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Geochronological data evaluation: Implications for the Proterozoic tectonics of East Gondwana

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Abstract

Geochronologic data and arguments proposing the division of the pre-Pan-African terranes of East Antarctica (Grenvillian Circum-East Antarctic Orogen) into three genetically different provinces, and the recognition of the Pan-African orogens as sutures are discussed. Problems of zircon U–Pb dating for identifying the absence of precursor Grenvillian events, which are superimposed and concealed by the dominant Pan-African event, are specifically highlighted. Major problems are (i) the lowering of closure temperatures under deformational and/or fluid-rich conditions, (ii) the weighted mean U–Pb concordant ages, (iii) the disappearance of an earlier discordia in the polymetamorphism, (iv) diversity in the interpretation of zonal structure of zircons, (v) general lack of discussion on the formative process of the dated zircons and (vi) the practical difficulty in zircon geochronology for providing evidence of the absence of earlier events superposed by a later high-grade event. It is stressed that it is difficult for zircon geochronology to provide robust evidence for eliminating the possibility of the existence of earlier tectonothermal events superimposed by later high-grade events. The presence of Grenvillian-age basement within the Pan-African terranes and the former linkage of now-separated Grenvillian age belts in East Antarctica cannot therefore be excluded from U–Pb zircon evidence alone.
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Keywords: Zircon geochronology; Age rejuvenation; East Antarctica; East Gondwana; Circum-East Antarctic Orogen; Pan-African Orogeny

1. Introduction

The existence of East Gondwana during the Neoproterozoic has been a subject of major recent debate in the geoscience world (e.g., Meert, 2001; Yoshida et al., 2003; Meert and Torsvik, 2003; Collins and Pisarevsky, 2005) mainly from arguments based on geochronologic data from Antarctica. The ~1000 Ma (actually ca. 900–1300 Ma) Circum-East Antarctic Orogen (CEAO) defined by Yoshida (1992, 1995a) as having assembled East Gondwana has been questioned on one hand from the distinct age differences within the orogen (Fitzsimons, 2000), and on another from evidence for the development of two Pan-African orogens crossing the CEAO (Fig. 1). As a result the hypothesis of the Pan-African assembly of East Gondwana has been proposed (Powell et al., 2001; Meert, 2001; Meert and Torsvik, 2003; Fitzsimons, 2003; Pisarevsky et al., 2003) and has resulted in some major re-considerations of the classical Rodinia configuration (e.g., Dalziel, 1991), as have been earlier attempts by some authors based mainly on palaeomagnetic studies (e.g., Meert et al., 1995; Meert and Van der Voo, 1997; Hensen and Zhou, 1997). The proposal by Fitzsimons (2000) is considered incomplete on the basis of three lines of evidence:

1. There is little distinct difference in age between the western and central sector of CEAO, with regard to the zircon and Sm–Nd mineral-whole rock ages as well as from the Rb–Sr whole-rock isochron ages. There is a difference in age of some hundred million years between the eastern sector and the remaining sectors of the CEAO. However, since a large orogen such as that fringing the East Antarctic continent might have been the result of the convergence of various plates at different times, precise synchronism of major orogenic phases is not necessarily to be expected.
2. Occurrence of rocks with ca. 1000 Ma ages is not rare within the two Pan-African orogens in East Antarctica. Furthermore, these terrains are characterized by superimposed

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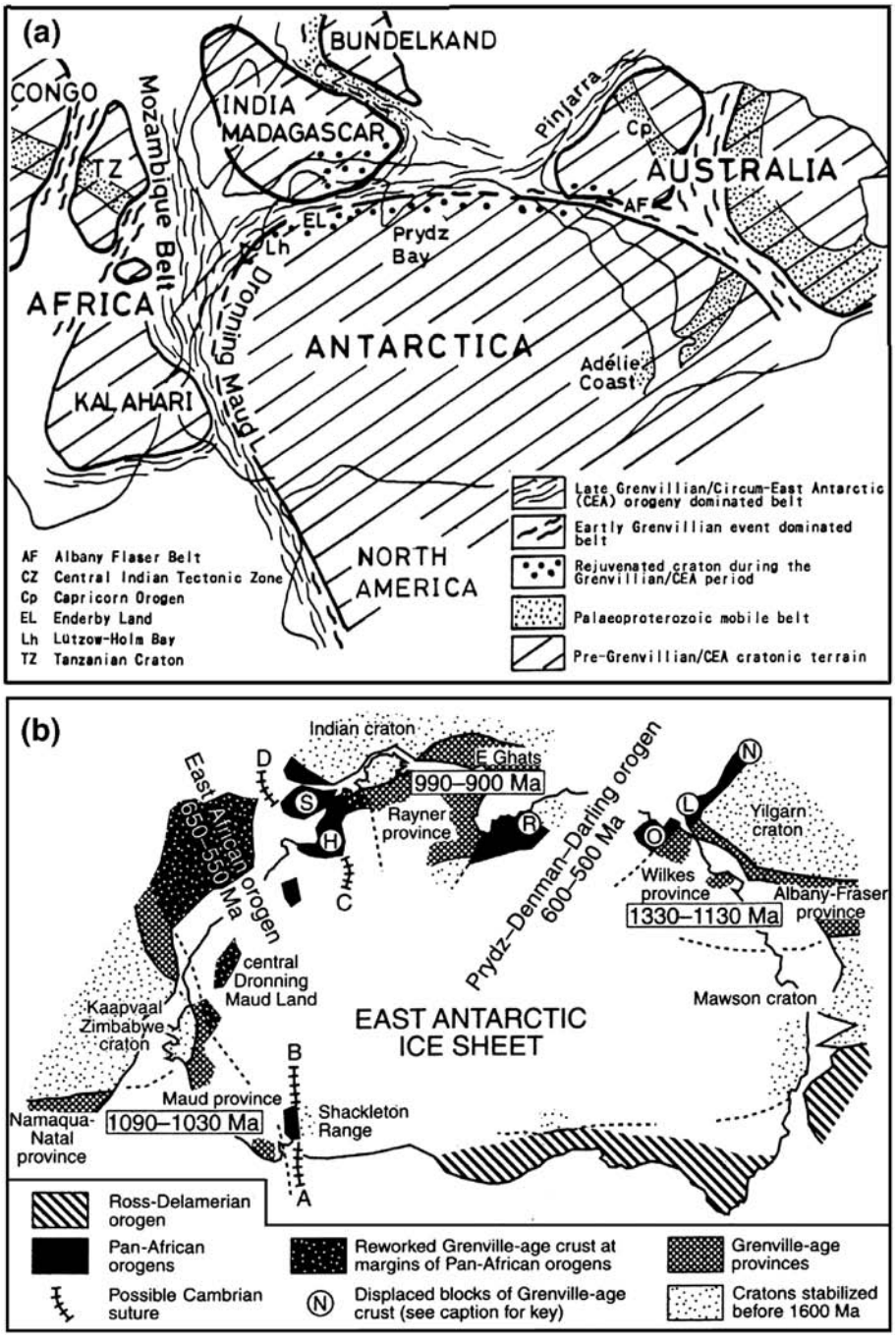


Fig. 1. (a) The Mesoproterozoic Grenvillian Circum-East Antarctic Orogen and its extensions in East Gondwana according to the classical model (modified after Yoshida, 1996). (b) Pre-Pan-African provinces and Pan-African orogens in East Antarctica (Reprinted from Fig. 2 of Fitzsimons, 2000, Geology, vol., 28, page 880).

deformations and metamorphisms, and by the lack of ophiolite. Such evidence supports the view that the Pan-African orogens are intracratonic rather than intercratonic; that is, they are not the result of suturing.

3. Detection of ca. 1000 Ma zircon ages as well as of Sm–Nd mineral-rock ages is rare or absent on rocks which initially had geochronologic components of ca. 1000 Ma and this could happen by the superposition of Pan-African events including strong deformational, high-temperature and/or fluid-rich metamorphism. Thus, the lack or scarcity of ca. 1000 Ma ages does not guarantee the lack of ca. 1000 Ma events within the Pan-African orogen in East Antarctica.

The present paper aims to discuss the geochronologic problems related to the above points from several examples of dating by U–Th–Pb and Sm–Nd and Rb–Sr systematics on rocks and minerals from the CEAO and Pan-African orogens in East Antarctica. Data from other aspects of investigation such as geology and palaeomagnetism are also relevant to the subject of East Gondwana tectonics but are beyond the scope of the

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present paper. General problems on these aspects have been discussed by Yoshida et al. (2003) and will not be taken further here.

2. Geochronologic data supporting the Pan-African assembly model of East Antarctica

Fitzsimons (2000) showed that pre-Pan-African terrains fringing East Antarctic crust, which had formerly been identified as the Circum-East Antarctic Orogen (CEAO) ~1000 Ma in age by Yoshida (1992, 1995a, 1996) do not form a single orogen. He considered that the CEAO is divided into three different crustal blocks with distinct ages and orogenic peaks separated by two Pan-African orogens.

Fitzsimons (2000) showed that pre-Pan-African terrains fringing East Antarctica are composed of, from east to west, the Wilkes Province, (peak orogenic ages of 1330–1130 Ma), the Rayner Province (990–900 Ma), and the Maud Province (1090–1030 Ma). The Pan-African Prydz-Denman-Darling Orogen (600–500 Ma) and East African Orogen (650–550 Ma), the latter having been shown to include the area from Lützow-Holm Bay (LHB) to Central Dronning Maud Land (CDML), separate the above three pre-Pan-African provinces (c.f. Fig. 1b). In the discussion Fitzsimons (2000) relied mostly on zircon U–Pb age data, and ages from other methods, notably whole-rock isochron ages, were discarded.

3. Age differences within the Circum-East Antarctic Orogen

Fitzsimons (2000) highlighted a distinct age difference between the Wilkes (eastern), Rayner (central) and Maud (western) provinces of the Mesoproterozoic terrains fringing East Antarctica (Fig. 1b), which were considered to form a continuous orogen (the CEAO, Fig. 1a) consolidated during the Grenvillian episode (ca. 900–1300 Ma) by James and Tingey (1983), Yoshida and Kizaki (1983), Yoshida (1992, 1995a, 1996), Yoshida et al. (2003).

Yoshida and Yoshida et al. considered that the eastern part of the CEAO (Albany-Flaser Orogen) experienced the earlier event of the orogeny, and that the orogen with the major orogenic period of ca. 1000–1100 Ma swings to the north into the Pinjarra Orogen (*sensu stricto*, Myers, 1993) of the western margin of Western Australia. There is thus no problem in the recognition of a relatively older signature at the eastern sector of the CEAO.

According to James and Tingey (1983), Yoshida and Kizaki (1983) and Yoshida (1995a, 1996), all other parts of CEAO, i.e. the western margin of Australia (Pinjarra Orogen, Myers, 1993), western part of the Wilkes Land, Queen Mary Land, Prydz Bay, Northern Prince Charles Mountains, Soya Coast to Lützow-Holm Bay, Eastern–Central–Western Dronning Mound Land are characterized by the superposition of low grade (western Dronning Maud Land) to medium-high grade (all other areas) Pan-African events over the ca. 1000–1200 Ma interval of orogeny. During the early 1980's, these areas except the Central Dronning Maud Land were found to have preserved ca. 1000–1200 Ma K–Ar, Rb–Sr, and/or U–Th–Pb ages (c.f. Yoshida and Kizaki, 1983 and references therein), whilst zircon U–Pb ages of the above range have appeared sporadically from the Rayner Complex (e.g., Black et al., 1987; White and Clarke, 1993), Prydz Bay (e.g., Kinny et al., 1993; Zhang et al., 1996), Lützow-Holm Bay (Shiraishi et al., 1992), and Central Dronning Maud Land (Jacobs et al., 1998).

Furthermore, rocks occurring in all these areas have a similar Nd model age range of around 1.3 to 2.0 Ma (Young et al., 1997; Zhao et al., 1997; Jacobs et al., 1998; Wareham et al., 1998; Yoshida et al., 1999; Mikhalsky et al., 1999). Thus, even though the CEAO is transected by distinct Pan-African orogens as mentioned below, all the above areas may well have formed a continuous orogen, the CEAO, and the Pan-African orogens may have superposed over parts of the CEAO and strongly rejuvenated it as pointed out by Yoshida et al. (2003).

Fitzsimons (2000) highlighted a distinctly younger age signature (ca. 990–900 Ma) in the central sector of the CEAO (Rayner Province) in contrast to the eastern and western sectors. He noticed this mostly from zircon U–Pb ages and discarded whole-rock isochron ages, which were formerly taken into account by several authors (e.g., Black et al., 1979, 1987; Tingey, 1982; Yoshida, 1995a) that there was a prevalence of ca. 1000–1100 Ma ages even up to 1300 Ma in the Rayner Complex. Yoshida et al. (2003) show that there are several U–Pb zircon ages as well as Sm–Nd mineral-whole-rock ages, in addition to the Rb–Sr whole-rock isochron ages supporting the prevalence of a ca. 1000–1100 Ma event in the Rayner Complex (Fig. 2). In Fig. 2, we observe U–Pb zircon ages of ca. 1000–1289 Ma scattered in both the eastern Rayner Complex and Northern Prince Charles Mountains. Sm–Nd mineral-whole-rock ages of ca. 1000–1150 Ma are also reported from Northern Prince Charles Mountains (Hensen and Zhou, 1995; Mikhalsky et al., 1999; Boger et al., 2000); they are as important as the U–Pb zircon ages since these kinds of age are considered to reflect the closure temperature of the Sm–Nd systematics within the mineral, and therefore, indicate that the peak metamorphism is a little earlier. Widespread ca. 1000–1300 Ma ages of zircon U–Pb and Rb–Sr whole rock isochron ages, which were discarded by Fitzsimons (2000), attest to the above argument. It has been pointed out that the above geochronologic characteristics are also recognized in the Eastern Ghats Belt of South India (e.g., Dobmeier and Raith, 2003), a belt which is considered to have been juxtaposed with the Rayner Complex during the Meso-Neoproterozoic period (Yoshida, 1995a; Yoshida et al., 2003; Dasgupta and Sengupta, 2003; Harley, 2003 and Fig. 1a).

There are apparently many younger ages of ca. 900–990 Ma in the Rayner Complex as summarized by Fitzsimons (2000) and supported by several references (e.g., Black et al., 1987; Harley, 2003). Even though considerable portions of the Rayner Complex may have suffered high-grade events at relatively later times than other areas of the Circum-East Antarctic Orogen, considerable numbers of these younger ages are considered to reflect late tectonic charnockite activity at around 945 and 985 Ma and its rejuvenation effect in the wider areas of the Rayner Complex, since the Mawson charnockite clearly

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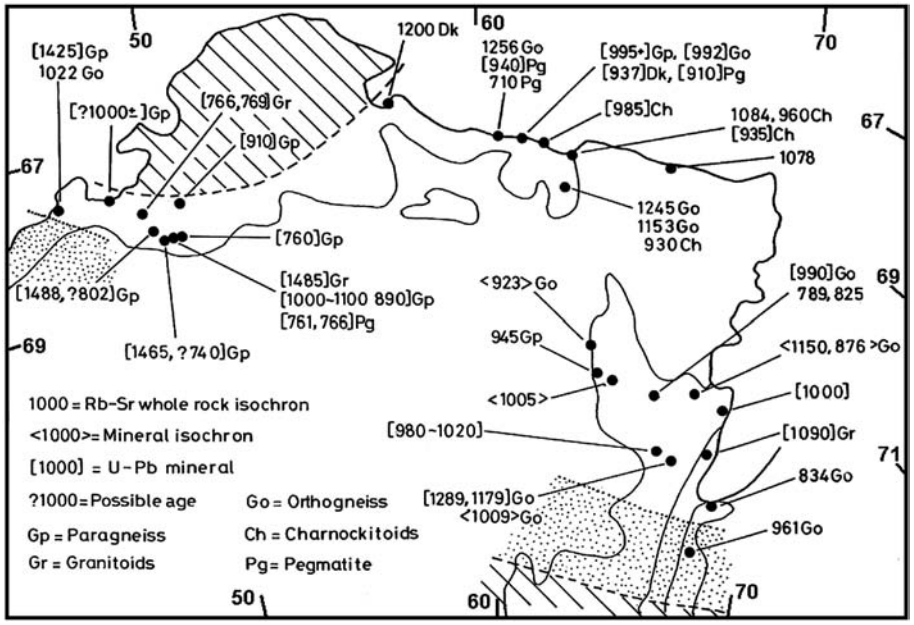


Fig. 2. Grenvillian ages in the Rayner Belt. Data are mostly from Sheraton et al. (1987), with some new data cited in the text. The stippled area is considered to be relatively younger rocks due to lower Sr initial ratios (Yoshida, 1994).

postdates the earlier deformational events recorded within the country gneisses (e.g. Young and Ellis, 1991; White and Clarke, 1993; Dunkley et al., 2002). It is pointed out that even though the Rayner Complex may be younger than other parts of the CEAO by some tens of million years, it can not necessary be considered to form a different orogen as maintained elsewhere.

In conclusion, there is no or small distinct age difference within the CEAO, except in the easternmost area (the Albany-Fraser Orogen). As mentioned below, this observation supports the continuation of the CEAO throughout its full length from the Western Dronning Maud Land to Western Australia (Pinjarra Orogen) as well as to the southwestern Australia (Albany-Fraser Orogen). This conclusion is reinforced by the sporadic occurrence of similar Grenvillian ages within the Pan-African orogens dissecting the CEAO. There do appear to be some tens to some hundreds of millions of years of age difference within the CEAO; but it may be pointed out that a large orogen should not necessarily exhibit peak orogenic ages throughout.

4. Geochronologic characteristics of Pan-African orogens in East Antarctica

4.1. Lützow-Holm Bay-Central Dronning Maud Land belt

The Lützow-Holm Bay (LHB)- Central Dronning Maud Land (CDML) Pan-African belt is one of the two possible major Pan-African sutures proposed by Fitzsimons (2000). This possible suture has also been recognized by Kriegsman (1993), Kleinschmidt and Buggisch (1994), Grunow et al. (1996), Jacobs et al. (1998), Kleinschmidt et al. (2002) and several other authors from a variety of considerations. This belt is generally regarded as the continuation of the eastern boundary fault of the Mozambique Belt to the north (in African coordinate), and connects to the Shackleton Range of the Transantarctic Mountains to the south where the Pan-African suture has been reported (Kleinschmidt and Buggisch, 1994; Talarico et al., 1999).

Geologically, this belt is characterized by strong deformation, high-grade amphibolite to granulite metamorphism and granitoid intrusions of Pan-African age, although ophiolites are developed only in the Shackleton Range (Talarico et al., 1999) and in the Mozambique Belt (Shackleton, 1993) including Madagascar (Collins et al., 2003) and nothing is known from the major part of East Antarctica.

Geochronologically, both the LHB and CDML areas are characterized by the superposition of ca. 500–600 Ma Pan-African ages over the ca. 1000–1100 Ma CEAO (Grenvillian) ages. LHB area is one of the first areas where the Pan-African orogeny was identified from zircon datings (Shiraishi et al., 1992, 1994). These authors attempted the first SHRIMP dating on 8 gneissic samples from throughout the LHB, Yamato Mountains and Belgica Mountains, and found that almost all samples carry extreme Pan-African ages, while only 4 samples carry ca. 1000 Ma ages. From their data, Shiraishi et al. (1992, 1994) concluded that the LHB area experienced only the ca. 500–600 Ma Pan-African orogeny. Four samples carrying ca. 1000 Ma ages were regarded as belonging to the Grenvillian terrain, from which they are either tectonically separated (e.g. Klippe) from surrounding terrains or their zircons are of detrital origin. Thus, Shiraishi et al. (1994) considered that the protolith of the LHB area is mostly Neoproterozoic sediments, with a maximum age provided by ca. 1000 Ma detrital zircon ages. The Pan-African timing of the high-grade metamorphism of the LHB area was further ascertained by Shimura et al. (1998) and Fraser et al. (2000).

The LHB area, however, has generally been regarded as belonging to the Grenvillian episode (e.g., Yoshida et al., 1983)

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based on Pb-Pb and Rb–Sr whole rock isochron ages (Maegoya et al., 1968; Shirahata, 1983; Shibata et al., 1986). Yoshida (1978, 1994, 1995b) and Yoshida et al. (2003) stress that the geology of the LHB area carries extensive evidence of superposed orogenic events, and these authors interpreted not only whole-rock isochron ages but also zircon SHRIMP ages to reflect the superposition of 1000 Ma and 500 Ma events. They considered that the effects of the ca. 500 Ma events would make the earlier ca. 1000 Ma event difficult to recognize. It should be noted that the LHB area is actually one of the areas where the Grenvillian zircon SHRIMP ages are well documented in the CEAO.

Extensive Pan-African ages have been detected in the CDML. Superimposition of amphibolite facies metamorphism associated with granitic activities over the granulite facies has been recognized properly within the CDML (e.g., Ravich and Soloviev, 1969; Sengupta, 1986). Sengupta (1993) considered that these events are Pan-African events superposed over Grenvillian events, and correlated them with those in the LHB area to the east; her view has recently been supplemented by a conventional zircon U–Pb study (Mikhalsky et al., 2003). In the Orvinfjella of the mainland CDML, Jacobs et al. (1998) reported zircon SHRIMP data identifying igneous and metamorphic Grenvillian (ca. 1137–1075 Ma) and Pan-African (ca. 608–516 Ma) ages. Their data have been widely accepted as pointing to the superposition of Pan-African high-grade magmatic-metamorphic events over Grenvillian basement in the CDML (e.g., Jacobs and Thomas, 2002; Bauer et al., 2003; Mikhalsky et al., 2003; Baba et al., 2003).

4.2. Prydz Bay area

The Prydz Bay area was generally considered to form a part of the CEAO characterized by ca. 1000 Ma high-grade events overprinted by local granitic activity and low-grade metamorphism (e.g., Tingey, 1982; James and Tingey, 1983; Yoshida and Kizaki, 1983; Sheraton et al., 1987; Kinny et al., 1993; Yoshida, 1995a; Sheraton et al., 1996). Both Grenvillian and Pan-African ages have been detected in this area, and distinct polymetamorphic and poly-deformational events have been detected (e.g., Sheraton et al., 1987; Dirks and Wilson, 1995).

Following the detection of Pan-African zircon Pb-Pb ages from late tectonic granite from the Larsemann Hills in the southeastern coast of the Prydz Bay by Zhao et al. (1992, 1993), the above classical idea has become suspect, especially on the east coastal area of the bay, in the context of the detection of Pan-African U–Pb and Pb-Pb zircon ages in these areas (e.g., Zhao et al., 1995; Carson et al., 1996; Fitzsimons et al., 1997). Zhao et al. (1995) even insisted that the protolith of paragneisses are younger than ca 1000 Ma, i.e., of Neoproterozoic sedimentation.

Referring to these geochronologic data and discussions, some workers have proposed a Pan-African suturing event surrounding Prydz Bay (Harley et al., 1995; Dirks and Wilson, 1995; Fitzsimons, 1997; Scrimgeour and Hand, 1997; Hensen and Zhou, 1997). Distinct Pan-African events have also been detected in the northern Prince Charles Mountains (Boger et al., 2000) as well as in the Grove Mountains (Zhao et al., 2000; Mikhalsky et al., 2001), which appear to support the extension of the Pan-African Prydz Bay belt (e.g., Boger et al., 2001; Fitzsimons, 2003; Zhao et al., 2003).

There are however, several ca. 1000 Ma measurements from zircon U–Pb, Pb-Pb ages, Sm–Nd garnet-whole-rock ages and Ar-Ar ages (Kinny et al., 1993; Hensen and Zhou, 1995; Zhang et al., 1996; Fitzsimons et al., 1997; Hensen et al., 1997; Zhao et al., 2000; Tong et al., 2002). In the central- to south-eastern Prydz Bay area, all the reported Grenvillian ages are from orthogneisses. This fact has been interpreted by many researchers that the orthogneisses represented the basement rocks of earlier times suffered Grenvillian metamorphism, while the paragneisses are of Neoproterozoic (e.g., Zhao et al., 2003). Hence, whether the eastern Prydz Bay area represents a suture is still a matter of debate (e.g., Dirks and Wilson, 1995; Tong et al., 1995; Zhang et al., 1996; Tong and Liu, 1997; Yoshida et al., 2003).

5. Discussion

Whether either one or both of the Pan-African (LHB-CDML and Prydz Bay) orogens in East Antarctica are Pan-African sutures or not it still is a subject of dispute. A key observation to identify a suture is the distinct difference in the palaeolocation of the crusts on both sides of the orogen. Unfortunately a proposal for Pan-African suturing in East Antarctica based on geochronological discussions, has prompted scientists of other subject areas most notably palaeomagnetism, to propose a palaeogeography based on inadequate data as pointed out by Yoshida et al. (2002, 2003, notably p. 69). At present critical examination of the geochronologic database has the key role to play in the recognition of suture(s) in East Antarctica.

The geochronological work supporting the recognition of the suture(s) in East Antarctica has mostly focused on one hand on the absence of Grenvillian *in-situ* zircon ages (ca. 900–1300 Ma) within the Pan-African orogens, and on another on age differences within the pre-Pan-African terrains of the CEAO. Since the data and arguments concerning the latter point are given above, only the former point will be discussed in this section.

Hölzl et al. (1994) noted the non-detection of Grenvillian ages in the granulite grade Highland Complex of Sri Lanka, which juxtaposes with the LHB area within the CEAO, and concluded that there was no Grenvillian metamorphism within this complex, although this region has formerly been regarded as metamorphosed during that episode (e.g., Cooray, 1984; Kröner et al., 1987; Yoshida et al., 1990), and good evidence for Grenvillian zircon age was reported by Kröner et al. (1987). Shiraishi et al. (1992, 1994) analyzed 8 gneissic samples from 8 localities in the LHB to Eastern Dronning Maud Land area, and since 4 samples did not carry a Grenvillian age, they concluded that it formed a Cambrian orogen with no preceding record of Grenvillian metamorphism. In contrast the same area was regarded principally as a Grenvillian orogen by Yoshida et al. (1983). As noted above a similar contention has also concerned the Prydz Bay area where Zhao et al. (1992) found only the Pan-African zircons from late tectonic granite. Zhao

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et al. (1995) considered that the protolith of paragneisses of the southeastern Prydz Bay area are of Neoproterozoic age, younger than the ≥1000 Ma Grenvillian episode, and that the area comprises a Cambrian suture. In contrast, this region has been regarded to form a principal part of the Grenvillian belt surrounding East Antarctica by other workers (Tingey, 1982; James and Tingey, 1983; Yoshida and Kizaki, 1983; Sheraton et al., 1987, 1996). Observations that question the Pan-African suturing hypothesis for the orogens in East Antarctica are summarized below.

5.1. Lowering of closure temperatures of U–Pb and Sm–Nd systematics under deformational and/or fluid-rich conditions

There are many data and discussions on the lowering of closure temperatures of U–Th–Pb systematics in zircons and Sm–Nd systematics in garnet and rocks (Fig. 3a,b,c). From experimental evidence Miura et al. (1980), Caruba et al. (1985) and Lee et al. (1997) and from field evidence Garanin et al. (1993), and Fitzsimons (1996, 1997) show that mobility of U–Pb systems occurs only above 850–900 °C, whilst Mezger et al. (1989) and Carson et al. (1996, 1997) show in contrast, that it occurs as low as 800 °C (Fig. 3a,b). These data and arguments have often been cited by later authors to support the time of high-grade metamorphic events identified from zircon ages.

However from experimental studies, Rizvanova et al. (2000) have shown that mobility of U–Pb systematics and recrystallization occur at around 400 °C in metamict zircons and above 650 °C in crystalline zircons under hydrothermal conditions. The mobility of U–Th–Pb systematics at lower temperature has also been clearly shown from field studies by Gebauer and Grünenfelder (1976), Gebauer and Grüenenfelder (1978) and by Sinha and Gloer (1978) down to amphibolite facies conditions (with or without concurrent deformation), and Gebauer and Grünenfelder (1976), Gebauer and Grüenenfelder (1978) further showed that it could occur down to greenschist facies conditions. Similar evidence was identified by several authors in both experimental (e.g., Poutiainen, 1995) and field studies (e.g., Wyne and Singha, 1988; Rubin et al., 1989; Hoskin et al., 1998; Schaltegger et al., 1999; Hartmann et al., 2000; Da Silva et al., 2000; Hogdahl et al., 2001; Mathieu et al., 2001; Liermann et al., 2002). It is noteworthy that there are many examples of the mobility around 500–400 °C under fluid-rich conditions which may or may not be associated with deformation.

The closure temperature of Sm–Nd systematics in garnet and in rock has received analogous studies (Fig. 3c). Studies by Cohen et al. (1988) showed that the closure temperature of Sm–Nd systems within garnet is around 900 °C or higher, a finding supported by the work of Foland et al. (2000) and cited in a variety of studies concerned with the geochronology of high-grade terrains. However, Hensen and Zhou (1995) showed that the mobility of Sm–Nd systematics in garnet can occur down to 700 °C and Mezger et al. (1992) identified mobility at temperatures down to ca. 550 °C. Several further field and experimental studies have shown that the mobility of rare earth elements within a rock can occur as low as ~250 °C under fluid-rich conditions (e.g., Tourpin et al., 1991; Gruau et al., 1992; Valsami-Jones and Cann, 1994; Lahaye et al., 1995).

The low temperature mobilities of both U–Pb and Sm–Nd systematics in zircon, garnet and in rocks support the view that resetting of older ages can occur during a later event. Such a factor could make it difficult to detect Grenvillian ages in the high-grade Pan-African terrains of East Antarctica.

5.2. Rejuvenation of zircon and monazite ages through a later tectonothermal event

As discussed above, deformational and fluid-rich conditions considerably lower the closure temperature of isotopic systematics of rocks and minerals, and thus, promote the rejuvenation of rock/mineral ages during later orogenesis. This phenomenon has long been known to rejuvenate isotopic systems in rocks occurring within shear zones that typically carry younger ages than the less deformed country rocks (e.g., Black et al., 1979).

As deformational effects are unequal from grain to grain according to the locus of the stress and unevenness of the micro domain, the rejuvenation itself is also unequal. Theoretically and almost always practically, more strained grains are more strongly rejuvenated. There are numerous reports of the rejuvenation of zircon and monazite ages from metamorphic terrains notably associated with detailed examination of microstructures (e.g., Black et al., 1987; Kröner et al., 1994; Ashwal et al., 1999; Da Silva et al., 2000).

An interesting recent study has identified evidence for extensive and strong resetting of monazite ages in Himalayan gneisses (Martin et al., 2006). Here the U–Th–Pb ages of xenocrystic monazite in metamorphic garnet show a smooth distribution with the maximum concentration of ages within the timeframe of the Miocene Himalayan metamorphism and a negligibly small distribution at ca. 500 Ma when the majority of the monazite grains are considered to have crystallized.

5.3. Quasi-concordant U–Pb zircon ages

From the principle of U–Th–Pb isotopic systematics, a concordant age for a zircon grain (or a micro domain) on $^{207}Pb/^{235}U$ against $^{206}Pb/^{238}U$ concordia diagram (Wetherill, 1956) has been regarded as the time of formation and of complete lead loss in the zircon. However, the concordant age may carry an uncertainty of some tens of million years. This is because the curvature of the concordia curve between the above time interval is small and a partial lead loss, which occurs some tens or hundreds of million years later than the earlier concordant zircon formation, often results in a quasi-concordant age some tens of million years younger than the time of the initial zircon formation (e.g., Black and Sheraton, 1990; Ashwal et al., 1999).

As a result of this uncertainty a group of scattered concordant or near concordant ages on zircons may be obtained. It is sometimes better to estimate the maximum age of the group as the minimum age of the earlier concordant zircon age, and the youngest age of the age group to suggest the maximum age of the later event resulting in the partial lead loss from the

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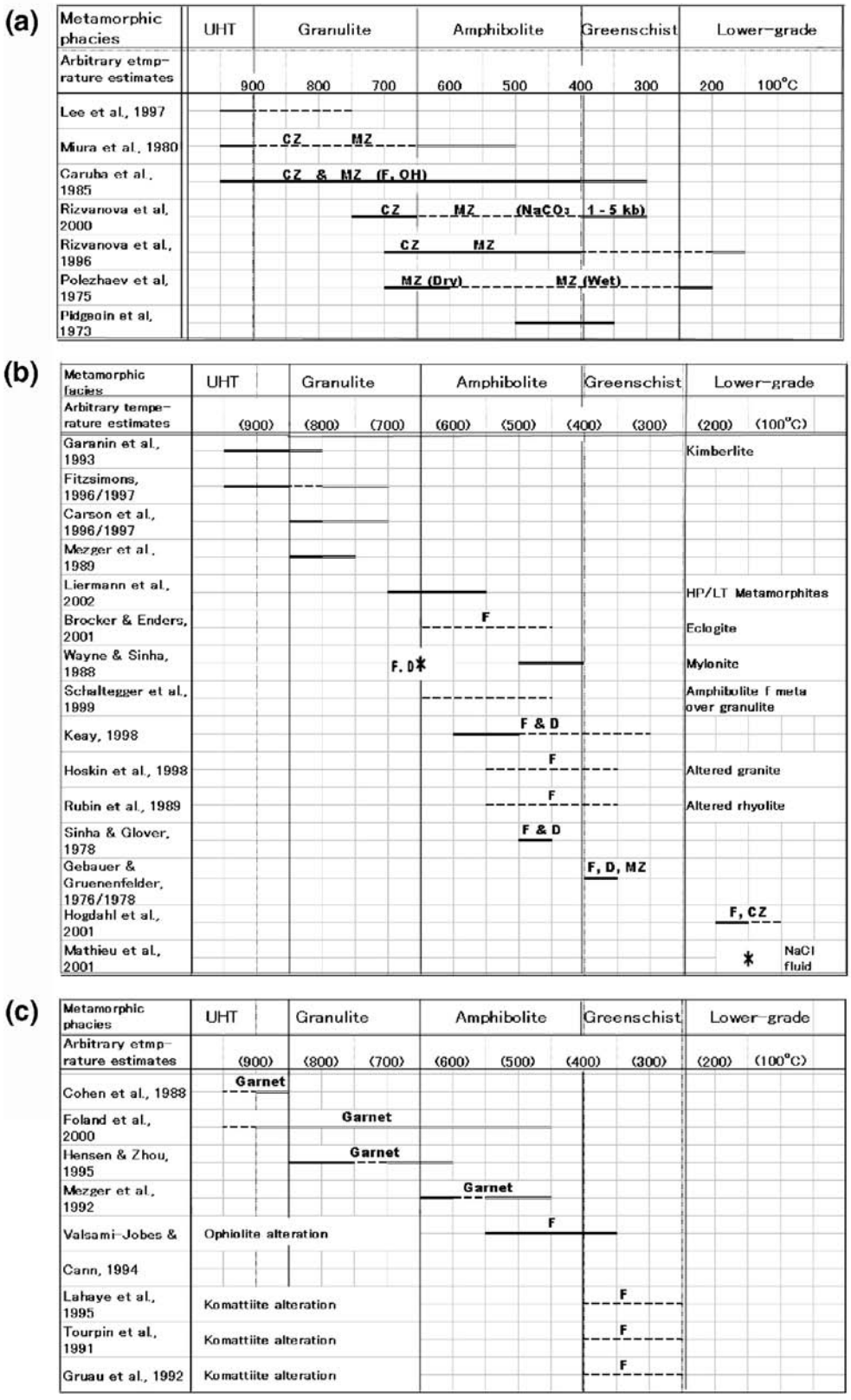


Fig. 3. Element mobility in zircon, garnet and rocks under a variety of temperature conditions. The solid line and an asterisk denote the mobility observed in that range of temperature or temperature point. Dashed line indicates the mobility observed imprecisely or the mobility considered possible. The double line indicates ranges where mobility was not observed. F: Fluid-rich condition. D: Deformational condition. CZ: crystalline zircon, MZ: metamict zircon. (a) Mobility of U–Pb systematics in zircons (Experimental studies). (b) Mobility of U–Pb systematics in zircons (Field studies). (c) Mobility of Sm–Nd systematics in garnet and rocks.

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zircons (Fig. 4a,b). Thus the weighted mean age obtained on a concordia of $^{238}U/^{206}Pb$ against $^{207}Pb/^{206}Pb$ concordia diagram of the kind used by Tera and Wasserburg (1972) on a scattered group of U–Pb concordant/near concordant ages may yield a misleading metamorphic age from study of a polymetamorphic terrain. Several ca. 700–800 Ma events have been identified in Enderby Land and Northern Prince Charles Mountains (e.g. Shiraishi et al., 1997; Hensen et al., 1997). There is a possibility that these events have also effectively rejuvenated preexisting Grenvillian ages in these areas, as was the result of 940–980 Ma activity in the Mawson Charnockite discussed elsewhere above.

5.4. The Concordia-discordia method problem

The concordia-discordia method on the $^{207}Pb/^{235}U$ against $^{206}Pb/^{238}U$ concordia diagram has been considered to be very effective for analyzing polymetamorphic terrain (e.g., Faure, 1988), and a variety of good results has been published from several high-grade terrains. The lack of lower intercept Grenvillian ages in granulites was one of major reasons why

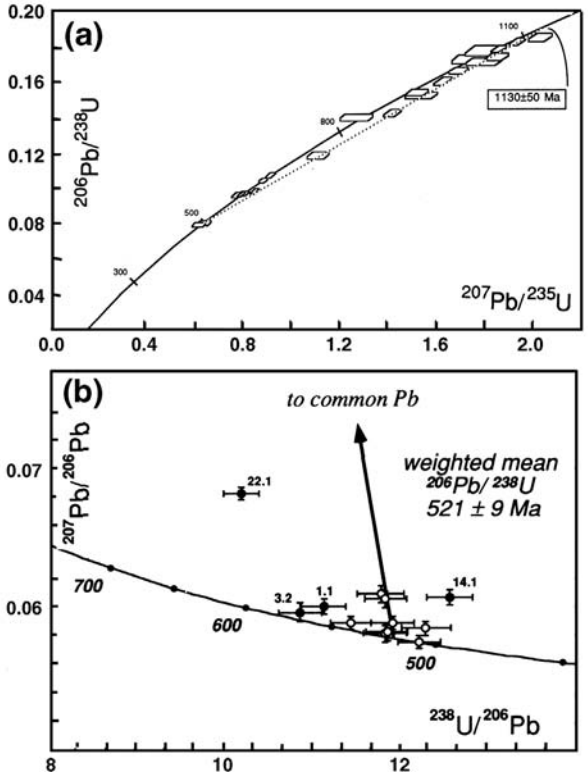


Fig. 4. Quasi-concordant age problems associated with zircon U–Pb ages. (a) Apparent concordant ages on $^{207}Pb/^{235}U$ against $^{206}Pb/^{238}U$ concordia diagram observed for a granite sample from Northern Victoria Land, Antarctica (Reprinted from Precambrian Research, vol. 48, Black and Sheraton, 1990, with the permission of Elsevier). There are several apparent concordant ages between ca. 500 and 1100 Ma. Some are considered to be rejuvenated ages by the authors. (b) A group of concordant and semi-concordant ages on the $^{238}U/^{206}Pb$ against $^{207}Pb/^{206}Pb$ concordia diagram, observed for a gneiss sample from Lützow-Holm Bay, Antarctica (Reprinted from Fig. 7 of Shiraishi et al., 1994, Jourl of Geology, vol. 102, page 54). The authors in this case used a weighted mean to derive a peak metamorphic age.

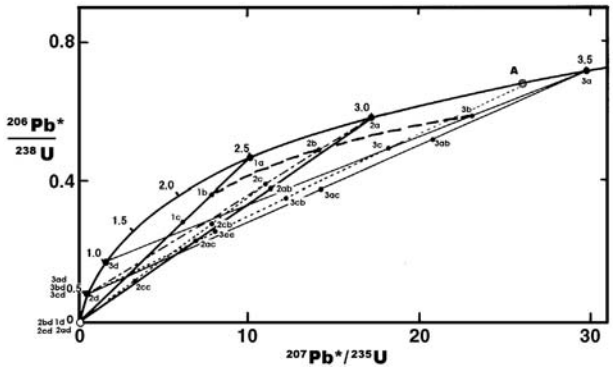


Fig. 5. U–Pb concordia-discordia ages. A U–Pb concordia-discordia diagram shows that two Pb-loss events on zircon grains initially formed 3.5 billion years (by) ago result in the scatter of analytical points of the zircons between the two discordias connecting 1.0–3.5 and 0.5–3.5 Ga. Dot and an open circle are the locations of zircons which have suffered various degrees of Pb-loss at various times. Straight lines are discordias of various ages. The double circle with a letter A is an example of a meaningless upper intercept age created by a group of zircons affected by both 1.0 and 0.5 Ga Pb-loss events. 0–1a–1b–2b–3b is a changing course of zircon compositions which suffered lead loss only during the ca 1 Ga (2.5 by after the original zircon formation) event. 0–2a–2ab–3ab is that suffered lead loss only during the ca. 0.5 Ga (3.0 by after the original zircon formation). Further explanation is given in text.

authors including Hölzl et al. (1994, Sri Lanka), Shiraishi et al. (1994, LHB granulites) and Dirks and Wilson (1995, Prydz Bay granulites) threw doubt on the existence of the Grenvillian events before the pervasive Pan-African lead-loss events.

However, in principle if the later Pb-loss event is strong, no lower intercept age of the earlier event would be expected. This is interpreted in Fig. 5 using the example of a zircon formed at 3500 Ma before the present that suffered Pb-loss events at ca. 1000 Ma and 500 Ma. If some zircons lost their Pb at 1000 Ma and were not affected by the ca. 500 Ma events, their U–Pb plots would be aligned on a discordia connecting 3500 Ma and 1000 Ma. In contrast, zircons which did not lose Pb at ca. 1000 Ma and instead lost their lead during the ca. 500 Ma events would yield U–Pb plots aligned along a discordia connecting 3500 Ma and 500 Ma. Zircons, which lost their lead during both the 1000 Ma and 500 Ma events, are scattered in the area between the two discordias above. Specifically they are aligned on a variety of discordias connecting 0.5 Ga and some meaningless upper intercept ages (for example, point A on Fig. 5) as explained below.

A metamorphic event at 1.0 Ga, i.e., 2.5 billion year (by) after the original zircon formation at 3.5 by ago resulted in a partial lead loss of a zircon, to shift its compositional location from 1a to 1b, 1c, 1d, etc. At 0.5 by after the above event, i.e., 3.0 by after the original zircon formation, 1b zircon gradually loses its lead and change composition along its concordia and comes to 2b, and 1c to 2c, and so on, these points forming a discordia connecting 3.0 Ga and 0.5 Ga. A second metamorphism at 3.0 by after the original zircon formation results in a lead loss on zircons and the composition of 2a zircon changes along the discordia connecting 2a and 0 point, e.g., 2ab, 2ac, 2ad. Other points, for an example, 2c zircon comes to 2cb point. At 3.5 by after the formation of the original zircon (at

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present time), all zircons lose their lead gradually along their concordia. For example, 2c zircon comes to 3c, while 2cb zircon comes to some location (3cb) between the discordias of 3.5–1.0 and 3.5–0.5 Ga, aligned on a discordia connecting 3c and 0.5 Ga, with a meaningless upper intercept age of A. Similarly, 2b zircon which partially loses lead and changes its composition to 2bb, comes to some other location between the two discordias of 1.0–3.5 Ga and 0.5–3.5 Ga and is aligned to show another discordia with a somewhat older upper intercept age than A.

Thus, the latest events can readily form the discordia and give the lower intercept age; in contrast, the earlier event is difficult to be detected by the discordia-concordia method when the later event resulted in extensive lead loss to zircons. We can estimate such conditions in a Pan-African Orogen superimposed over the Grenvillian Orogen in East Antarctica, because the later event affected the lead loss over almost all zircons producing the absence of concordant 1000 Ma zircon ages. This means that all zircons were affected by the later lead loss event, and simultaneously it means that no discordia connecting 3500 Ma and 1000 Ma will be observed. The above arguments show that the absence of the lower intercept age of 1000 Ma does not necessarily indicate the absence of a 1000 Ma lead loss event.

The above argument also draw our attention that in a rock, the more we can detect Pan-African ages, the less chance we may detect the Grenvillian ages; and this can also be the case for the concordant ages. Actually this is the case for the Pan-African orogens in East Antarctica.

5.5. Formative process of zircons in relation to the age of the zircon formation

Zircon forms mostly by magmatic processes, sometimes by the metamorphic processes, and also, by hydrothermal processes (e.g., Hartmann et al., 2000). In the interpretation of zircon ages, identification of their formative process is also important in addition to considerations of the conditions discussed above. With metamorphic zircons there is also the question of when the zircon formed during a metamorphic episode of say, at least ca. 100 million years of duration. Zircon cannot form without a supply of zirconium and once formed it cannot dissolve except under high temperature and/or fluid-rich conditions. Even though zirconium supply occurs, the formation of zircon preferentially takes place around an existing zircon grain. Degeling et al. (2001) point out that the break down of garnet leads to the formation of zircons and a supply of zirconium is liable to accompany this process and Kelley and Harley (2004) show good field examples from the Rayner Complex of East Antarctica. The decomposition of garnet is possible from a variety of metamorphic reactions. For example, garnet may decompose along the prograde path of the P–T–t curve of a metamorphism, first forming staurolite and then forming orthopyroxene; along the retrogressive path, it may decompose, first forming spinel+quartz, then biotite plus aluminosilicate, and later chloritoid plus biotite (Fig. 6). These events may be different in age by some tens of million years as shown in Fig. 6. It means that great caution is required when constraining the time of formation of zircons with the precision of a few tens of million years; it may not be possible without the examination of formative process of zircons. All the above arguments indicate that confidant explanation of a zircon age can only be obtained by careful analysis of the formative process of the analyzed zircons, including careful analyses of microtextures and chemistry of related minerals. Some tens of million years difference is not important in the differentiation of the Pan-African and Grenvillian ages, but is meaningful in identifying age difference within the CEAO. It is also very important for the timing of magnetization in palaeomagnetic studies, notably of Neoproterozoic East Gondwana.

Further, the lack of examination of the formative process of zircons in a geochronological study may sometimes leads to the overlooking of differentiation of the ages of superimposed events such as the Pan-African and Grenvillian.

5.6. Interpretation of zircon inner structures

There is a variety of interpretations of zoning patterns in zircons (e.g., Hartmann et al., 2000). From the point of view of the present study, the core-rim relationship is important (Fig. 7). The recognition that both the Prydz Bay and the LHB area seem to be devoid of Grenvillian metamorphism has been supported by the zircon micro domain and the Grenvillian age is considered to have an inherited detrital origin (i.e. the age is derived from exotic zircons and not the *in situ* zircons). However, it is difficult to identify from the core of the zircon whether it is exotic or *in situ*.

Hartmann et al. (2000) and Da Silva et al. (2000) show numerous examples of transformation of original Palaeoproterozoic zircons into Pan-African zircon either from the surrounding of the margin or from deposition in cracks, or even from the core and within the inner zones of a zircon grain. Wyne and Singha (1988) showed that deformed fragments of

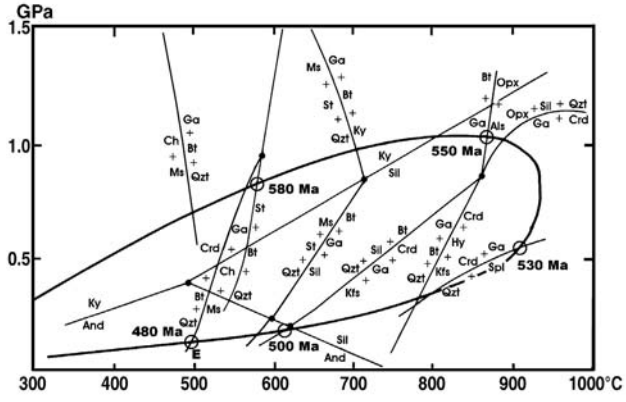


Fig. 6. P–T–t path and zircon formation. Possible formation of zircons associated with garnet break down plotted (as circles with ages) on a simplified petrogenetic grid. Notice that there are several possibilities for zircon formation under a prograde-retrograde P–T–t (pressure–temperature–time) metamorphic trajectory. Ages given are arbitrary and refer to several studies of metamorphic terrains. The base petrogenetic grid is referred after Shuto and Osanai (2002).

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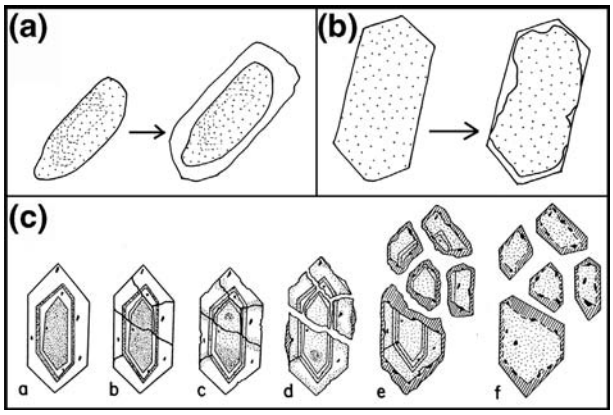


Fig. 7. Different interpretations of zircon core-rim relationships. (a) An inherited, detrital zircon of 2040 Ma rimed by 530 Ma zircon. The sample is taken from garnet-biotite gneiss from West Ongul Island, Antarctica (Redrawn after Shiraishi et al., 1994). (b) An original igneous zircon of ca. 1999 Ma partly altered to metamorphic zircon at ca. 516 Ma. The sample is taken from the Camboriu granitic gneiss, southeastern Brazil (Redrawn after Hartmann et al., 2000; Da Silva et al., 2000). (c) Igneous zircon of ca. 600 Ma broken and recrystallized by amphibolite facies mylonitization at ca. 460 Ma (summary figure). Samples were taken from mylonitized Henderson Gneiss of granite origin from North Carolina, USA (Reprinted from Contributions to Mineralogy and Petrology, vol. 98, 1988, Fig. 7a-f, page 118, Physical and chemical response of zircons to deformation, with kind permission of Springer Science and Business Media).

old zircons in mylonite were surrounded by new zircons and thus preserved as xenomorphic cores within an idiomorphic later overgrown rim. Thus, without careful examination of many examples of occurrence of zoned zircons in a rock sample as well as in many rock samples, identification of the exotic origin of the core part of a zircon is difficult. Zircon geochronology of both the LHB and Prydz Bay areas appears not to be associated with this kind of study, and is thus devoid of the necessary evidence to identify the true signature of their origin. Actually, Shiraishi et al. (1994) clearly stated that there are two explanations for the core zircons yielding Grenvillian ages, either they are relic *in situ* zircons or exotic detrital zircons; they arbitrarily selected the latter for their interpretation.

5.7. The total difficulty in identifying the absence of Grenvillian events from zircon geochronology

Recognition of the absence of a Grenvillian event from zircon geochronology alone is very difficult in Pan-African orogens where precursor Grenvillian metamorphism is suspected (Fig. 8). The reason for this is discussed below, in addition to the above-mentioned problems, and it is noted that the lack of detection of a 1000 Ma age does not point to the absence of 1000 Ma high-grade metamorphism.

There is no guarantee that a 1000 Ma zircon is large enough in grain size or thick enough in a layer of a zoned grain to be detectable by SHRIMP (~20 μm) or by EPMA (~2 μm). Further insufficient numbers or micro domains may have been sampled to make the detection of a 1000 Ma grain likely. Furthermore, even though there may be practically no 1000 Ma zircons, we cannot exclude the possibility that the 1000 Ma event did not precipitate zircon, or that once formed, zircons were completely dissolved and disappeared during a later event. It is important to realize that zircon grains within a rock have unequal histories and strained conditions, etc. (e.g., Kröner et al., 1994), and therefore only a lucky grain or a suitable small domain may be recrystallized on a specific occasion. Furthermore, the samples collected are negligibly small compared with the scale of the geology.

As a result of the above problems, the detection of only a very small number of a ca. 1000 Ma age dates should not be neglected; rather, we have to understand that there should be a great amount of back evidence in nature that have shown us one small tip of evidence. This is a very important point to be stressed to consider the Pan-African orogens in East Antarctica.

6. Conclusion

The Pan-African assembly hypothesis of East Antarctica proposed by Fitzsimons (2000) mostly from geochronologic data has lead on to a variety of studies proposing a Pan-African assembly of East Gondwana. These conclusions have been critically discussed from a geochronological point of view and the main conclusions are:

1. The Rayner Complex, i.e., central part of the Grenvillian Circum-East Antarctic Orogen (CEAO) records evidence of older ages than 990 Ma not only from Rb–Sr whole-rock isochron determinations but also from U–Pb zircon as well as Sm–Nd garnet ages. Apparent younger ages of the Rayner Complex pointed out by many authors (e.g., Fitzsimons, 2000) are considered to be the result of the predominant activity of the Mawson Charnockite formation and

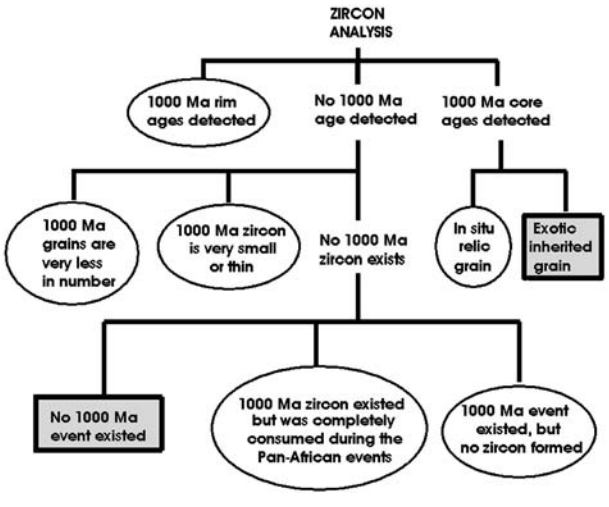


Fig. 8. The evidence required for a precursor event to be detected by zircon geochronology. The flow diagram shows that in a geochronological study of an area that has suffered strong metamorphism at ca. 500 Ma, confirmation of the absence of a possible ca. 1000 Ma event is very difficult. The examination follows a path in which only the two gray squares are cases where no 1000 Ma event occurred. All other 6 cases encircled by ovoids point to the existence of a ~1000 Ma Grenvillian event.

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equivalent igneous activity, and their rejuvenation effect. Thus, the Grenvillian orogen with a similar age range of ca. 1000 Ma to 1200 Ma is traceable from the Pinjarra Orogen (*sensu stricto*, Myers, 1993) in the western margin of Western Australia, Denman Glacier, Prydz Bay, Northern Prince Charles Mountains, Rayner Complex, Soya Coast, Lützow-Holm Bay, Eastern Dronning Maud Land, Central Dronning Maud Land to Western Dronning Maud Land. The Albany-Fraser Orogen is older than the other parts of the CEAO by several hundred million years, and may represent an older phase of the Circum-East Antarctic Orogeny.

In principle, there is no need for precise synchronism in the timing of the peak orogeny within such a large orogenic belt surrounding East Antarctica, since colliding blocks within the area surrounding East Antarctica need not necessarily belong to the same plate, but rather they could comprise several different plates with differing ages of juxtaposition.

2. Two Pan-African orogens do carry several Grenvillian ages as noted above and are characterized by polymetamorphic/deformational signatures noted elsewhere, attesting to the high possibility of Grenvillian events before the Pan-African orogenesis. Evidence regarded as supporting the absence of Grenvillian events in these orogens is not definitive and needs to be further examined. From the viewpoint of geochronology, it is almost impossible to verify the absence of Grenvillian events prior to Pan-African orogenesis.
3. Problems of detection of earlier events in high-grade polymetamorphic terrains by U–Pb zircon geochronology have been discussed in detail. Lowering of closure temperatures, uncertainty in the weighted mean method of U–Pb concordant ages, difficulty of identification of earlier events by the concordia-discordia method in a polymetamorphic terrain, lack of examinations in most studies of the formative processes of the zircons providing the U–Pb ages, problems of interpretation of zircon intragrain microstructures, and the small numbers of samples used for analysis compared with the large size of natural geologic bodies have all been highlighted. All these points support the idea that we cannot exclude the possibility of earlier events in polyorogenic terrains from geochronology only. Thus, the possibility of Grenvillian events within the Pan-African orogens in East Antarctica cannot be excluded from U–Pb zircon data alone.
4. The present study highlights some problems associated with the geochronological study of East Antarctic orogenic history, although it is beyond the scope of this paper to include a thorough discussion of this tectonics; specifically the evidence for Pan-African sutures is not addressed here. Detailed studies in other fields, especially including structural geology, petrology and palaeomagnetism, are all required to study East Antarctica and clarify its relationship to the remainder of East Gondwana, and specifically to provide reliable correlation throughout the CEAO.

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