Group in the Pilbara. The second sequence consists predominantly of 2.68–2.43 Ga quartz arenites, dolomites and banded iron formations that in the Kaapvaal Craton comprise the Ghaap/Griqualand Groups, and in the Pilbara comprise the Hamersley Group. The third sequence is 2.47–2.20 Ga siliciclastic and volcanic units with basal glacigene deposits that in the Kaapvaal Craton comprise the Postmasburg Group (or the lower Pretoria Group), and in the Pilbara comprise the lower part of the Wyloo Group. These similarities led Cheney (1996) to propose that the now widely separated Kaapvaal and Pilbara Cratons were once continuous, forming part of a continent, named Vaalbara. This hypothesis is supported by the paleomagnetic data of Zegers et al. (1998) who showed that, when the virtual paleomagnetic poles (at ~ 2.87 Ga) of two major igneous layered complexes, the Millinda Complex (2860 ± 20 Ma, U–Pb zircon) in the Pilbara and the Usushwana Complex (2871 ± 30 Ma, U–Pb zircon) in the Kaapvaal, are placed on the present N-pole, the two cratons are brought into close proximity, consistent with Cheney's Vaalbara model.

4. Assembly of the supercontinent Columbia

Like Rodinia, which is interpreted to have been assembled along the globally distributed 1.1–0.9 Ga "Grenvillian" orogens (Dalziel, 1991, 1997; Hoffman, 1991; Dalziel et al., 2000; Meert and Powell, 2001), the supercontinent Columbia is considered to have assembled along the globally distributed 2.1–1.8 Ga collisional orogens (Rogers and Santosh, 2002; Condie, 2002). Zhao et al. (2002a) have given a thorough overview on the tectonic histories of global 2.1–1.8 Ga collisional orogens and their bearing on the assembly of the supercontinent Columbia, which can be briefly summarized as follows.

Archean blocks in North America were sutured along the 1.95–1.80 Ga Trans-Hudson Orogen or its age-equivalent orogens (Fig. 5a; Hoffman, 1988, 1989a, 1989b; Van Kranendonk et al., 1993; Park, 1995; St-Onge et al., 2001). For example, the Trans-Hudson Orogen, the largest and best-exposed Paleoproterozoic orogenic belt in North America was formed in an Andean-type convergent collision of the Archean Slave–Rae–Hearne (SRH) and Superior Cratons (Fig. 5a; Hoffman, 1988, 1989b), which resulted in the closure of the intervening Manikewan ocean, entrapping the intervening Paleoproterozoic juvenile terranes and generating the Wathaman batholith as the ensialic magmatic arc at the suture zone, and thus forming the bounding fold belts at the cratonic margins (Hoffman, 1990; Symons, 1994). The 1.9–1.8 Ga Penokean Orogen is considered to have resulted from collision between the Archean Superior Craton and the Mashfeld terrane (Fig. 5a; Holm et al., 1988; Windley, 1992). The ~1.9 Ga Taltson–Thelon Orogen formed as a result of collision between the Archean Slave and Rae Cratons (Fig. 5a; Hoffman, 1988; 1989b). The 1.95–1.84 Ga Wopmay Orogen developed as a result of collision between the Archean Slave Craton and the Hottah terrane of the Nahanni continental block (Hoffman and Bowring, 1984). In the northeastern part of North America, there are a number of narrow Paleoproterozoic mobile belts, including the New Quebec, Foxe, Makkovik, Ungava and Torngat Orogens (Fig. 5a). These orogens are interpreted as 1.9–1.8 Ga continent–continent or arc–continent collisional belts joining the Archean Superior, Rae and Nain Cratons and associated arcs (e.g. Narsajuq arc; Van Kranendonk et al., 1993; St-Onge et al., 2001). It deserves mentioning that controversy still remains about whether all of the Trans-Hudson orogen in North America resulted from collision of two discrete continental blocks. For example, Aspler and Chiarenzelli (1998) and Chiarenzelli et al. (1998) interpreted the Trans-Hudson belt as an orogen that opened and then closed back on itself around the same pole of rotation around which it opened. Similarly, Chacko et al. (2000) and De et al. (2000) regarded the Taltson–Thelon Orogen as intracontinental rather than a plate margin.

Like the Archean cratonic blocks in North America, other Archean blocks around the world were also welded along Paleoproterozoic (2.1–1.8 Ga) collisional belts. In Greenland, the Archean North Atlantic and Disko Cratons were sutured along the 1.9–1.8 Ga, east–west trending, Nagssugtoqidian Orogen (Fig. 5b; Kalsbeek et al., 1987; Kalsbeek and Nutman, 1996; Kalsbeek, 2001; Bridgwater et al., 1990; Whitehouse et al., 1998). In Baltica (Eastern Europe), the Kola, Karelia, Volgo–Uralia and Sarmatia (Ukrainian) Cratons were joined by the 1.9–1.8 Ga Kola–Karelia, Volhyn–Central Russian and Pachelma Orogens (Fig. 5c; Berthelsen and Makers, 1986; Bogdanova, 1993, 1999; Bogdanova et al., 1996). In Siberia, the Aldan Shield and other Archean blocks were connected by the 1.9–1.8 Ga Akitkan Orogen (Fig. 5d; Rosen et al., 1994; Condie and Rosen, 1994; Frost et al., 1998; Jahn et al., 1998). In South America and West Africa, Amazonia and São Francisco Cratons collided, respectively, with West African and Congo Cratons along the 2.1–2.0 Ga Transamazonian and Eburnean Orogens (Fig. 5e and f; Swapp and Onstott, 1989; Bertrand and Jardim de Sá, 1990; Feybesse and Milesi, 1994; Ledru et al., 1994; Alkmim and Marshak, 1998; Zhao et al., 2002b). In Australia, the Yilgarn and Pilbara Cratons collided along the 2.0–1.9 Ga Capricorn Orogen (Fig. 5g; Myers, 1990; Tyler and Thorne, 1990; Myers et al., 1996; Occhipinti et al., 1998; Pirajno et al., 1998; Cawood and Tyler, in press), and the Kimberley and North Australian Cratons were amalgamated along the ~1.85 Ga Halls Creek Orogen (Bodorkos et al., 1999, 2000, 2002; Sheppard et al., 1999). In India, the South and

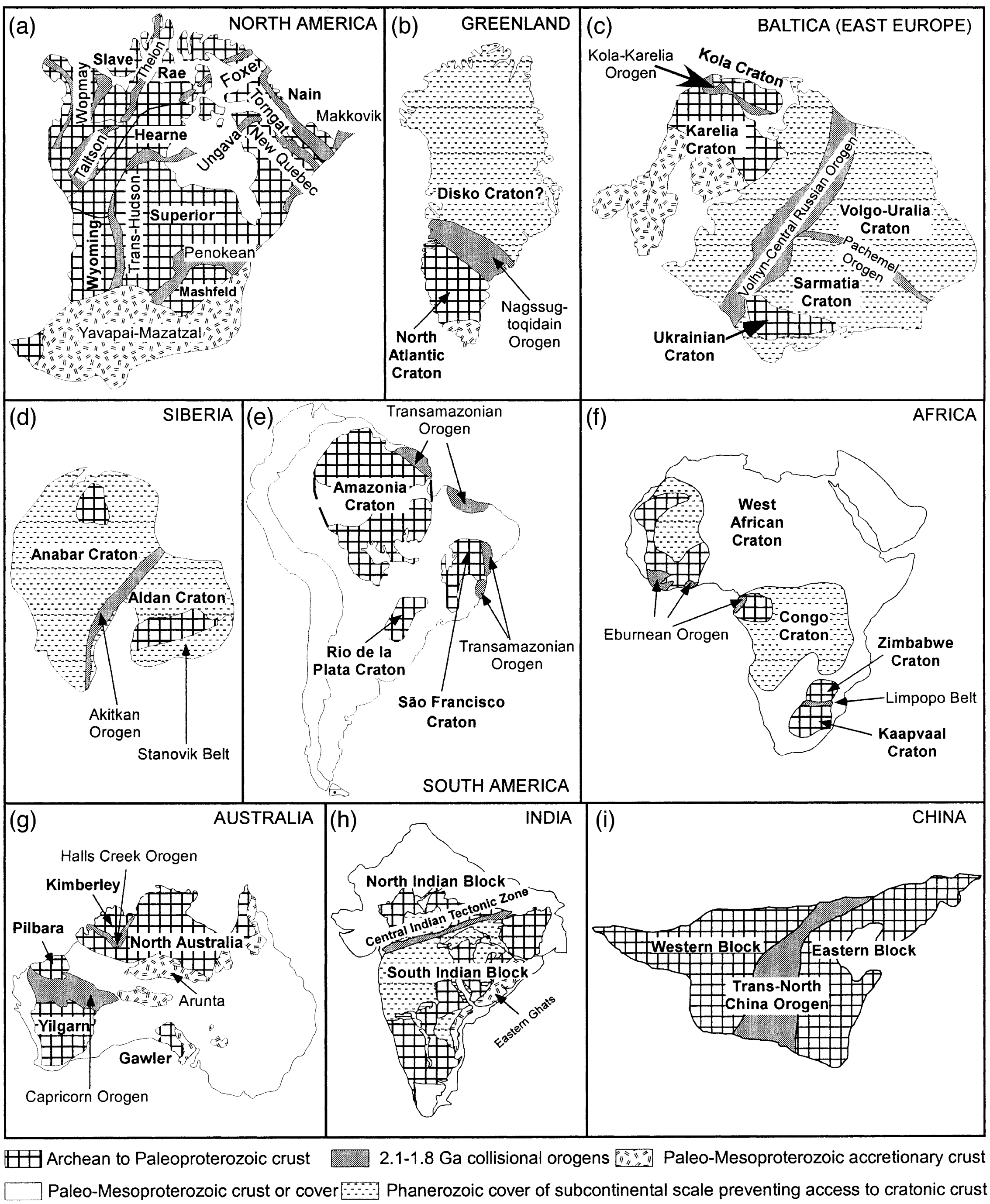

Fig. 5. Schematic tectonic map showing distribution of major 2.1 – 1.8 Ga orogens and associated cratonic blocks. Data sources: (a–b), Hoffman (1988, 1989b); (c) Bogdanova (1993); (d) Frost et al. (1998); (e) Alkmim and Marshak (1998) and Sadowski (2000); (f) Bertrand and Jardim de Sá (1990); (g) Myers et al. (1996) and Bodorkos et al. (1999, 2000); (h) Zhao et al. (2003a); and (i) Zhao et al. (2001a).

North Indian Blocks were combined along the ~1.8 Ga Central Indian Tectonic Zone (Fig. 5h; Jain et al., 1991; Yedekar et al., 1990; Rogers and Gird, 1997; Bhowmik et al., 1999; Acharyya, 2003; Bhowmik and Roy, 2003). In North China, the discrete Archean to Paleoproterozoic Eastern and Western Blocks independently developed during the Archean and collided along the Trans-North China Orogen at ~1.85 Ga (Fig. 5i; Zhao, 2001; Zhao et al., 1999a,b, 2000, 2001a,b, 2002c; Guo and Zhai, 2001; Guo et al., 1999, 2002; Liu et al., 2002; Wilde et al., 2002; Kröner, 2002; Kröner et al., 2001, 2004). In South Africa, the Limpopo Belt was previously referred to as a classic example of an Archean collisional belt, resulting from the collision of the Kaapvaal and Zimbabwe Cratons at 2.6 Ga ago (Light, 1982). Recently, however, a ~2.0–1.9 Ga tectonothermal event has been revealed for the Limpopo Belt by directly dating garnet, clinopyroxene and titanite using the Pb–Pb and Sm–Nd garnet chronometers and the Ar–Ar step heating technique for amphibole (Kamber et al., 1995), the Pb stepwise leaching of metamorphic minerals (Holzer et al., 1998) and the SHRIMP zircon U–Pb technique (Kröner et al., 1999). Together with lithological, structural and metamorphic data and previously published ages, these new results led to a reinterpretation of the tectono-metamorphic history of the Limpopo Belt, with the main phase of collision between the Zimbabwe and Kaapvaal cratons occurring between 2.0 and 1.9 Ga (Holzer et al., 1998; Kröner et al., 1999). Therefore, 2.1–1.8 Ga collisional orogens have been recognized on nearly every major continental block; they are considered to have recorded global-scale collisional events that resulted in the final assembly of Columbia from a jumble of Archean to Paleoproterozoic cratonic blocks (Condie, 2002; Rogers and Santosh, 2002, 2003; Zhao et al., 2002a).

5. Growth of the supercontinent Columbia

Ample evidence shows that, following its assembly at ~1.8 Ga, the supercontinent Columbia underwent long-continued (1.8–1.2 Ga) accretion along some of its continental margins.

In North America, Greenland, Scotland and Baltica, a 1.8–1.2 Ga magmatic accretionary zone extends from Arizona through Colorado, Michigan, southern Greenland, Scotland, Sweden and Finland to western Russia, bordering the present southern margin of North America, Greenland and Baltica (Gower et al., 1990; Park, 1992; Karlstrom et al., 2001; Rogers and Santosh, 2002). As shown in Fig. 6, this mega-magmatic zone comprises the 1.8–1.7 Ga Yavapai and Central Plains Belts, 1.7–1.6 Ga Mazatzal Belt, 1.5–1.3 Ga St. Francois and Spavinaw Granite–Rhyolite Belts and 1.3–1.2 Ga Elzevirian Belt in the southwestern North America; the 1.8–1.7 Ga Makkovikian Belt and 1.7–1.6 Ga Labradorian Belt in northeastern North America; the 1.8–1.7 Ga Ketilidian Belt in Greenland; the 1.8–1.7 Ga Malin Belt in the British Isles; and the 1.8–1.7 Ga Transscandinavian Igneous Belt, 1.7–1.6 Ga Kongsberggian–Gothian Belt, and 1.3–1.2 Ga early Sveconorwegian Belt in Baltica (Gower et al., 1990; Condie, 1992; Muir et al., 1992; Gorbatschev and Bogdanova, 1993; Chadwick and Garde, 1996; Åhäll and Gower, 1997; Åhäll and Larson, 2000; Karlstrom et al., 2001). The occurrence of temporally and petrologically similar rocks across a distance of thousands of kilometers between these continents is impressive and appears to support the existence of the Columbia supercontinent. Petrological and geochemical studies indicate that this large magmatic belt includes dominantly juvenile volcanogenic sequences and granitoid suites resembling those of present-day island arcs and active continental margins (Nelson and DePaolo, 1985; Bennet and DePaolo, 1987), representing subduction-related episodic outgrowth along the continental margin of Columbia (Karlstrom et al., 2001; Rogers and Santosh, 2002).

In South America, a 1.8–1.3 Ga accretionary zone occurs along the western margin of the Amazonia Craton. It includes the 1.80–1.45 Ga Rio Negro–Juruena Belt and 1.45–1.30 Ga Rondonian–San Ignacio Belt (Fig. 7; Sadowski and Bettencourt, 1996; Tassinari and Macambira, 1999), both of which consist of supracrustal and igneous rocks that formed during subduction at the western margin of Amazonia. The north–northwest-striking Rio Negro–Juruena Belt includes the Juruena terrane in the northwest and the Rio Negro terrane in the southwest (Fig. 7), which are considered to represent either a partially preserved subduction-related magmatic arc or two successively accreted island arcs (Tassinari and Mac-

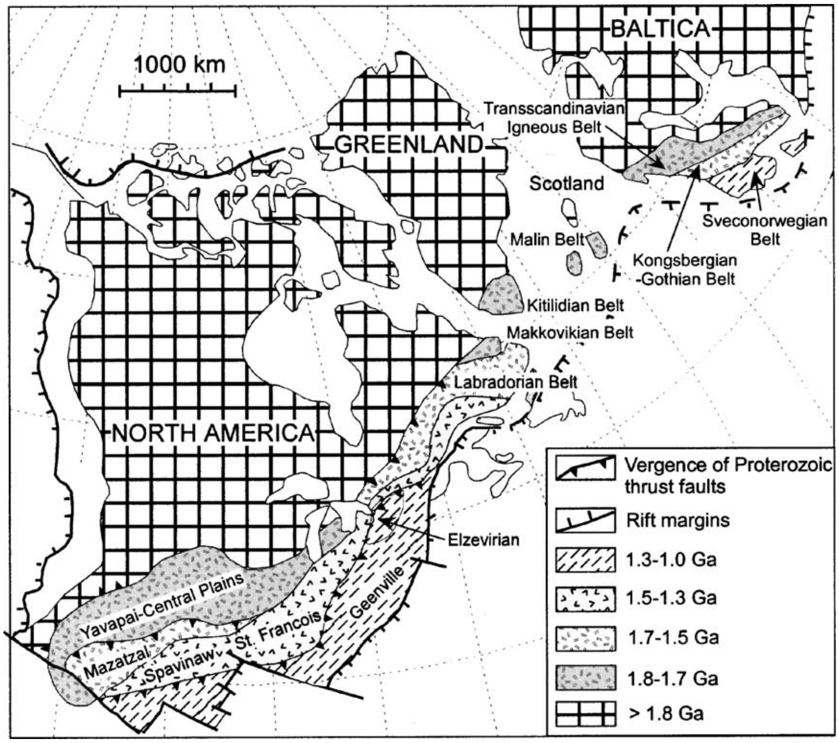

Fig. 6. Spatial distribution of a long-lived (1.8–1.3 Ga) subduction-related magmatic belt along the southern margin of North America, Greenland, Scotland and Baltica (modified after Karlstrom et al., 2001).

ambira, 1999; Geraldes et al., 2001). Geraldes et al. (2001) divided the southwestern Rio Negro–Juruena Belt into three major subduction-related suites: (a) the 1.79–1.74 Ga Alto Jauru suite in the eastern part of the belt, representing a mainly juvenile island arc formation; (b) the ~ 1.55 Ga Cachoeirinha suite in the central part of the belt and consisting of calc-alkaline plutons emplaced into Alto Jauru suite, interpreted as the roots of a continental margin arc built upon basement comprising the Alto Jauru suite; and (c) the 1.52–1.47 Ga Rio Alegre volcanic and mafic plutonic suite intruded by the ~ 1.45 Ga Santa Helena batholith that is interpreted as the magmatic core of a juvenile arc accreted to the edge of the Alto Jauru suite. The 1.5–1.3 Ga Rondonian Belt consists of the extensively reworked crust of the Rio Negro–Juruena Province and juvenile magmatic arc formations, represented by the dominantly potassic calc-alkaline granitic Pensamiento Complex of which the youngest magmatic units are the 1325 ± 45 Ma (Rb/Sr whole rock) Piso Firme Granophyre and the 1283 ± 33 Ma (Rb/Sr whole rock) Orobayaya Granite (Sadowski and Bettencourt, 1996).

In Australia, 1.8–1.5 Ga accretionary magmatic belts are extensively exposed as inliers surrounding the southern and eastern margins of the North Australian Craton and the eastern margin of the Gawler Craton, represented by the Arunta, Mt. Isa, Georgetown, Coen and Broken Hill Inliers (Fig. 8; Myers, 1993; Myers et al., 1996). The development of these inliers was previously interpreted to be associated with ensialic rifting, mafic underplating, volcanism and granitoid magmatism, and a subsequent compressive tectonothermal event term the Barramundi Orogeny at 1880–1850 Ma (Etheridge et al., 1987; Wyborn, 1988). However, recent geological data show that major magmatic events associated with the formation of these inliers coincided with significant tectonic events that occurred along a complex convergent plate boundary that developed on the southern margin of the proto-Australian (North Australia) continent (Giles et al., 2002). The Arunta Inlier

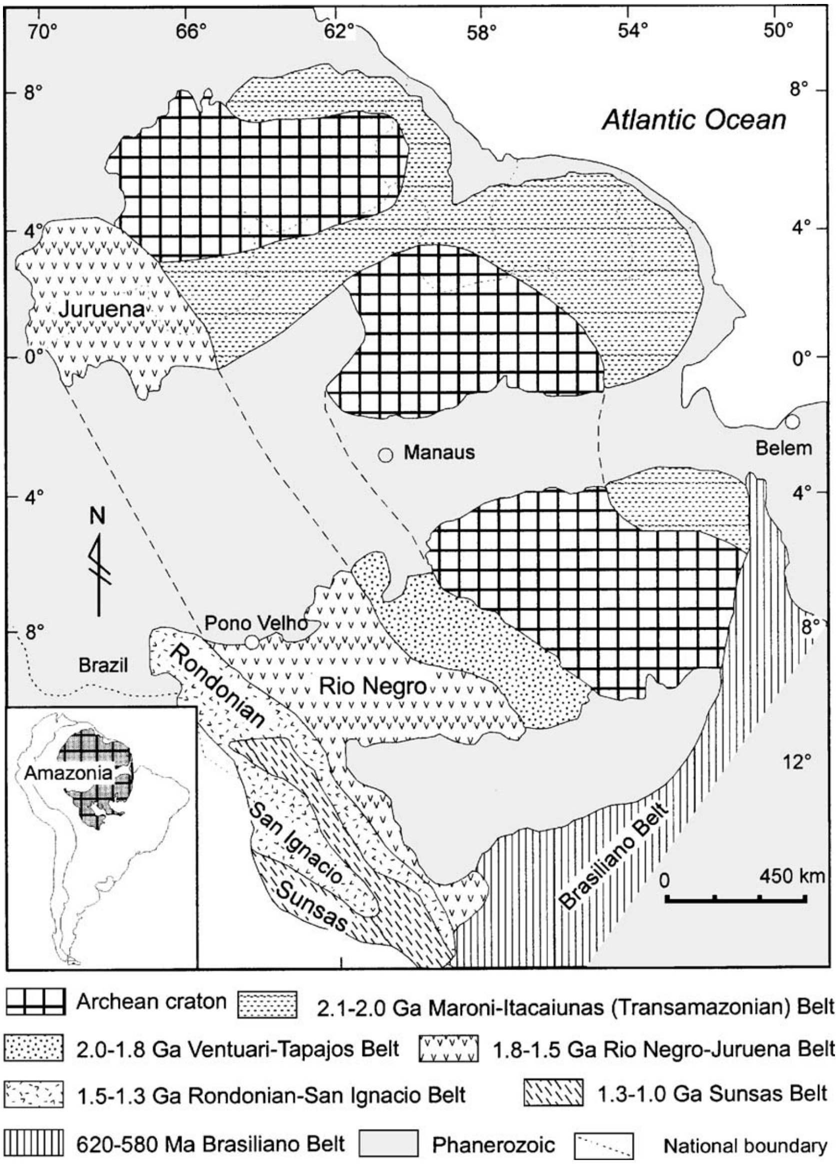

Fig. 7. Spatial distribution of the 1.8–1.5 Ga Rio Negro and Juruena Belts and 1.45–1.3 Ga Rondonian and San Ignacio Belts along the western margin of the Amazonia Craton in South America (after Geraldes et al., 2001).

is the largest one in Central Australia (Fig. 8), covering approximately 200,000 km^2. Conventionally, it has been divided into three major tectonic provinces (Northern, Central and Southern), each of which contains high-grade metamorphosed sedimentary and volcanic rocks that have been assigned to three major divisions, with Division 1 being the oldest and Division 3 the youngest (Steward et al., 1984). Granitoid rocks ranging in age from 1.8 to 1.6 Ga constitute a major component in the Arunta Inlier, represented by five episodes of granitoid magmatism at 1820, 1770–1750, 1730–1710, 1660–1650 and 1615–1590 Ma (Clarke et al., 1990; Collins and Williams, 1995; Zhao and Bennet, 1995; Zhao and McCulloch, 1995). Geochemical and Nd isotopic data indicate that most granitoid rocks in the Arunta Inlier

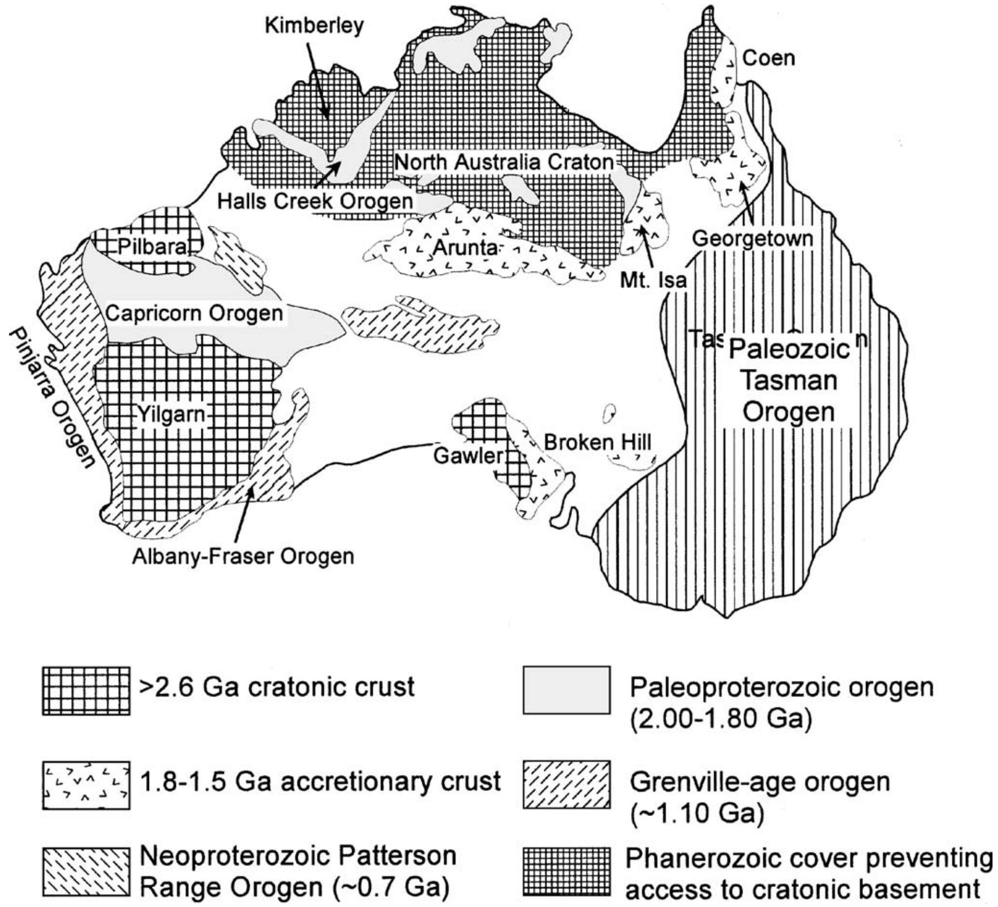

Fig. 8. Schematic map showing the 1.8–1.5 Ga accretionary magmatic belts around the eastern and southern margins of the North Australia Craton and the eastern margin of the Gawler Craton (Modified after Myers et al., 1996; Bodorkos et al., 1999, 2000).

formed by fractionation of arc-type magmas and/or partial melting of arc-related intrusions or underplates in a subduction-related continental margin setting (Zhao and McCulloch, 1995). Recently, Giles et al. (2002) demonstrated that, like the Arunta Inlier, other inliers (the Mount Isa, McArthur, Georgetown, etc) along the eastern margin of the North Australia Craton also formed in a long-lived subduction-related continental margin environment.

In China, a 1.8–1.4 Ga accretionary magmatic zone, called the Xiong'er belt (Group), extends along the southern margin of the North China Craton, covering a total area of more than 25 000 km^2 (Fig. 9). The belt is separated from the pre-1.8 Ga basement of the North China Craton in the north by the Lintong–Tongguan–Sanmenxia–Yichuan–Baofeng–Luohe Fault and from the 1.6–1.4 Ga Guanping Ophiolite Complex in the south by the Luonan–Luanchun–Fangcheng Fault (Fig. 9). The major lithologies in the belt are andesitic lava, dacitic porphyry and minor intermediate to acid volcanic tuff, which are lithologically similar to rock associations of an Andean-type continental margin arc (Chen, 1992; Chen and Zhao, 1997; Zhao et al., 2003b). Various tectonic–magmatic discrimination diagrams also show that the volcanic rocks of the Xiong'er belt have affinities to magmatic arcs. Spatially, the Xiong'er volcanic rocks show an increase in K_2O, K_2O/Na_2O, K_2O/SiO_2, K_2O+Na_2O and K_2O+Na_2O/Al_2O_3 from south to north (Cheng and Fu, 1992), implying subduction to the north, which is consistent with the spatial distribution of the Xiong'er belt and the

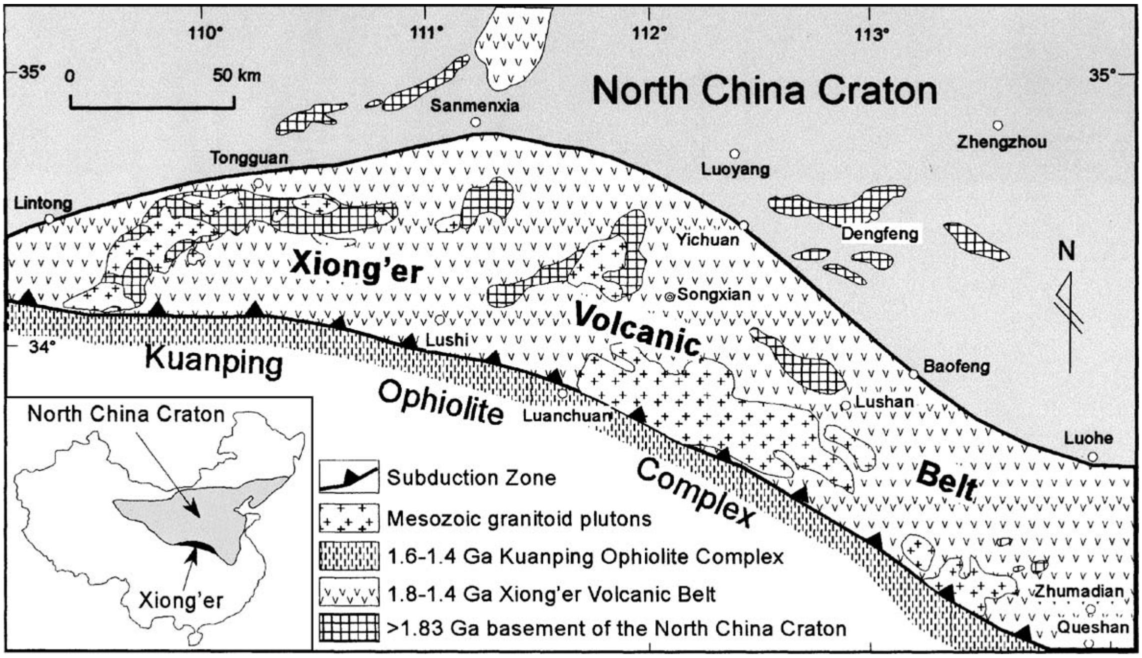

Fig. 9. Geological map showing the Xiong'er volcanic belt, a 1.8–1.4 Ga accretionary complex along the southern margin of the North China Craton, modified after Chen and Fu (1992).

Kuanping ophiolite complex (Fig. 9). The Luonan–Luanchun–Fangcheng Fault separating the Xiong'er belt from the Kuanping ophiolite complex has been interpreted as an important tectonic boundary along which the Qinling oceanic crust, represented by the Kuanping ophiolite complex, was subducted beneath the southern continental margin of the North China Craton, leading to the formation of the Xiong'er continental margin arc (Chen and Fu, 1992; Chen and Zhao, 1997). This implies that in the supercontinent Columbia, the southern margin of the North China Craton may not have been connected with any other continents, but was bordered by an active subduction zone.

6. Fragmentation and final breakup of the supercontinent Columbia

One of the most characteristic features of Mesoproterozoic geology is widespread continental rifting and anorogenic magmatism (Anderson, 1983; Windley, 1979, 1989, 1993, 1995; Green, 1992; Tarney, 1992; Wiebe, 1992; Anderson and Morrison, 1992), presumably due to mantle plume diapirism and asthenospheric upwelling. Windley (1995) proposed that Mesoproterozoic anorogenic magmatism developed as a result of the breakup of a supercontinent that formed before 1.65 Ga. Condie (2002) reconstructed the possible configuration of continental fragments following the initial fragmentation of a Paleo-Mesoproterozoic supercontinent at about 1.5 Ga (Fig. 10), based on the distribution of Mesoproterozoic rift belts, anorogenic magmatic assemblages and mafic dike swarms. Rogers and Santosh (2002) considered that the fragmentation of the supercontinent Columbia began about 1.6 Ga ago, associated with continental rifting and anorogenic magmatism, and continued until its final breakup at about 1.2 Ga, marked by widespread emplacement of mafic dike swarms.

6.1. Mesoproterozoic continental rifting

Continental rifting in the Mesoproterozoic (1.6–1.2 Ga) was recorded along both the margins and interior of Columbia (Rogers and Santosh, 2002), forming a large number of ensialic and continental-margin rift sequences, representatives of which are the

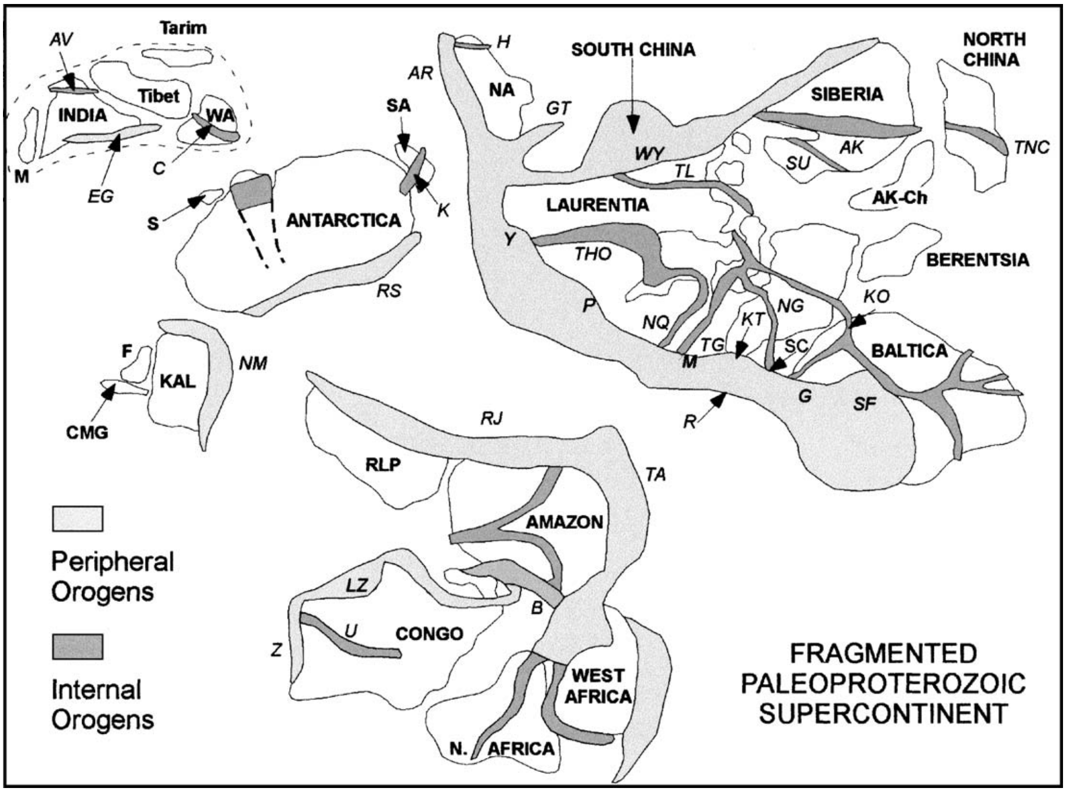

Fig. 10. Possible configuration of continents following the breakup of a Paleo-Mesoproterozoic supercontinent at about 1.5 Ga (Condie, 2002). Map locations: M—Madagascar; WA, NA, SA—West, North and South Australia, respectively. S—Sri Lanka; F—Falkland–Malvinas plateau; CMG—Coats Land, Maudheim, Gunehogna in Antarctica; KAL—Kalahari craton; RLP—Rio de la Plata craton; R—Rockall; SC—Scotland. Orogens: *AK*—Akitkan (1.9–1.8 Ga); *AR*—Arunta (1.85–1.7 Ga); *AV*—Aravalli (1.5–1.4 Ga); *B*—Birimian (2.2–2.0 Ga); *C*—Capricorn (1.95–1.80 Ga); *EG*—Eastern Ghats (1.5 Ga); GT—Georgetown (1.6–1.5 Ga); G—Gothian (1.7–1.6 Ga); *H*—Hooper (1.85–1.8 Ga); *K*—Kimban (1.85–1.70 Ga); *KT*—Ketilidian (1.9–1.8 Ga); *KO*—Kola (2.0–1.9 Ga); *L*–Labradorian (1.7–1.6 Ga); *LZ*—Luizian (2.2–2.1 Ga); *M*—Makkovikian (1.9–1.8 Ga); *NG*—Nagssugtoqidian (1.9–1.8 Ga); *NM*—Namaqua (2.0–1.8 Ga); *NQ*—New Quebec (1.9—1.8 Ga); *P*—Penokean (1.9–1.8 Ga); *R*—Rayner (1.9–1.5 Ga); *RJ*—Rondonian–Juruena (1.8–1.55 Ga); *RS*—Ross (1.8–1.7 Ga); *SU*—Sutam (1.9 Ga); *SF*—Svecofennian (1.9–1.8 Ga); *TL*—Taltson (1.9–1.8 Ga); *TG*—Torngat (1.9–1.8 Ga); *TA*—Transamazonian (2.2–2.0 Ga); *THO*—Trans-Hudson (1.9–1.8 Ga); *TNC*—Trans-North China (1.9–1.8 Ga); *U*—Ubendian (2.0–1.9 Ga); *WY*—Wopmay–Angara (2.0–1.5 Ga); Y—Yavapai (1.8–1.7 Ga); Z—Zambezi (1.5–1.4 Ga).

Belt–Purcell Supergroup in North America, Telemark Supergroup in Baltica, Riphean Aulacogens in Siberia, Kalahari Copper Belt in Southern Africa, Kibaran Belt in eastern and central Africa, and Zhaertai–Bayan Obo Belt in North China.

The Mesoproterozoic Belt–Purcell Supergroup of the northwestern US and adjacent Canada is exposed over 130,000 km^2, resting on a NW-trending trough bounded by a network of extensional syndepositional faults in Montana, Idaho, northeastern Washington, and British Columbia (Fig. 11; Green, 1992). The supergroup is composed of thick successions of deep-water turbidites succeeded by thick cycles of shallow-water clastic and carbonate sediments, which are interleaved with bimodal volcanics, including continental tholeiitic basalts and minor rhyolites (Green, 1992). The strata of the Belt–Purcell Supergroup have long been interpreted as continental rift deposits, with the likely inclusion of intracontinental lacustrine sediments (Harrison et al., 1974; Winston, 1986, 1990; Winston and Link, 1993; Blewett et al., 1998), although the early Belt–Purcell basin may have received episodic fluxes of marine water (Luepke and Lyons, 2001). Geochronological data, in combination

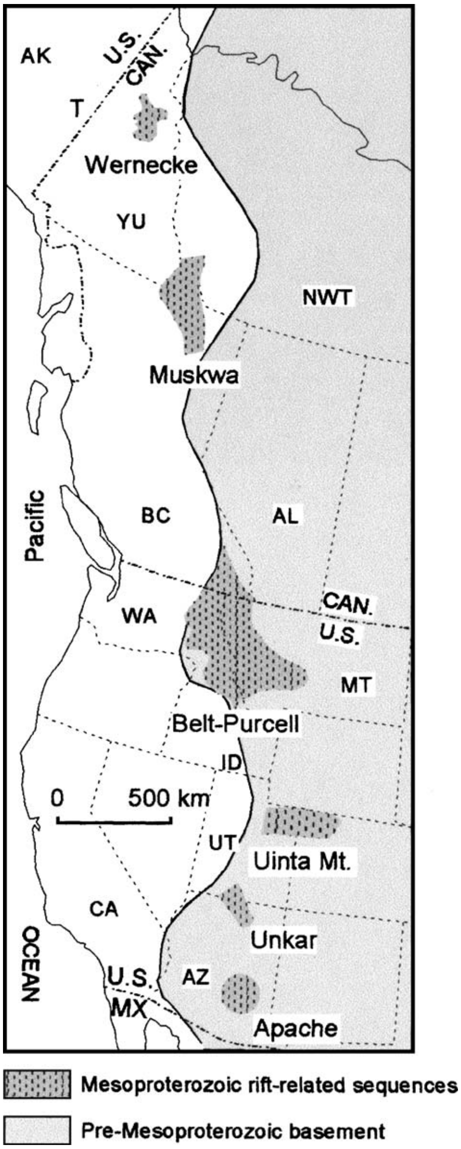

Fig. 11. Sketch map showing localities of Mesoproterozoic rift-related sequences along the western margin of North America. After Green (1992), Rogers and Santosh (2002) and other data sources. Abbreviations: T—Tindir Group; O—Ogilvie Mountains; AK—Alaska; AL—Alberta; AZ—Arizona; BC—British Columbia; CA—California; ID—Idaho; MT—Montana; NWT—Northwest Territories; UT—Utah; WA—Washington.

with paleomagnetic results, constrain the lower and upper sequences of the Belt–Purcell Supergroup to the periods 1470–1440 and 1370–1300 Ma, respectively (Kidder, 1992, 1998; Winston and Link, 1993; Anderson and Davis, 1995; Luepke and Lyons, 2001), as indicated by lower Belt–Purcell synsedimentary mafic intrusions and upper Belt mafic and felsic extrusive rocks (Sears and Price, 2002). Delaney (1981) and Green (1992) correlate the lower sequences of the Belt–Purcell Supergroup with the Muskwa and Wernecke Supergroups in northwestern Canada, and Rogers and Santosh (2002) suggest that the Uinta, Unkar and Apache Groups in Utah and Arizona represent the southern extension of the Belt–Purcell Supergroup; they together constitute a large Mesoproterozoic continental rift system along the western margin of Laurentia, with most of its length being covered by younger deposits (Fig. 11). Provenance studies of the Belt–Purcell sediments corroborate a western sediment source (Harrison et al., 1974; Winston, 1986; Frost and Winston, 1987; Ross et al., 1992; Blewett et al., 1998), and the development of this long-lived (1.5–1.3 Ga) rift system is considered to have been associated with the breaking away of the western Laurentia continental margin from its Mesoproterozoic conjugate, whose candidates include Siberia (Sears and Price, 1978, 2000, 2002), Australia (Ross et al., 1992; Borg and DePalo, 1994; Blewett et al., 1998), South China (Li et al., 1995, 1996) or eastern India (Rogers and Santosh, 2002).

In Baltica, the Mesoproterozoic Telemark Supergroup is considered to have formed in a long-lived, rift-related, basin that developed along a N–S trending trough on older Proterozoic crust in southern Norway (Fig. 12; Green, 1992; Brewer and Menuge, 1998; Bingen et al., 2001). The supergroup can be divided into the Rjukan, Seljord and Bandak Groups (Dons, 1960). The Rjukan Group consists predominantly of rhyolites overlain by basic volcanic rocks, whose geochemical signature points to a within-plate setting of continental rifting (Bingen et al., 2001). Volcanic activity in the Rjukan Group is estimated at 1.51 Ga by zircon U–Pb data on two rhyolite horizons (Dahlgren et al., 1990; Sigmond et al., 1997), and is nearly coeval with inter-orogenic magmatism of 1.51–1.50 Ga gabbro–dolerite–granite associations and ~ 1.46 Ga N–S trending mafic dyke swarms in Baltica (Åhäll and Connelly, 1996, 1998). The Seljord Group overlies the Rjukan Group and is dominated by quartzite. The Bandak Group unconformably overlies both the Rjukan and Seljord Groups, and contains acid and basic volcanic rocks overlain by immature

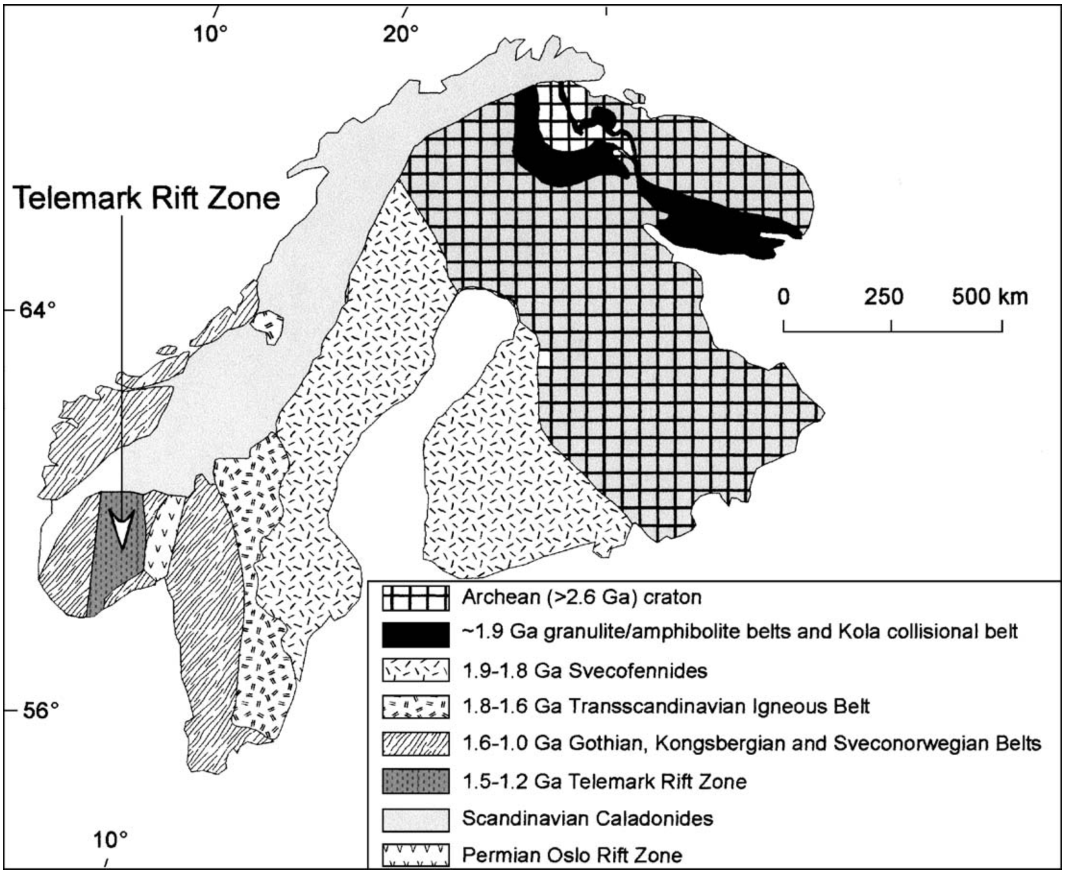

Fig. 12. Schematic geological map of the Baltica Shield, showing the location of the Mesoproterozoic Telemark rift zone (after Bingen et al., 2001).

sediments, which are interpreted as a rift-basin fill. A rhyolite from the Bandak Group gives a zircon U–Pb crystallization age of ca. 1.16 Ga, interpreted to be the youngest age of the Telemark Supergroup. Green (1992) suggests that the extensional event leading to the development of the Telemark rift basin may have been related to the clockwise rotation and separation of Baltica with respect to Laurentia in the late Mesoproterozoic, which is consistent with paleomagnetic data that indicate a period of independent movement between 1200 and 950 Ma, during which Baltica appears to have rotated $\sim 80^\circ$ clockwise relative to Laurentia (Park, 1995).

In Siberia, several Mesoproterozoic rift-related basins (aulacogens) lie between the Oleniok uplift and the Anabar block (Zonenshain et al., 1990; Green, 1992), and also along the southeastern margin of the Siberia Craton (Khudoley et al., 2001). The sedimentary–volcanic formations in these aulacogens are generally referred to as the Riphean succession. The Riphean rifting between the Oleniok uplift and the Anabar block is exemplified by the Udzh aulacogen, which involved considerable thinning of the continental crust and deposition of 7 to 9 km of sediments that were associated with widespread alkaline-ultramafic magmatism, including kimberlite (Zonenshain et al., 1990). The Riphean succession along the southeastern margin of the Siberia Craton was traditionally divided into the lower Riphean (1650–1350 Ma) Uchur Group, middle Riphean (1350–1000 Ma) Aimchan and Kerpyl Groups, and upper Riphean (1000–650 Ma) Lakhanda and Uy Groups (Surkov et al., 1991). However, subsequent dating of mafic sills that intrude the Uy Group yields a Sm–Nd isochron age of

942 ± 18 Ma (Pavlov et al., 1992) and U–Pb baddeleyite ages of 974 ± 7 and 1005 ± 4 Ma (Rainbird et al., 1998), showing that the Uy Group is possibly older than 1000 Ma. Therefore, the whole Riphean succession of southeastern Siberia may have formed during the Mesoproterozoic. The Riphean succession comprises predominantly volcanoclastic-terrigenous and volcanoclastic-carbonate rocks in terrestrial to shallow marine sedimentary environments, which were associated with emplacement of alkaline-ultramafic plutons (Surkov et al., 1991; Khudoley et al., 2001). Sedimentological and stratigraphic studies indicate that the Riphean sedimentary basin of southeastern Siberia was initiated by rifting that subsequently failed, allowing the development of a long-lived intracratonic sedimentary basin (Khudoley et al., 2001). This period of cratonic rifting may have been related to the breaking away of Siberia, either from western Laurentia with the development of the Belt–Purcell–Wernecke Basins, as speculated by Sears and Price (1978, 2000, 2002), or from northern Laurentia at about ∼ 1.5 Ga, as proposed by Condie and Rosen (1994), Frost et al. (1998) and others (e.g. Rainbird et al., 1998).

The late Mesoproterozoic Kalahari Copper Belt is a northeasterly trending rift system that developed along what are now the western and northern margins of the Kalahari Craton of southern Africa (Fig. 13; Borg, 1988; Green, 1992). Defined by a series of block-faulted basins on the Paleo-Mesoproterozoic basement, the belt extends 1800 km from central Nambia through northern Botswana to northern Zimbabwe, and contains a thick basal accumulation of high-K rhyolites, pyroclastic rocks and minor coarse clastic sediments, succeeded by a rift-fill sequence of up to 3 km of continental redbeds and interbedded basalts, which are overlain by shallow marine and possibly lacustrine sediments with native Cu mineral-

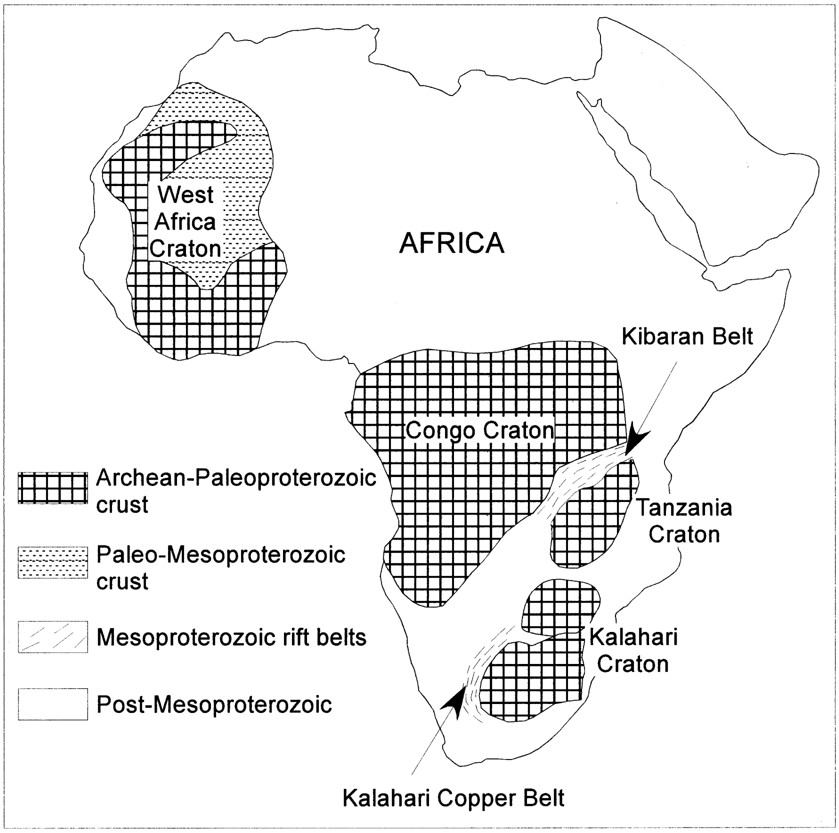

Fig. 13. Sketch showing the Mesoproterozoic Kalahari Copper and Kibaran belts in Africa. Modified after Kröner (1980) and Green (1992).

ization. The basalts are tholeiites having within-plate geochemistry, indicating a continental rift setting (Borg, 1988). The timing of the onset of this rifting subsidence is not well constrained, being estimated at about ~ 1300 Ma (Green, 1992). It still remains unclear whether rifting of the Kalahari Copper Belt led to separation of an unknown continental block from the Kalahari Craton.

The Mesoproterozoic Kibaran belt is considered to be a rift system that separates the Congo Craton on the west from the Tanzania Craton on the east (Fig. 13; Green, 1992; Tack et al., 1994; Fernandez-Alonso and Theunissen, 1998). Evidence for rifting includes a thick sequence (11–14 km) of clastic sediments, named the Burundi Supergroup, and bimodal volcanics. The Burundi Supergroup is composed mainly of quartzitic and pelitic rocks, and poorly sorted and mature conglomerates, which are interpreted as products of rifting subsidence (Green, 1992). The bimodal volcanics are abundant tholeiites and minor rhyolites that mainly occur in the upper sequence of the supergroup. Associated with these volcanics are granitoid plutons and mafic and ultramafic intrusions, which can be divided into two distinct magmatic suites, both including A-type granites (Tack et al., 1994). The first is the 350-km-long Kabanga–Musongati suite with an emplacement age of 1275 ± 11 Ma (U–Pb zircon age) and is mainly composed of mafic and ultramafic layered rocks with subordinate A-type acidic differentiates moderately enriched in incompatible elements (Tack et al., 1994). The second is the 40-km-long Gitega–Makebuko–Bukirasazi suite with an emplacement age of 1249 ± 8 Ma (U–Pb zircon age) and is composed of A-type granitoids rocks (Tack et al., 1994). Geochemical data indicate an old continental lithospheric mantle origin for both the magmatic suites (Tack et al., 1994); probably related to mantle plume diapirism and asthenospheric upwelling, which led to fragmentation of the Congo and Tanzania Cratons in the late Mesoproterozic (Klerkx et al., 1987; Key and Ayres, 2000; Rogers and Santosh, 2002).

The Mesoproterozoic Zhaertai–Bayan Obo Rift Zone (1.6–1.2 Ga) occurs along the northern margin of the Northern China Craton, extending from Zhaertai in the west, through Bayan Obo and Huade, to Weichang in the east, and is up to 1500 km long (Fig. 14; Zhao et al., 2003b). The best-studied segment of the rift zone is the Bayan Obo Belt, which is internationally known for hosting the largest REE deposit in the world. The east–west-trending Bayan Obo Belt is separated from the late Archean to Paleoproterozoic basement of the North China Craton in the south by the Damaoqi–Shangdu Fault and overthrust by the Caledonian Baoerhantu–Wenduermiao accretionary complex and the Hercynian Suolunshan–Xilabulunhe accretionary complex in the north along the Wulanbaolige and Xilabulunhe Faults, respectively (Zhou et

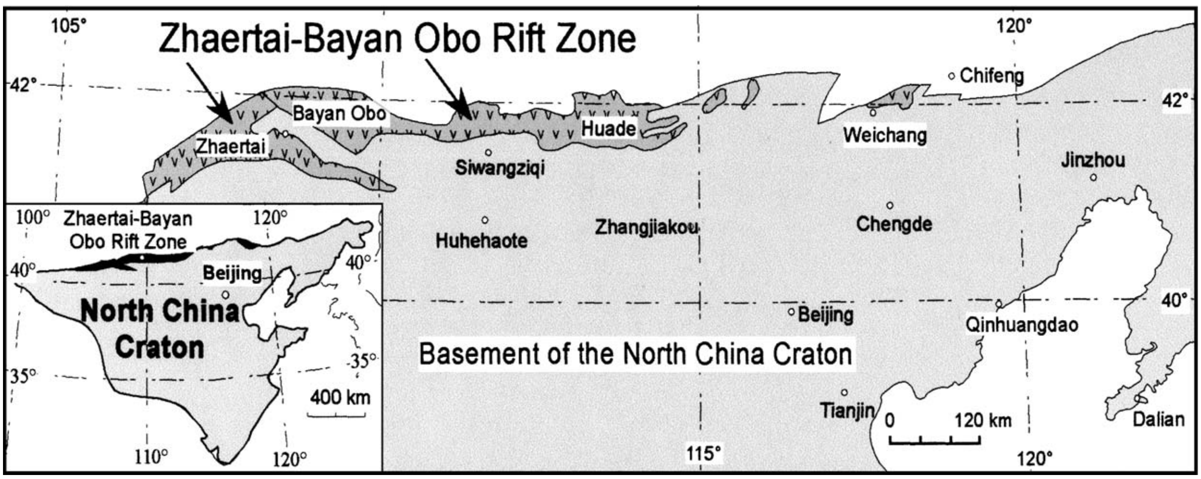

Fig. 14. Sketch showing the Mesoproterozoic Zhaertai–Bayan Obo rift zone along the northern margin of the North China Craton (After Zhao et al., 2003b).

al., 2002). The rift belt comprises sedimentary and volcanic assemblages metamorphosed to sub-greenschist facies. The sedimentary assemblages, conventionally termed the "Bayan Obo Group", comprise conglomerates, pebble-bearing sandstones, turbiditic graywackes, shales, limestones and iron formations, indicative of a transition from a rapidly subsiding ensialic basin or continental marginal environment to a stable continental shelf environment (Zhou et al., 2002). The volcanic assemblages are typically bimodal in composition, with tholeiitic to alkalic basalts dominant and subordinate K-enriched rhyolites, which are associated with alkaline intrusions including alkaline syenite, aegirinite and carbonatites. Field relationships show that magmatism in this rift belt was intense at the early stage and waned as sedimentation increased, though it continued throughout the whole time of rifting (Zhou et al., 2002). The lithological assemblages of the Zhaertai, Huade and Weichang Belts can be well correlated with those of the Bayan Obo Belt, indicating they belonged to the same rift zone. The development of this rift is considered to have been associated with the separation of the North China Craton from some other cratonic block during the fragmentation of the supercontinent Columbia.

6.2. Mesoproterozoic anorogenic magmatism

Mesoproterozoic anorogenic magmatism has been recognized in most cratonic blocks in the world, especially in North America, Greenland and Baltica, where anorogenic magmatic assemblages occur along a huge belt that extends from Fennoscandia, through southern Greenland and northeastern Canada, to the southwestern US (Fig. 15; Gower et al., 1990; Anderson and Morrison, 1992; Green, 1992; Wiebe, 1992; Windley, 1989, 1993, 1995; Åhäll and Connelly, 1998; Åhäll et al., 2000; Barnes et al., 2002). The development of this extensive anorogenic magmatic belt is considered to be associated with the fragmentation of a Paleo-Mesoproterozoic supercontinent (Windley, 1995; Condie, 2002; Rogers and Santosh, 2002).

In Baltica, the anorogenic magmatism is represented by the Fennoscandian rapakivi granites and minor rhyolites, which formed during separate magmatic episodes between 1.60 and 1.30 Ga (Fig. 15; Åhäll et al., 2000; Karlstrom et al., 2001). In North American, the only known orogenic event between 1.6 and 1.1 Ga was the 1.21- to 1.33 Ga Elzevirian orogeny of the southern Adirondack Mountains (Daly and McLelland, 1991), and throughout much of the continent, the period from 1.6–1.2 Ga is characterized by a wide spectrum of anorogenic activity, including emplacement of anorthosite massifs, charnockite intrusions, diabase dyke swarms, batholiths of potassic rapakivi granite, and eruption of coeval rhyolites (Anderson, 1983; Windley, 1989, 1993, 1995; Gower et al., 1990; Anderson and Morrison, 1992; Green, 1992; Wiebe, 1992; Karlstrom et al., 2001; Barnes et al., 2002). These anorogenic rocks occur along a 5000-km-long and 1000-km-wide belt that extends from southern California to Labrador, marginal to and within the Grenville Orogenic Belt (Fig. 15; Windley, 1995; Karlstrom et al., 2001). The belt can be broadly subdivided into southern and northern zones (Windley, 1995).

The southern zone is in the mid-continental USA where it is known as the granite–rhyolite province and consists predominantly of 1.5–1.3 Ga rhyolitic ash-flow tuffs and epizonal granitic plutons, with a noticeable lack of anorthosites, charnockites and gabbros (Anderson and Morrison, 1992; Windley, 1995; Barnes et al., 2002). Within the zone, the principal granites are potassic, iron-enriched biotite and hornblende granites and two-mica granites. The earliest is the ~1485 Ma Wolf River granites in Wisconsin (Anderson, 1980), and the youngest was emplaced at ~1296 Ma in Labrador (Schärer et al., 1986). Over fourteen 1.44- to 1.46 Ga anorogenic granite and rhyolite localities have been identified from Kentucky through southern Illinois, Missouri, and Kansas to southern Nebraska, and more than 40 plutons with an age range from 1.40 to 1.46 Ga occur in southern Wyoming, Colorado, New Mexico, Arizona, southern Nevada and southern California (Anderson and Morrison, 1992). The general character of these granites is "anorogenic" or A-type rapakivi-textured granites (Anderson and Morrison, 1992; Frost et al., 1999). Anderson (1983) suggested that the rhyolites and granites could have formed by 30% fusion of calc-alkaline crust, which is coincident with the fact that these rhyolites and granites in the USA are situated exactly within the area of juvenile 1.8–1.6 Ga calc-alkaline arc-accretionary orogens (Wind-

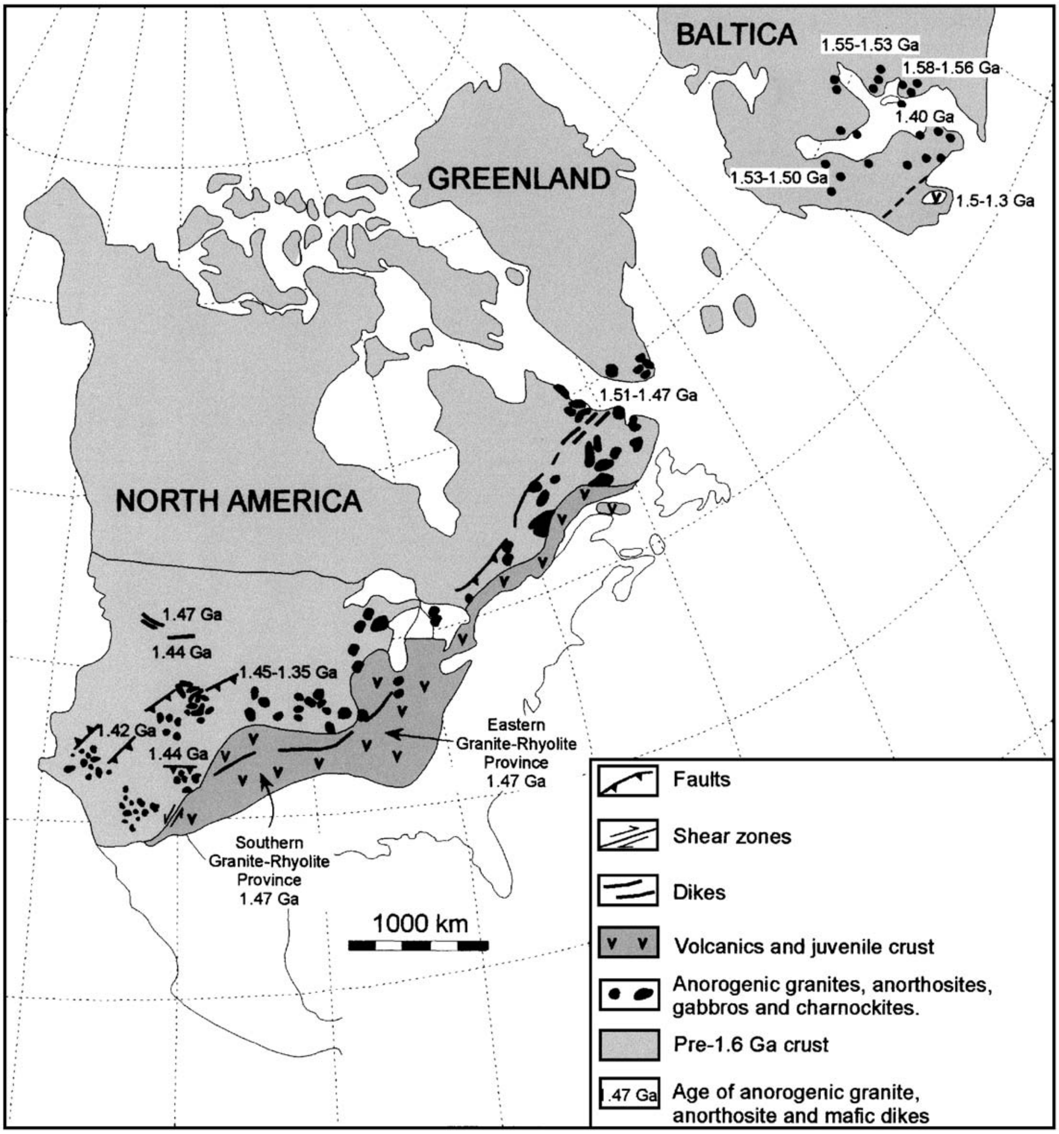

Fig. 15. Distribution of Mesoproterozoic anorogenic magmatism in North America, Greenland and Baltica. Modified after Karlstrom et al. (2001), Anderson and Morrison (1992) and Windley (1993, 1995).

ley, 1995). Windley (1995) pointed out that the most appropriate modern analogue for the rhyolitic type of Mesoproterozoic anorogenic magmatism in the mid-continental USA is the chemically similar, Jurassic Tobifera rhyolite field in Argentina in South America, in which basic and ultrabasic rocks are also noticeably absent (Kay et al., 1989). The Tobifera rhyolites occur within a Paleozoic arc-accretionary orogen (Dalziel et al., 1976; Winslow, 1976; Ramos et al., 1986) as a 1-km-thick sheet adjacent to the Gondwana continental margin (Windley, 1989). According to Windley (1995), both the Phanerozoic and Mesoproterozoic rhyolitic suites formed after the amalgamation of a supercontinent and were associated with its breakup and the formation of a new ocean; these were the Atlantic and Grenvillian oceans, respectively.

The northern zone comprises massif-type anorthosites and associated charnockites, gabbros and granitic intrusives that are widespread within and adjacent to the Grenville orogen in northeast Canada. Representative anorogenic intrusions in this zone include the 1.46–1.44 Ga Michikamau and Harp Lake anorthosites, 1.43 Ga Michael gabbro, 1.41 Ga Flowers River gabbro-anorthosite, 1.38 Ga Shabogamo gabbro, 1.35 Ga Parry Sound anorthosite and Whitestone gabbro-anorthsite, 1.34–1.32 Ga Red Wine and Arc Lake peralkaline plutons, 1.32–1.29 Ga Nain anorthosite, 1.3 Ga Kiglapait gabbro, and 1.27–1.26 Ga peralkaline granites, rhyolitic tuffs and anorthosites in the Flowers River region (Ashwal and Wooden, 1985; Emslie, 1985; Hill and Miller, 1990; Martignole et al., 1993). These intrusions have an age range close to the predominant age of the rhyolites and granites in the southern zone of the mid-continental USA, and Windley (1995) concluded that these rocks were likewise emplaced during anorogenic processes that led to the fragmentation of a Mesoproterozoic supercontinent.

In South America, Mesoproterozoic rapakivi granites and associated mafic and ultramafic rocks have been found in almost all cratonic provinces (Dall' Agnol et al., 1999). In the Rondonia Tin Province of the Amazonian Craton, rapakivi granites were emplaced during discrete episodes of magmatism between ca. 1600 and 1000 Ma, forming the 1606–1532 Ma Serra da Providencia Intrusive Suite, 1406 Ma Santo Antonio Intrusive Suite, 1387 Ma Teotonio Intrusive Suite, 1346–1338 Ma Alto Candeias Intrusive Suite, 1314–1309 Ma Sao Lourenco–Caripunas Intrusive Suite, 1082–1074 Ma Santa Clara Intrusive Suite, and 998–974 Ma Younger Granites of Rondonia (Bettencourt et al., 1999). Of these rapakivi granite suites, the former five pre-Grenvillian suites can be geochronologically correlated with the Fennoscandian rapakivi granites and similar rocks in North America and South Greenland; representing extensional anorogenic magmatism associated with the fragmentation of a pre-Rodinia supercontinent. The two youngest rapakivi suites, the Santa Clara Intrusive Suite and Younger Granites of Rondonia, seemingly represent inboard magmatism during the waning stages of the 1.1–1.0 Ga Sunsas/Grenville collisional orogeny (Bettencourt et al., 1999).

Anorogenic rapakivi granites and associated mafic and ultramafic rocks with similar ages are also reported from the Ural Mountains and the Ukraine (Aberg, 1988), Peninsular India (Leelanandam and Reddy, 1988; Leelanandam, 1990), Africa (Conradie and Schoch, 1986), and North China (Yu et al., 1994). They are considered to record a global-scale extensional event that led to the fragmentation of the Paleo-Mesoproterozoic supercontinent (Windley, 1995).

Global-scale Mesoproterozoic anorogenic magmatism is also recorded by the presence of widespread alkaline ultrabasic rocks, represented by kimberlites, lamproites and carbonatites in West Africa, South Africa, India, South America and Western Australia (Dawson, 1989). They include the 1.6–1.2 Ga Premier kimberlites and carbonatites on the margin of the Kaapvaal Craton (Phillips et al., 1989), 1.4–1.2 Ga Mali kimberlites on the west margin of the Congo Craton and on the southwest margin of West African Craton (Haggerty, 1982), ~ 1.2 Ga kimberlites in the Kimberley area of Western Australia (Pidgeon et al., 1989), and ~ 1.2 Ga Majhgawan kimberlites and lamproites in the Indian Shield (Paul et al., 1975). These plate-scale kimberlite activities represent important mantle upwelling (plume) episodes that resulted in the dispersion of these cratonic blocks from the supercontinent Columbia (Zhao et al., 2002a).

At about 1.35–1.21 Ga, the intrusion of many mafic dike swarms and associated basaltic extrusions in North America (Le Cheminant and Heaman, 1989, Ernst et al., 1995, 2001; Ernst and Buchan, 2003), Greenland (Cadman et al., 2001), Baltica (Park, 1992), North China (Ma et al., 1987), South America (Raposo and D'Agrella, 2000), East Antarctica (Sheraton et al., 1990), Western Australia (Qui et al., 1999; Pidgeon and Nemchin, 2001) and Central-South Australia (Mortimer et al., 1988), marks a major episode of plate-wide extension and rifting (Windley, 1995). One of the most widely distributed dyke swarms of this period is the MacKenzie swarm, which extends over an area of 2.7 million km^2 and dramatically fans over an arc angle of about 100° from a point in northwest Canada (Fig. 16; Ernst et al., 1995). Near the focus of the swarm are thick accumulations of coeval basalts (the Coppermine River basalts) and large layered ultramafic–mafic intrusions (the Muskox intrusions). Precise U–Pb baddeleyite dating has shown that the Mackenzie dyke swarm and the Muskox intrusion were emplaced within a short period of less than 5 million years beginning at 1272 Ma ago, with all the dykes

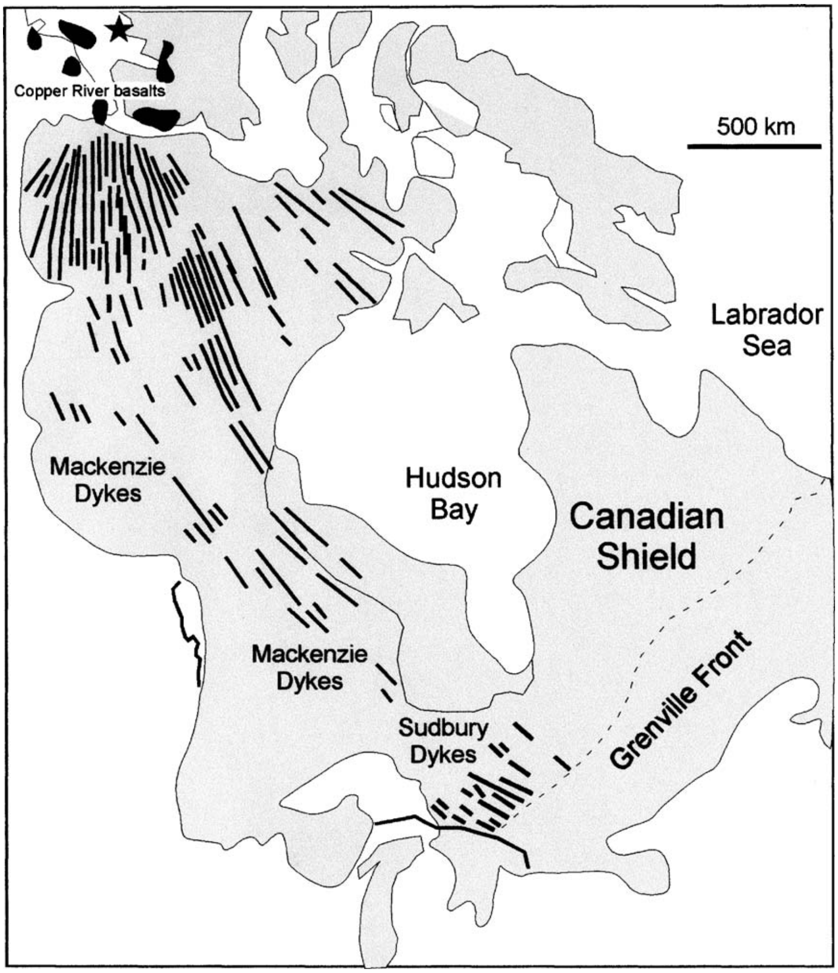

Fig. 16. The 1.27 Ga Mackenzie giant radiating dike swarm and coeval Copper River basalts, and 1.24 Ga giant linear Sudbury swarm of North America (Ernst et al., 1995). Star marks Mackenzie mantle-plume center defined by convergence of the dikes.

being injected at 1267 ± 2 Ma (U–Pb baddeleyite age; Le Cheminant and Heaman, 1989). Based on the radiating pattern, the stratigraphic evidence for uplift in the local region preceding magma injection, and magnetic fabric data indicating vertical flow in the focal region and lateral flow at all distances beyond, Ernst et al. (1995) interpreted the swarm as having been initiated by a mantle plume impinging into the lithosphere. The associated Coppermine River basalts have trace element and Sr–Nd–Pb isotopic characteristics comparable to those of Phanerozoic flood basalts, and this provides further evidence for a mantle plume and hotspots at the northern apex of the radiating Mackenzie swarm connected with the opening of an ocean (Dupuy et al., 1992). The ~1.24 Ga Sudbury dyke swarm (Fig. 16) in southern Canada is also thought to be related to the opening of an ocean (Windley, 1995). Other similar-aged dike swarms in North America include the 1.32 Ga Seal Lake dike swarm, 1.32 Ga Harp dike swarm, and 1.21 Ga Mealy dike swarm. These mafic dyke swarms together constitute a plate-wide extensional episode that may have marked the youngest piercing points at which the continental blocks in the Paleo-Mesoproterozoic supercontinent Columbia can be geologically linked and signaled the final breakup of the supercontinent (Rogers and Santosh, 2002; Zhao et al., 2002a).

7. Summary and discussion

In Earth's history, supercontinents containing most of the earth's continental crust are known to have formed on at least two separate occasions; the younger supercontinent Pangea formed at 300–250 Ma by accretion and amalgamation of fragments produced by breakup of the older one, Rodinia, which was assembled at ∼ 1000 Ma along Grenville-age sutures (Powell et al., 1993; Dalziel, 1995, Dalziel et al., 1997, 2000; Meert and Powell, 2001; Torsvik, 2003). It now seems likely that Rodinia formed by accretion and assembly of fragments from an even older supercontinent (Rogers and Santosh, 2002; Condie, 2002; Meert, 2002; Sears and Price, 2002; Wilde et al., 2002; Zhao et al., 2002a, 2003b). This pre-Rodinia supercontinent contained almost all of the world's continental blocks that were amalgamated along global 2.1–1.8 Ga collisional orogens (Fig. 17), including the 2.1–2.0 Ga Transamazonian and Eburnean Orogens in South America and West Africa, respectively; 1.95–1.85 Ga Trans-Hudson Orogen or its equivalents (Taltson–Thelon, Wopmay, New Quebec, Foxe, Makkovik, Ungava and Torngat Orogens) in North America; 1.9–1.8 Ga Nagssugtoqidian Orogen in Greenland; 1.9–1.8 Ga Kola–Karelia, Volhyn–Central Russian and Pachelma Orogens in Baltica (including East Europe); 1.9–1.8 Ga Akitkan Orogen in Siberia; 2.0–1.9 Ga Limpopo Belt in South Africa; 2.0–1.9 Ga Capricorn Belt in Western Australia; ∼ 1.8 Ga Trans-North China Orogen in North China; and ∼ 1.8 Ga Central Indian Tectonic Zone in India.

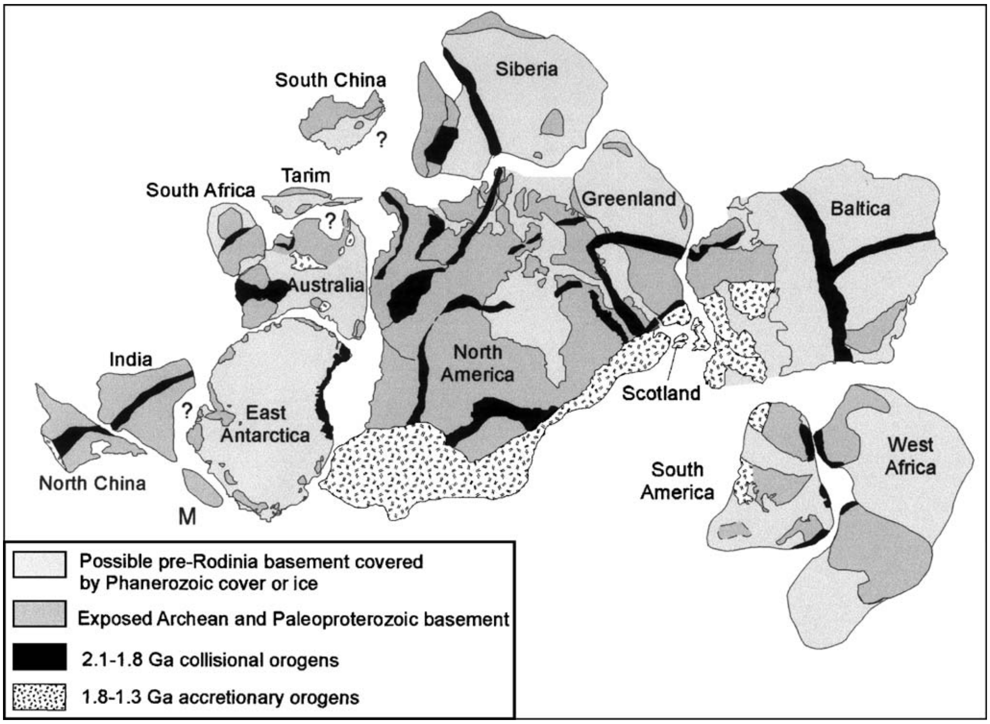

Fig. 17. A possible reconstruction of Paleo-Mesoproterozoic supercontinent Columbia (revised after Zhao et al., 2002a). 1.8–1.3 Ga accretionary orogens include the 1.8–1.7 Ga Yavapai and Central Plains Belts, 1.7–1.6 Ga Mazatzal Belt, 1.5–1.3 Ga St. Francois and Spavinaw Granite–Rhyolite Belts and 1.3–1.2 Ga Elzevirian Belt in the southwestern North America; the 1.8–1.7 Ga Makkovikian Belt and 1.7–1.6 Ga Labradorian Belt in northeastern North America; the 1.8–1.7 Ga Ketilidian Belt in Greenland; the 1.8–1.7 Ga Malin Belt in the British Isles; the 1.8–1.7 Transscandinavian Igneous Belt, 1.7–1.6 Ga Kongsberggian–Gothian Belt, and 1.3–1.2 Ga early Sveconorwegian Belt in Baltica; the 1.80–1.45 Ga Rio Negro–Juruena Belt and 1.45–1.30 Ga Rondonian–San Ignacio Belt along the western margin of South America; the 1.8–1.5 Ga Arunta, Musgrave, Mt. Isa, Georgetown, Coen and Broken Hill inliers along the southern and eastern margins of the North Australian Craton; and the 1.8–1.4 Ga Xiong'er Belt along the southern margin of the North China Craton. Names of 2.1–1.8 Ga collisional orogen are shown in Fig. 2b.

Following its final assembly at ~ 1.8 Ga, the supercontinent Columbia underwent a long-lived, subduction-related outgrowth along some of its continental margins (Fig. 17), forming a number of accretionary zones, of which the most representatives are the huge 1.8–1.3 Ga magmatic accretionary belt bordering the present southern margin of North America, Greenland and Baltica; the 1.80–1.45 Ga Rio Negro–Juruena Belt and 1.45–1.30 Ga Rondonian–San Ignacio Belt along the western margin of South America; the 1.81.5 Ga Arunta, Musgrave, Mt. Isa, Georgetown, Coen and Broken Hill inliers along the southern and eastern margins of the North Australian Craton; and the 1.81.4 Ga Xiong'er Belt along the southern margin of the North China Craton.

Fragmentation of this supercontinent began about 1.6 Ga ago, in association with development of continental rifting along the western margin of Laurentia (Wernecke, Muskwa, Belt, Purcell, Uinta, Unkar and Apache supergroups or groups), southern margin of Baltica (Telemark Supergroup), southeastern margin of Siberia (Riphean aulacogens), northwestern margin of South Africa (Kalahari Copper belt), and northern margin of North China (Zhaertai-Bayan Obo belt). The fragmentation was also associated with widespread anorogenic activity including emplacement of anorthosite massifs, charnockite intrusions, diabase dyke swarms and potassic rapakivi granites, and eruption of coeval rhyolites, especially in North America, Greenland, Baltica and South America. The final breakup of the supercontinent occurred at about 1.31.2 Ga, marked by the emplacement of the McKenzie, Sudbury, Seal Lake, Harp and Mealy mafic dike swarms and coeval eruption of flood basalts (e.g. Coppermine River basalts) in North America. This indicates that the breakup of the supercontinent Columbia was followed rapidly by the assembly of another supercontinent Rodinia, which was assembled along the globally distributed Grenvillian orogens at ~ 1.0 Ga (Dalziel, 1995, Dalziel et al., 1997, 2000).

At present, a full configuration for the proposed supercontinent Columbia is not possible because of some unresolved issues, especially concerning controversial fits of Siberia with Laurentia (Sears and Price, 1978, 2000; Condie and Rosen, 1994; Frost et al., 1998, Rainbird et al., 1998; Smethurst et al., 1998; Ernst et al., 2000; Piper, 2000, 2003); Australia/East Antarctica with Laurentia (Moores, 1991; Ross et al., 1992; Borg and DePalo, 1994; Blewett et al., 1998; Burrett and Berry, 2000, 2002; Karlstrom et al., 2001; Wingate et al., 2002); South America with Laurentia (Sadowski and Bettencourt, 1996; Dalziel, 1997; Sadowski, 2002); and North China with Siberia, Baltica or India (Li et al., 1996; Condie, 2002; Wilde et al., 2002; Zhao et al., 2003a). Nevertheless, we believe that the weight of evidence supports the existence of a Paleo-Mesoproterozoic supercontinent (Fig. 17) that was assembled during global-scale collisional events in the period 2.11.8 Ga, underwent long-lived (1.8–1.3 Ga) subduction-related accretion, was fragmented between 1.6 and 1.2 Ga and finally broke up at ~ 1.2 Ga.

Acknowledgements

We thank John J.W. Rogers and Georg R. Sadowski for their constructive comments that helped to clarify our discussion. This research was financially supported by Hong Kong RGC Grants (7090/01P, 7055/03P and 7048/03P), HKU Seed Funding for Basic Research Programs and a China NSFC grant (40002015).

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