Chapter 6

The Antarctic Continent in Gondwana: a perspective from the Ross Embayment and Potential Research Targets for Future Investigations

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6.1 Introduction

The geological record has, for a long time, demonstrated that Antarctica holds a key role in deep-time supercontinent reconstructions. More recently, because of the well-documented relevance of the polar regions' processes in influencing the global changes of both ocean circulation and climate patterns, Antarctica has increasingly shown to be of similar importance in the context of paleoenvironmental and paleoclimatic investigations, particularly in those focused on the Cenozoic greenhouse to icehouse evolution.

Geological data in the Ross Embayment, the large region between the Transantarctic Mountains and West Antarctica including the Ross Sea and Ross Ice shelf, have been collected since the very early phase of heroic exploration (Anderson, 1965). Samples of the fossilised Glossopteris flora were found in the last field camp of the last expedition of Captain Scott's party, on their way back to Ross Island. Du Toit's (1937) Gondwana reconstruction was proven on the basis of the results of these first expeditions. The Mesoproterozoic Rodinia supercontinent (Dalziel, 1991; Li et al., 2008; Merdith et al., 2017a,b; Moores, 1991; Wingate et al., 2002) provides another example of where fundamental pieces of evidence revealed the

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central position held by Antarctica with respect to the other continental blocks, even though the details of some of the proposed configurations differ from each other.

Several air images were collected during an early reconnaissance period in the 1930s−40s, and the first detailed mapping started in the late 1950s−60s. From the 1960s to the early years of the 21st century, geological expeditions have faced the challenging conditions of remote field work necessary to evaluate the geological record held within the Antarctic rocks. Progressively revealed by the increased research activities of national and multinational expeditions following the 1957−1958 International Geophysical Year, to the activities planned during the International Polar Year 2007/2008, the geological record of Antarctica has now acquired a valuable role in comprehending paleoenvironmental and paleoclimatic investigations, particularly those focused on the initiation of Cenozoic glaciations, the stability of the polar ice sheets, and the complex interactions among tectonic, sedimentary and climatic processes. The last century also included some challenging drilling projects both onshore or very close to the coast (e.g., DVDP, CIROS, CRP and ANDRILL) and offshore (e.g., DSDP leg 28; IODP Expedition 318, IODP Expedition 374) (see McKay et al., 2021, figure 5). The former were benchmarks for the first proximal glaciological reconstructions of the entire continent. The latter played a critical role in the first set up of modern paleoclimate research, providing the first record of Cenozoic oxygen isotope data in the Southern Ocean. In a mostly ice-covered continent as Antarctica, remarkable results on the bedrock geology were also obtained from geophysical (e.g., Golynsky et al., 2018 and ref. therein), geochronological and geochemical studies on rocks and mineral detrital fragments included in metasedimentary and sedimentary rocks of various ages in circum Antarctic outcrops (Cooper et al., 2011; Elliot, 2013; Elliot and Fanning, 2008; Elliot et al., 2015, 2016, 2017; Elsner et al., 2013; Estrada et al., 2016; Goodge et al., 2001, 2002, 2004a,b, 2010, 2012, 2017; Henjes-Kunst et al., 2004; Paulsen et al., 2015, 2016; Paulsen et al., 2017; Rocchi et al., 2015; Schulz and Schüssler, 2013; Stump et al., 2006; Veevers and Saeed, 2008, 2011, 2013; Wisoczanski and Allibone, 2004). Recent detailed reviews of the Antarctic geological features discussed here are reported in detail within Kleinschmidt (2021a).

As a result of many international collaborative projects, and several national programs supporting geoscientific research, the big picture of the Antarctic plate, its relations to the surrounding ones and its most prominent tectonic features, are increasingly becoming clearer. Knowledge of basement geology, and cover sedimentary rocks, determined by conventional and more recent analytical methods have opened up future field and laboratory-based investigations. With the switch in many areas from the initial phase of reconnaissance work to more detailed high-resolution mapping and sampling, geophysical surveys, the intense use of new facilities providing isotopic data and remote sensing data, the anatomy of the basement is now revealed as much more variegated and complex than expected from the preliminary data. Nevertheless, areas where there is insufficient information, or lack of appropriate level of resolution, still persist and require new basic research efforts before a robust reconstruction of the basement deformational and petrological evolution, as well of the sedimentary processes in spatially associated basin set-ups, can be considered as fully documented.

After a brief description of the present-day geodynamic setting and a summary of the tectonostratigraphic framework, we focus on three main topics: (1) the important phase of Proterozoic to Paleozoic crustal evolution, which marked the transition between the two supercontinents of Rodinia and Gondwana; (2) the complex interplay between tectonic, sedimentary and paleoclimatic processes recorded in the Paleozoic to Mesozoic cover rocks; and (3) the relationships between compressional and extensional events in West Antarctica during Phanerozoic time after the end of the Ross Orogeny, and the close links between onshore geological data and offshore (sedimentary cores) information to decipher the younger, Cenozoic to Holocene, events.

The chapter describes the evolution of the Antarctic continent from its inclusion as part of the Gondwana supercontinent to the breakup of this landmass and the repositioning of Antarctica at southern polar latitudes since the Early Cretaceous (c. 120 Ma). The chapter also highlights some of the most interesting paleoclimatic issues, which are considered essential to improve our understanding of the polar climate and ice ages and their influences on Earth's climate system in the Cenozoic to present time. Our intention is to give a general overview, which should be complementary to the more detailed information included in the accompanying chapters devoted to specific aspects of the geological record. The chapter's conclusions highlight some of the persisting open problems in Antarctica's tectonic evolution, and areas where major research themes are needed both on- and off-shore.

6.2 The Antarctic plate and the present-day geological setting of the Ross Embayment

The Antarctic continent comprises two primary tectonic regions: (1) East Antarctica and (2) West Antarctica, with the associated West Antarctic Rift System (WARS) (Fig. 6.1B). East Antarctica is thought to feature Precambrian continental crust generally c.35−45 km thick (Bentley, 1991), and up to c.58 km in the Gamburtsev Mountains (Ferraccioli et al. (2011) and ref. therein) that are stable, coherent and often topographically high (Cogley, 1984) material that held a central position in the Paleozoic supercontinent of Gondwana (Tingley, 1991), as it did in the Mesoproterozoic supercontinent Rodinia (Dalziel, 1991; Moores, 1991). In contrast, West Antarctica is an amalgamation of low-lying, 20−35 km thick, mainly younger crustal blocks (Dalziel and Elliot, 1982; Janowski and Drewry, 1981; Jordan et al., 2020).

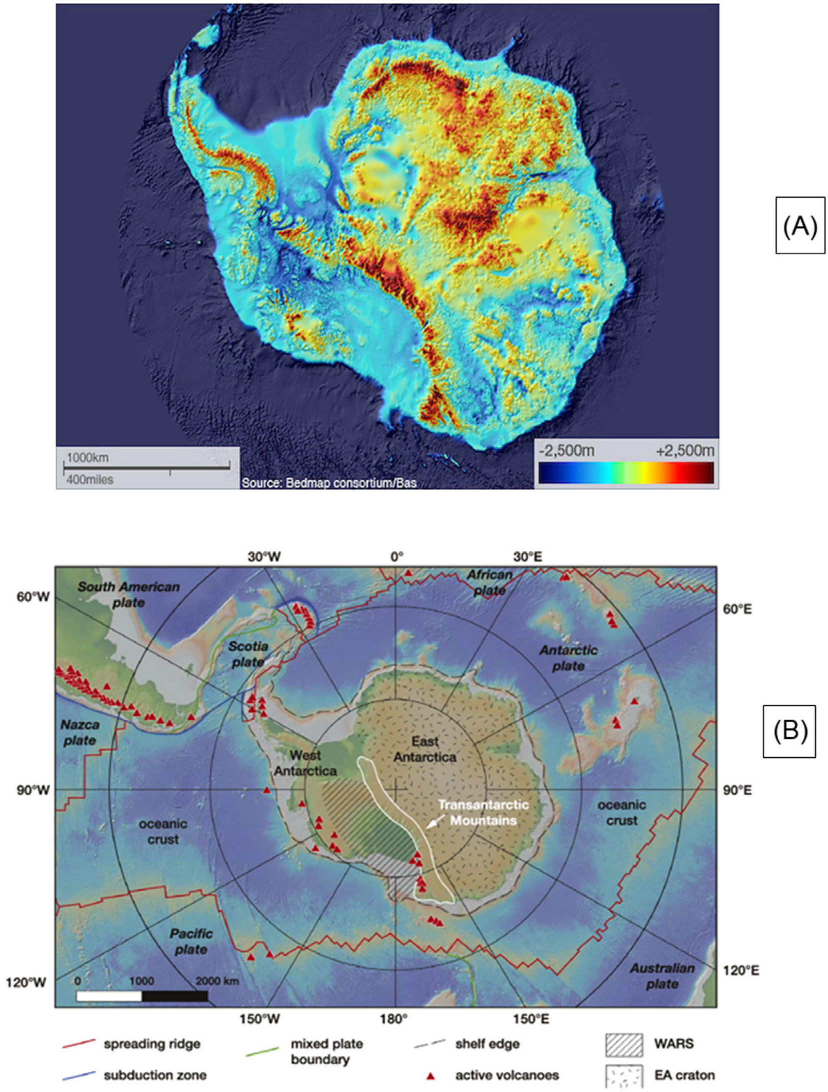


FIGURE 6.1 (A) Map showing the main Antarctic bed topography, based on BEDMAP 2 (Fretwell et al., 2013). (B) Schematic tectonic map of the Antarctic plate, showing bounding spreading ridges and subduction zones (after Goodge, 2020). Continental Antarctica extends to the shelf edge. Thick cratonic crust and lithosphere of East Antarctica (dashed pattern) and extensional West Antarctic rift system (WARS, diagonal ruling) are separated by the intraplate Transantarctic Mountains. Base from GeoMapApp (http://www.geomapapp.org).

The West Antarctic Rift lithosphere (WARS) compares well to other major Cenozoic continental rift systems (Behrendt, 1999). In the Ross Sea, the WARS borders the Transantarctic Mountains on their eastern side as a broad region of thinned continental crust associated with Cretaceous and episodic Cenozoic extension (Behrendt et al., 1991a,b; Behrendt, 1999). The WARS has high heat flow (83–126 mWm^2) (Berg et al., 1989; Blackman et al., 1987a,b) and thick (5–14 km) sedimentary basins with recent faulting (Cande and Leslie, 1986; Cooper et al., 1987; Hamilton et al., 2001). The Ross Archipelago (LeMasurier and Thomson, 1990) is currently active with fumarolic activity associated with alkaline volcanism at Mt. Erebus and Mt. Melbourne. The crust in the WARS is currently 20–25 km thick (Cooper et al., 1997; Jordan et al., 2020; Trehu, 1989).

The Transantarctic Mountains are approximately 2500 km long and 200 km wide, dividing East Antarctica from West Antarctica with peaks that rise over 4 km above sea level. Crustal thickness under the Transantarctic Mountains varies between 20 and 45 km (Bannister et al., 2003; Busetti et al., 1999; Cooper et al., 1997; Kanao et al., 2002; ten Brink et al., 1993, 1997). These mountains differ sharply from most mountain ranges of similar size and lateral extent because their formation did not reflect any compressional orogenic phases, but rather a process thought to have been directly-linked and causally-related to the development of the WARS (e.g., Studinger et al., 2004; ten Brink et al., 1997, and ref. therein). The Transantarctic Mountains were uplifted 6–10 km in an asymmetric tilt block formation and underwent denudation from the Cenozoic to the Cretaceous (Fitzgerald, 1992, 1995; Studinger et al., 2004). The later part of the uplift and denudation phases occurred under persisting spreading/extension within the western Ross Sea (Cande et al., 2000) and has been concomitant with voluminous sediment infilling from the Late Eocene/Oligocene to present (Barrett et al., 2000, 2001; Hamilton et al., 2001).

Detailed information about the different reconstructions and tectonic scenarios for Antarctica in Gondwana can be found in a range of previous work, including in Lawver et al. (1998), Larter et al. (2002), Boger and Miller (2004), Fitzsimons (2000a,b, 2003), Cawood (2005), Collins and Pisarevsky (2005), Payne et al. (2009), Boger (2011), and Merdith et al. (2017a,b).

Here, we will focus on the major geological constraints, key regions and datasets available in Antarctica, which represent a basis for reconstructing the southern continents in Gondwana. In this context, our review will also face some of the most controversial aspects that are presently under debate concerning the reconstruction of the main phases before and during the amalgamation of the Gondwana supercontinent. The geological evolution of the Antarctic continent is reviewed in two main periods: (1) before c. 450 Ma, covering the processes that were active during the amalgamation of Rodinia and Gondwana; and (2) c. 450 Ma to present day, including all the major events that occurred after the final stage of Gondwana amalgamation along its paleo-Pacific margin.

6.3 East Antarctica

6.3.1 The Main Geological Units during the Paleoproterozoic−Early Neoproterozoic Rodinia Assemblage

As with all other continents, Antarctica comprises a number of Archaean/Early Proterozoic cratons (older than c.1.5 Ga) surrounded by successively younger belts that formed, and/or accreted to the continental margins, as products of convergent plate tectonic events such as subduction of oceanic crust underneath continental crust and/or collision of two former separated continents (Fig. 6.2). The Precambrian geological evolution of the East Antarctica is still highly debated, but some key points are now well defined by a large number of recent papers (see Aitken et al., 2016; Bauer et al., 2003a,b, 2009; Bogdanova et al., 2009; Boger, 2011; Boger et al., 2015; Cawood and Buchan, 2007; Collins and Pisarevsky, 2005; Daczko et al., 2018; Evans, 2009; Fitzsimons, 2000a,b, 2003; Godard and Palmeri, 2013; Golynsky and Jacobs, 2001; Golynsky et al., 2018; Goodge, 2002, 2020, 2021; Goodge and Severinghaus, 2016; Goodge et al., 2002, 2008, 2010, 2012, 2017; Harley et al., 2013; Jacobs, 2009; Jacobs and Lisker, 1999; Jacobs and Thomas, 2002, 2004; Jacobs et al., 1998, 2003a,b, 2008a,b, 2015, 2017, 2020; Kleinschmidt, 2021a,b; Kleinschmidt and Boger, 2008; Läufer, 2021; Läufer et al., 2021; Li et al., 1995, 2008; Liu et al., 2006, 2007, 2009, 2013, 2014, 2016, 2017, 2018; Meert and Torsvik, 2003; Ménot, 2021; Merdith et al., 2017a,b; Mieth and Jokat, 2014; Mikhalsky and Kamenev, 2013; Mikhalsky et al., 2013, 2015; Morrissey et al., 2017; Osanai et al., 2013; Paech et al., 2005; Palmeri et al., 2018; Payne et al., 2009; Pisarevsky et al., 2003; Riley et al., 2020; Roland, 2021; Ruppel et al., 2015, 2018, 2020; Siddoway, 2021; Smellie, 2021; Thomson, 2021; Tucker et al., 2017; Veevers et al., 2016; Wang et al., 2020; Yoshida et al., 2003). In Fig. 6.2 a schematic highly-speculative geological map is proposed with the hypothesised sub-ice extent of the main crustal domains. Archaean-Proterozoic Cratons are confined to East Antarctica and several are generally accepted as follows:

- the small 'Grunehogna Craton' (c. 3.1 Ga basement − Marschall et al., 2010, 2013) covered by flat-lying and undeformed sediments older than 1 Ga, interpreted as a fragment of the African 'Proto-Kalahari Craton';
- the 'Napier Craton' (considered a fragment of the 'Dwarhai Craton' in India) (Kelly and Harley, 2005);
- the 'Mawson Craton' (Fitzsimons, 2000a,b, 2003; Naumenko-Dèzez et al., 2020 and ref. therein) exposed only in limited areas near the continental margin in Terre Adélie and in George V Land, and their Australian counterpart the 'Gawler Craton', extending into the Miller and Geologists Ranges of the central Transantarctic Mountains;
- the 'Ruker Craton', probably extending to south-west and the cryptic 'Valkyrie Craton' (part of the 'Crohn Craton') (Boger, 2011; Boger et al., 2006);

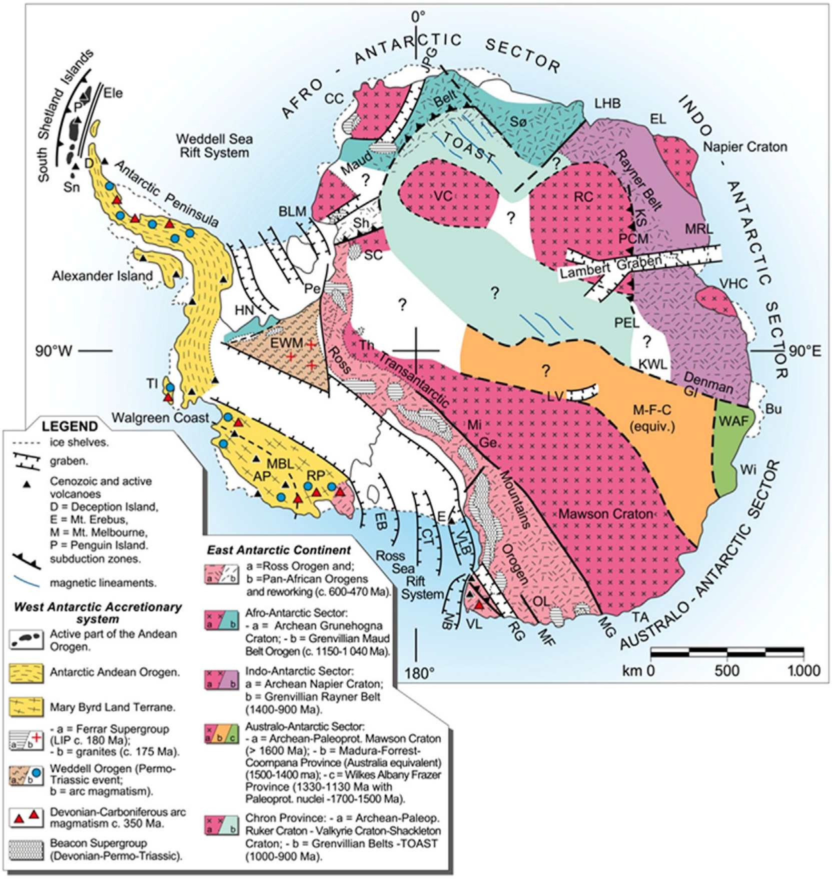


FIGURE 6.2 Schematic geological map of Antarctica (modified after Kleinschmidt, 2014). The map shows the possible sub-ice extent of the cratonic areas, the four 'Grenvillian' belts, the Pan-African orogenic belts and the Ross Orogen. The map also shows the distribution of Beacon Supergroup and Ferrar Supergroup outcrops and of the two main Mesozoic and Cenozoic orogens (Ellsworth or Weddell Orogeny and Antarctic Andean Orogen) which formed along the paleo-Pacific margin of Gondwana in West Antarctica, and the major intra-plate fracture zones in East Antarctica. Abbreviations: *AP*, Amundsen Province; *BLM*, Bertrab, Littlewood, Moltke Nunataks; *Bu*, Bunger Hills; *CT*, Central Trough; *EB*, Eastern Basin; *EL*, Enderby Land; *Ele*, Elephant Island; *EWM*, Ellsworth-Whitmore Mountains; *GC*, Grunehogna Craton; *Ge*, Geologists Range; *HN*, Haag Nunatak; *JPG*, Jutul Penck Graben; *KL*, Kemp Land; *KWL*, Kaiser-Wilhelm-II-Land; *KS*, Kuunga Suture; *LHB*, Lützow-Holmbukta; *LV*, Lake Vostok; *MBL*, Marie Byrd Land; *MF*, Matusevich strike−slip fault; *MG*, Metz Glacier; *Mi*, Miller Range; *MRL*, Mac.Robertson Land; *NB*, Northern Basin; *OL*, Oates Land; PCM, Prince Charles Mts.; *Pe*, Pensacola Mts.; *PEL*, Princess Elizabeth Land; *RG*, Rennick Graben; *RC*, Ruker Craton; *RP*, Ross Province; *SC*, Shackleton Cratonic domain; *Sh*, Shackleton Range; *Sn*, Snow Island; *Sø*, Sør Rondane; *TA*, Terre Adélie; *Th*, Thiel Mts.; *TI*, Thurston Island; *TOAST*, Tonian Oceanic Arc Super Terrane; *VC*, Valkyrie Craton; *VHC*, Vestfold Hills Craton; *VL*, Victoria Land; *VLB*, Victoria Land Basin; *WAF*, Wilkes-Albany-Frazer Orogen; *Wi*, Windmill Islands. The Precambrian to Cambrian geological structure of East Antarctica, which is ice covered for the most part, is still highly speculative.

- the southern Shackleton Range (Riley et al., 2020; Will et al., 2009 and ref. therein) and perhaps the eastern Thiel Mountains (Ford, 1963) (part of the 'Mawson Craton'?);
- the 'Vestfold Hills' cratonic fragment in Prydz Bay region (Clark et al., 2012; Zulbati and Harley, 2007); and
- the BLM cratonic block in the Coats Land, which is an enigmatic crustal domain, hypothesised to be of pre-Grenvillian age (Kleinschmidt and Boger, 2008) or alleged Paleoproterozoic age (Jacobs et al., 2015), where only an undeformed c. 1100 Ma rhyolites and granophyres cover emerges from the ice. In a recent paper, Riley et al. (2020) dated a granite pegmatite cobble ice-transported to and recovered from the Brunt Ice Shelf: its U-Pb zircon age (c. 1100 Ma) is similar to that of the felsic volcanic cover, but also to the widespread Grenvillian age of the Maud Belt.

The exposed basement of these cratonic blocks, whose internal evolutionary history is not detailed here, only occasionally covered by almost flat-lying platform sediments (e.g., Thiel Mountains), consists of high-grade gneisses and granulite facies metamorphic complexes showing radiometric ages older than c. 1500 Ma up to just over 3800 Ma. Since the relationships between the different cratons are still poorly known, a tectonic reconstruction is mainly speculative. How and when did these cratonic nuclei assemble together? Until about the late 2010s the prevailing hypothesis was that in a long-lasting period, from the Mesoproterozoic to early Neoproterozoic (c. 1300−800 Ma), their docking was related to the Grenvillian age orogens well known in the Laurentian paleoplate (e.g., Bogdanova et al., 2009; Li et al., 2008; Meert and Torsvik, 2003 and ref. therein).

Along the margins of these cratons, large accretionary belts are found. These orogenic belts, separating the cratons as internal belts in East Antarctica, or as elongated peripheral belts at the paleo-Pacific margin of Gondwana, developed during two major orogenic cycles spanning in time from c.1.3−0.9 Ga (Grenvillian-aged orogens), through 500−600 Ma (Ross and the Pan-African orogens). As in eastern North America and other continents, the Grenvillian-aged orogens in Antarctica are considered to record the fundamental orogenic event marking the amalgamation of the Rodinia supercontinent in Meso-Neoproterozoic times. Until the early 1990s, only one very long Antarctic Grenvillian orogen was assumed, following the Antarctic coast as a 250 km wide strip from Coats Land in the west up to George V Land in the east, occasionally specially named the 'Circum East Antarctic Mobile Belt' (Yoshida, 1992). Such an extension has been demonstrated to be incorrect for Terre Adélie and George V Land and unproven for Coats Land, where the tiny outcrops represented by three little groups of nunataks (Littlewood, Bertrab and Moltke) consist of c.1100 Ma rhyolites and granophyres that are absolutely undeformed (Kleinschmidt and Boger, 2008; Storey et al., 1994). The existence of a Circum East Antarctic Mobile

Belt is no more supported by new geological evidence, which instead indicates the presence of at least four distinct Grenvillian-aged domains (Fitzsimons, 2003; Jacobs et al., 2015, 2020; Ruppel et al., 2015, 2020 and ref. therein) (Fig. 6.2) described as follows:

1. the ‘Maud Belt’, extended from western to eastern Dronning Maud Land to parts of Sør Rondane, interpreted as the trace of an active Grenvillian continental margin at the ‘proto-Kalahari Craton’ (Jacobs et al., 2008a,b, 2015, 2020 and ref. therein);
2. the ‘Rayner Belt’ (Yoshida and Kizaki, 1983) that in Enderby, Kemp and MacRobertson Lands separates the ‘Napier Craton’ from the southern Prince Charles Mountains cratonic area (‘Ruker Craton’) and which was verified in the northern Prince Charles Mountains (e.g., Boger et al., 2002; Morrissey et al., 2016);
3. the ‘Wilkes Province Belt’ (Fitzsimons, 2000a; the ‘Wilkes-Albany-Frazer Orogen’ in Fig. 6.2) of Wilkes Land that is exposed in the Bunger Hills and Windmill Islands, and merges and divides the Vestfold Hills domain from the Mawson continent; and
4. the ‘TOAST’ (‘Tonian Oceanic Antarctic Super Terrane’) domain interpreted as an oceanic arc belt of Tonian age (Jacobs et al., 2008a,b, 2015, 2020).

At the same time, some detailed research proposed that these Grenvillian belts are active continental arcs accreted to the proto-Kalahari and proto-Indian cratons, far from the Mawson-Gauler cratons (Jacobs, 2009; Jacobs et al., 1998, 2003a,b, 2008a,b; Liu et al., 2009; Mikhalsky et al., 2009); a proposal now widely accepted (see Boger, 2011; Jacobs et al., 2020; Merdith et al., 2017a,b and ref. therein). Below a short summary is given of these events, subdivided in the four different crustal domains according the proposed paleoplate affinity (Fig. 6.2).

1. ***Afro-Antarctic Sector***. In this wide region the basement is composed by high-grade ortho and paragneisses (the Grenvillian ‘Maud Belt’) dated at 1170 to 1050 Ma followed by high-grade metamorphism at c. 1080–1030 Ma as part of an active continental margin on the Kalahari craton at that time (Arndt et al., 1991; Bauer et al., 2003a,b; Board et al., 2005; Grantham et al., 2013; Groenewald et al., 1995; Jacobs, 2009; Jacobs et al., 1998, 2003a,b, 2008b, 2009, 2015, 2020; Kamei et al., 2013; Paulsson and Austrheim, 2003; Ruppel et al., 2020; Wang et al., 2020; Will et al., 2009). The sector extends from the Coats Land, through the Dronning Maud Land to the Sør Rondane region, north of the ‘Forster Magnetic Anomaly’. Evidence of a late stage (c.785-760 Ma) magmatic event in the Schirmacher Oasis suggests a reactivation of this Grenvillian continental margin (Jacobs et al., 2020).
2. ***Indo-Antarctic Sector***. Here, covering circum-Antarctica coastal regions from Lützow-Holmbukta Bay to the Denman Glacier, the wide and composite ‘Rayner Belt’ (or ‘Rayner Complex’) is characterised by magmatic and metamorphic events dated c. 1400–800 Ma (Kelly and Harley, 2004;

Kelly et al., 2002; Kelsey et al., 2007; Liu et al., 2009, 2016, 2017, 2018; Mikhalsky and Leitchenkov, 2018; Mikhalsky et al., 2009, 2013, 2015; Phillips et al., 2006) testifying a long-lived Grenvillian active continental arc. Magmatic rocks of 'charnockite type' and granulites are common. The composite character of this orogenic belt is strengthened by the presence of crustal domains with Paleoproterozoic protoliths ages of felsic to mafic orthogneisses (e.g., in the Lützow-Holm Bay area, Dunkley et al., 2020; Takahashi et al., 2018 and ref. therein). The western boundary with the 'Maud Belt' in the Lützow-Holmbukta region is still highly debated and inferred mainly from geophysical data (see Ruppel et al., 2018). In the oceanward side two Archaean cratonic fragments are present: the Napier Craton (Kròl et al., 2020 and ref. therein), and the Vestfold Craton with Indian affinity (Clark et al., 2012; Zulbati et al., 2007).

3. ***Australo-Antarctic Sector***. Contrary to the sectors described above, the Australo-Antarctic sector, extending between the Denman Glacier region to the northern coast of northern Victoria Land, was in Precambrian-Paleozoic time always welded to the Australian units. Along the coastal rocks exposure a long-lasting research activity defined different lithotectonic units from the ('Lachlan Orogen' equivalent) Robertson Bay Terrane, to the Bowers Terrane and the Wilson Mobile Belt ('Ross Orogen'; Tessensohn and Henjes-Kunst, 2005 and ref. therein), to the Mertz Glacier Shear Zone where the 'Mawson Craton' is transected (Di Vincenzo et al., 2007; Lamarque et al., 2018; Naumenko-Dèzes et al., 2020; Talarico and Kleinschmidt, 2003 and ref. therein). To the west of the 'Mawson Craton', according to Payne et al. (2009), Morrissey et al. (2017), and Liu et al. (2018), a younger, poorly exposed terrane remains hypothetical: the Australian 'Madura-Forrest-Coompana Province' (M-F-C) equivalent province (Fig. 6.3). This domain is proposed also on the basis of a magnetic grain different from that of the Mawson Craton (Golynsky et al., 2018) and as a source area for the glacially-derived granitoid clasts dated c. 1500−1400 Ma by Goodge et al. (2008, 2010, 2017). The Bunger Hills and Windmill Islands regions are located and interpreted as the north-western border zone of the Mawson Continent correlable to the Grenvillian 'Albany-Frazer Orogen' in Australia developed at the margin of the Yilgarn West Australian Craton (Liu et al., 2018; Morrissey et al., 2017; Tucker et al., 2020). The paleo-plate boundary between the Australo-Antarctic sector and the Indo-Antarctic sector is not exposed. Its position is generally inferred to be located within the wide (c. 250 km) region between Mirny Station and the Denman Glaciers (Aitken et al., 2014, 2016; Daczko et al., 2018; Fitzsimons, 2000a,b, 2003; Tucker et al., 2017, 2020). In Fig. 6.2 this boundary is located along Mirny Fault according to the proposal of Daczko et al. (2018). This is a key region (see discussion in Tucker et al., 2020) where both the 'Kuunga Orogen' and the 'Pinjarra Orogen' overlapped over a Paleo-Meso-Proterozoic basement (Daczko et al., 2018; Mikhalsky et al., 2015).

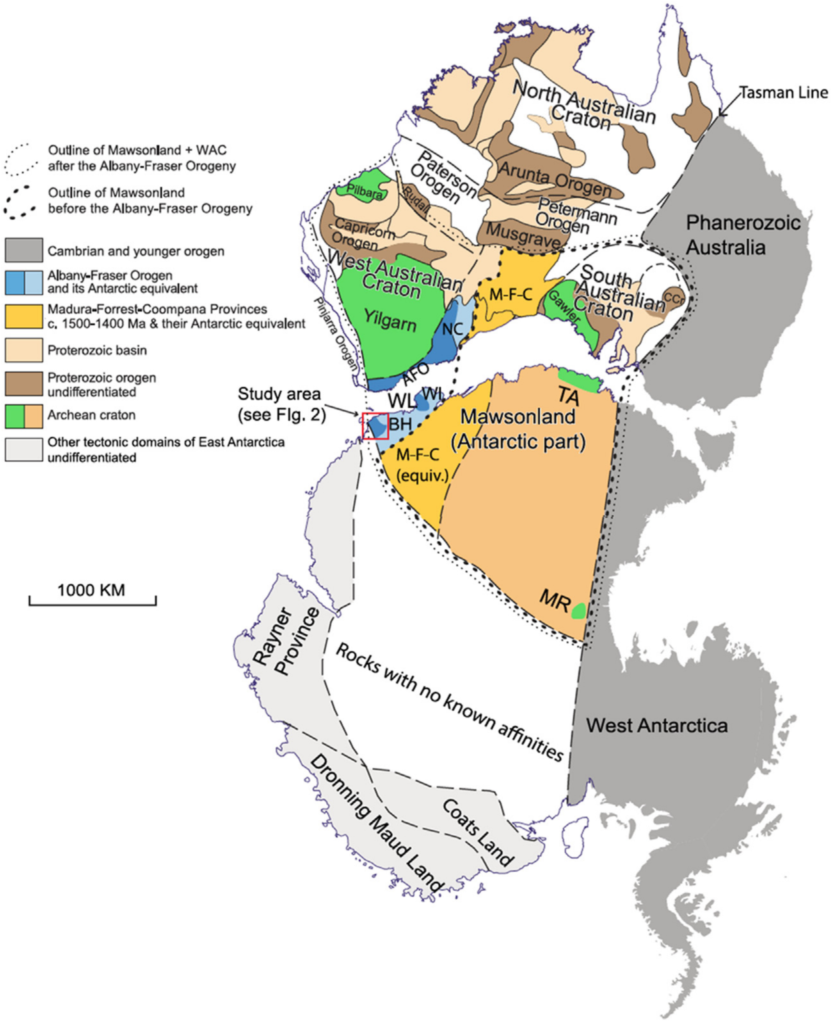


FIGURE 6.3 Tectonic map of Australia and Antarctica in a Gondwana configuration (after Liu et al., 2018). Abbreviations: *AFO*, Albany-Fraser Orogen; *BH*, Bunger Hills; *CCr*, Curnamona Craton; *M-F-C*, Madura-Forrest-Coompana Provinces; *Mr*, Miller Range; *NC*, Nornalup Complex; *TA*, Terre Adélie craton; *WI*, Windmill Islands; *WL*, Wilkes Land.

In the Wilkes−Albany-Frazer Orogen, in spite of the mainly ice covered field situation, detailed geochronological and petrological studies on Obruchev Hills, Bunger Hills and Windmill Islands (Morrissey et al., 2017; Tucker et al., 2017; Tucker et al., 2020; Zhang et al., 2012 and ref. therein), aided by aerogeophysical surveys on the subglacial tectonic structure (Aitken et al., 2014, 2016; Maritati et al., 2016), has led to better knowledge of the geological evolution of this province. Evidence for felsic to mafic magmatism and high-grade metamorphism are widespread and dated between c.1345 and c.1130 Ma, coeval with Stages I and II of the Albany-Frazer Orogen in Australia (Spaggiari et al., 2015). For some lithological domains an Archaean to Paleoproterozoic protoliths age is proposed (Black et al., 1992; Sheraton et al., 1992, 1993; Tucker et al., 2017).

4. **The '*Crohn Province*'**. This wide domain of East Antarctica, mainly covered by ice, matches what Boger (2011) defined the 'Crohn Craton'. Due to the extremely poor conditions of exposure, all statements about this province are highly speculative. Available field and geophysical data suggest that it is a composite crustal block made by cratonic nuclei (the 'Ruker Craton' and the cryptic 'Valkyrie Craton') and enveloping mobile belts at least in part of Grenvillian age (c.1000−900 Ma) as proposed for the TOAST in SE Dronning Maud Land (Elburg et al., 2014, 2016; Jacobs et al., 2015, 2018; Kamei et al., 2013; Ruppel et al., 2018, 2020 and ref. therein). These age data (SHRIMP or LA-MC-ICPMS on zircon) are mainly obtained on a gabbro-tonalite-trondhjemite-granodiorite (meta)intrusive suite (GTTS, Kamei et al., 2013) cropping out in the Sør Rondane Mountains region (a Pan-African sliver of a Tonian juvenile island arc according to Ruppel et al., 2020), and from samples selected from moraine deposits in the SW Terrane of this region (Jacobs et al., 2017). A characteristic structural feature in some areas of these belts are an alternating parallel, elongated, magnetic anomalies trending NW-SE (Mieth and Jokat, 2014; Mieth et al., 2014; Ruppel et al., 2018). According to Ruppel et al. (2020) the TOAST first accreted to the cryptic Valkyrie Craton in early Neoproterozoic times, then during the Ediacaran-Early Cambrian was sandwiched against the Grenvillian Maud Belt. Geophysical data provide the best constraints on subglacial geological structures but there is little information on their age. Fitzsimons (2003) suggested three possible paths for extension in East Antarctica of the composite Pinijarra Orogen (assuming a Pan-African age for these tectonothermal events) and one of these, the PD3 path, transects East Antarctica along the boundary between the Crohn Province and the Mawson Continent including the Gamburtsev Mountains subglacial region. There are no robust data supporting this hypothesis, however. Instead, as the Ruker domain has Pb-isotopic composition similar to basement terranes in Indo-Antarctica (Flowerdew et al., 2013) a possible alternative model is that proposed by Boger (2011): first the composite oceanic Crohn Province accreted in Grenvillian time to the Mawson Continent, then the Indo-Antarctic Sector collided along the 'Kuunga Suture' during the Pan-African events.

6.3.2 From Rodinia breakup to Gondwana (c. 800−650 Ma)

The amalgamation phase of Gondwana necessarily involved the aggregation of various continental fragments, which derived from the fragmentation and dispersal of a former supercontinent – variously named Ur-Gondwana (Hartnady, 1991), Katania (Young, 1995), Paleopangea (Piper, 2000) or Rodinia (McMenamin and McMenamin, 1990). However, the precise configuration and modality of breakup of this supercontinent are yet not completely known and these uncertainties obviously propagate to the formulation of tectonic models for the constructive phase of Gondwana. Since the focus of this paper is on the Cretaceous–Cenozoic record, we will briefly summarise the main recent results in the reconstruction of the Late Precambrian–Early Paleozoic tectonic evolution of Antarctica in Gondwana avoiding a detailed review of Rodinia models: SWEAT (Dalziel, 1991; Hoffman, 1991; Moores, 1991); AUSWUS (Karlstrom et al., 1999); AUSMEX (Wingate et al., 2002); and reviews by Dalziel (1997), Meert and Torsvik (2003), Evans (2013), Merdith et al. (2017a,b) and alternative models (e.g., Paleopangea; Piper, 1982, 2000). As recently summarised by Merdith et al. (2017a), and shown in Fig. 6.4, in the SWEAT fit, eastern Antarctica is pushed against the southwest United States, while Australia lies further north near the United States-Canadian border (Dalziel, 1991; Hoffman, 1991; Moores, 1991; Wingate et al., 2002); in the AUSWUS fit, eastern Australia is matched against the southwest United States of America (Karlstrom et al., 1999); in the AUSMEX, Australia has only a small connection with Laurentia, where the north tip of Queensland fits against Mexico (Wingate et al., 2002), Kalahari Craton is shifted further south to accommodate Mawson; in the 'Missing-Link' model (Li et al., 2008), South China is considered as a continental sliver between Australia and Laurentia.

In spite of significant uncertainties about the precise reconstruction of the global paleogeography in Neoproterozoic time, and the still incomplete geochronological framework, most authors agree that the breakup of Rodinia led to the development of extensive passive continental margins which are documented in the late Neoproterozoic (c. 750–600 Ma) record of most present-day continents (Cawood, 2005; Dalziel, 1991, 1992, 1997; Goodge, 2020; Goodge et al., 2002, 2004a,b; Meert and Van der Voo, 1997; Powell et al., 1993). In Antarctica, following Dalziel's hypothesis (Dalziel, 1991), the first stage of Rodinia breakup involved a rifting phase which started at c. 800–750 Ma leading to the separation of Laurentia from the East Antarctica-Australia block and the formation of the intervening proto-Pacific Ocean. This process was accompanied by the drift of many cratonic blocks, presently exposed in South America and Africa (Amazonian, Rio de la Plata, Western African and India + South China cratons) (i.e. West Gondwana). While Laurentia remained thereafter always separated from Mawson-Australian cratons, the Kalahari and India + South China moved against proto-East Antarctica and collided with it. Most workers agree that key evidence of this evolution is stored in the 'Mozambique Belt' (Holmes, 1951), 'East African Orogen' (Stern, 1994), or 'East African Antarctica Orogen' (Jacobs et al., 2003a, 2015), and that

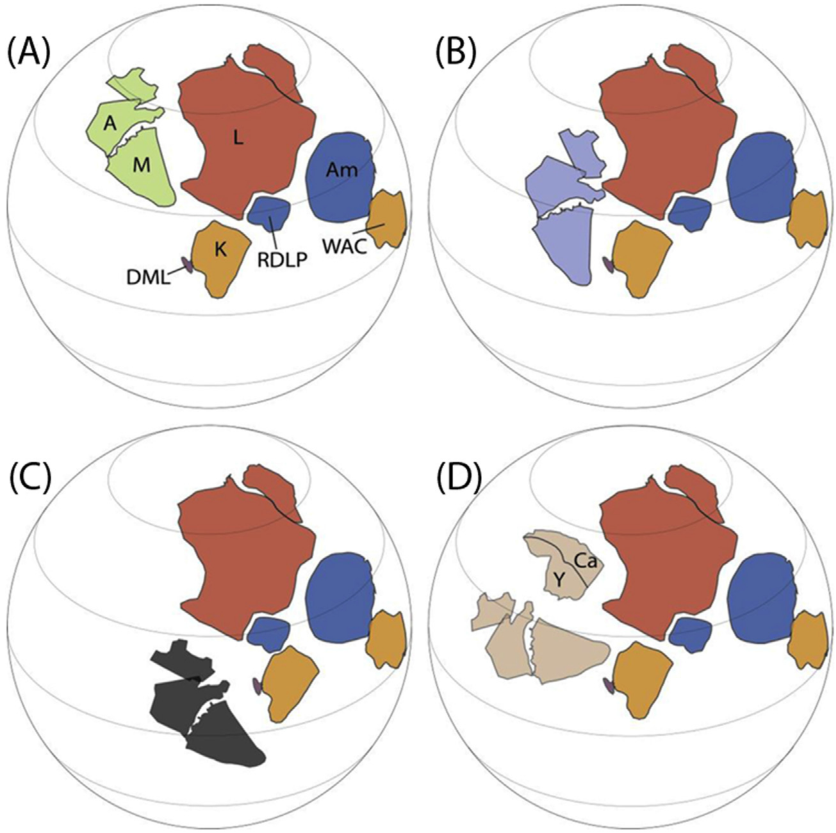


FIGURE 6.4 The different configurations of Laurentia with Australia-East Antarctica and South China with pre-Neoproterozoic geology overlain. Laurentia is fixed in its present-day position with all other blocks rotated relative to it at 800 Ma (from Merdith et al., 2017a). (A) SWEAT fit (green); (B) AUSWUS fit (blue); (C) AUSMEX (black); (D) Missing-Link model (tan). *A*, Australia; *Am*, Amazonia; *Ca*, Cathaysia (part of South China); *DML*, Dronning Maud Land; *K*, Kalahari; *L*, Laurentia; *M*, Mawson (East Antarctica); *RDLP*, Rio de la Plata; *WAC*, West Africa Craton; *Y*, Yangtze (part of South China). The colour of the other blocks represents their present-day geographical position. South America – dark blue; Africa – orange; Antarctica – purple; North America – red; China – yellow.

this extensive orogenic belt formed as a result of the closure of a 'Mozambique Ocean' and subsequent collision and amalgamation of East and West Gondwana during the Pan-African events (Dalziel, 1992; Jacobs et al., 2015; Shackleton, 1996; Stern, 1994 and ref. therein).

The Neoproterozoic-Cambrian process of amalgamation in Antarctica involved the following three regions: (1) the Afro-Antarctic Sector; (2) the Indo-Antarctica Sector; and (3) Wilkes Land in the Australo-Antarctic Sector. All are currently the subject of much debate.

1. **Afro-Antarctic Sector**. A review of structural and geochronological data from Dronning Maud Land-Lützow Holm Bay region and comparison with the adjacent (in Gondwana) Falkland Microplate and south-eastern Africa, led Jacobs and Thomas (2002) to corroborate the proposal by Jacobs et al. (1998) of a southward continuation of the Mozambique Belt into Dronning Maud Land in Antarctica. The Grenvillian 'Maud Belt' suffered a pervasive high-grade tectono-thermal overprinting and reworking at c. 650−500 Ma testifying the Pan-African collision of the West Gondwana with the East Antarctic proto-continent. The Pan-African events are characterised by strong variable metamorphic events up to the granulitic grade. In the Northern Shackleton Range only Pan-African ages have been detected: here the metamorphic basement is characterised by extensive thrust and nappe tectonics, widespread and distinct late orogenic collapse structures, thick molasse formations ('Blaiklock Glacier Group') (see Kleinschmidt, 2021a, 2021b). In most of Dronning Maud Land, and the eastern Sør Rondane region (Ruppel et al., 2020), metamorphic reworking up to HP-HT granulite grade is pervasive and also syn- and post-orogenic magmatism (Baba et al., 2010, 2013; Board et al., 2005; Elburg et al., 2016; Grantham et al., 2013; Jacobs et al., 2008a; Kleinhans et al., 2013; Osanai et al., 2013; Paech, 2004; Pauly et al., 2016; Roland, 2004; Roland and Olesch, 2004; Ruppel et al., 2020; Shiraishi et al., 2008). Many alternative locations have been proposed for the Pan-African suture in the Dronning Maud Land (e.g., figure 1 in Mieth and Jokat, 2014, and ref. therein). If the TOAST region is unbundled from the Afro-Antarctic Grenvillian Belt, a plausible suture-boundary between the Grenvillian Afro-Antarctic sector and the 'Crohn Province' is represented by the Forster Magnetic Anomaly (Riedel et al., 2013). A Pan-African suture zone in the Shackleton Range was suggested by Grunow et al. (1996) and conclusive evidence was found by Talarico et al. (1999) who described relics of ophiolites consisting of serpentinites and amphibolites with N-type MORB to OIB geochemistry and a maximum Sm−Nd age of c. 900 Ma. Metamorphic reworking of these rocks occurred under variable high P to medium P conditions in the eastern Shackleton Range; a stage of eclogite-facies metamorphism at c. 510 Ma has been recently reported by Schmädicke and Will (2006) in the central Shackleton Range. A detailed review of the Shackleton Range geology was recently published by Kleinschmidt (2021a, 2021b). On the basis of these discoveries, and thrust patterns in both the Shackleton Range and Western Dronning Maud Land, Kleinschmidt et al. (2002) proposed that the ophiolites may have formed part of a connection between the Paleo-Pacific and the Mozambique oceans, in the way the Drake Passage is linking the present Pacific and the Atlantic Oceans. The Mozambique Belt or EAAO ('East African Antarctic Orogen', as renamed by Jacobs and Thomas, 2002) and its continuation in Antarctica show a general N−S trending (Jacobs and Thomas, 2002).

The 'Maud Belt' belt has therefore provided sound evidence in contrast to the classical assumption that the whole Eastern Gondwana formed during the consolidation of Rodinia in the Mesoproterozoic time and supported the proposal that the 'Maud Belt' collided only in Neoproterozoic time with the 'Crohn Provence'. It then remained tectonically stable until the modern continents rifted from Gondwana in the Mesozoic. The Lützow Holm–Prydz Bay–Pan-African structures were initially considered to be part of a wider Pan-African belt termed the 'Kuunga Orogen' by Meert et al. (1995) or 'Kuunga Suture' by Boger and Miller (2004), and interpreted as the result of the collision between East Antarctica (+Australia) and India (+Madagascar + Sri Lanka) at c. 535–520 Ma, after the amalgamation of India with the rest of Gondwana along the Mozambique suture (Fig. 6.5). More recent data suggest that there are actually three orogenic belts involved in the orogenic collisions at about the same Pan-African period of c. 650–500 Ma in East Antarctica, highlighted by a often high-grade metamorphic reworking and intrusive bodies of Pan-African age (Fig. 6.2): (1) the belt in the Shackleton Range–Dronning Maud Land–southern Sør Rondane–Lützow-Holmbukta region ('East Antarctic Orogen' or 'East Antarctic Belt' of Jacobs et al., 1998, renamed 'Lützow Holm Belt' by Fitzsimons, 2000b); (2) a wide elongated portion of the Grenvillian 'Rayner Belt' from the Enderby Land to the Wilhelm II Land through the southern Prince Charles Mountains–Grove Mountains ('Kuunga Suture' according to Boger et al., 2002); and (3) the belt in the Denman Glacier region, interpreted as prolongation of the Leeuwin Complex of Australia's Pinjarra Orogen (e.g., Fitzsimons, 2000b). However, the exact extent of these orogens is doubtful because of the extensive ice cover, and this applies especially to the Denman Glacier belt.

As proposed by Boger et al. (2001, 2002), Gondwana's amalgamation in Antarctica may have taken place in two steps: the first before 550 Ma and the second after 550 Ma. The first step involved the amalgamation of West Gondwana with 'Indo-Antarctica' (i.e. India and the 'Rayner Belt' – 'Napier Craton') documented in Dronning Maud Land. The second step led to the aggregation of these terranes to the rest of East Gondwana, i.e. the rest of East Antarctica thus producing the 'Kuunga Suture'. Geophysical data from Mieth and Jokat (2014) suggest that the Lützow Holm Belt (=East Antarctic Orogen) is not a continuous belt from the Shackleton Range to the Lützow-Holmbukta, but interrupted by a crustal domain characterised by NW–SE trending aeromagnetic anomalies. They indicate a NW-directed Tonian indenter pushing apart the Shackleton Range region to the west and the Sør Rondane–Lützow-Holmbukta region to the east during Pan-African times (Jacobs et al., 2015; Ruppel et al., 2015) (Fig. 6.6). The indenter has been recently interpreted as an Oceanic Arc Super Terrane (Tonian Oceanic Arc Super Terrane – TOAST; Jacobs et al., 2015) that is sandwiched in between Kalahari margin and Indo-Antarctica – Rukerland. In the Sør Rondane

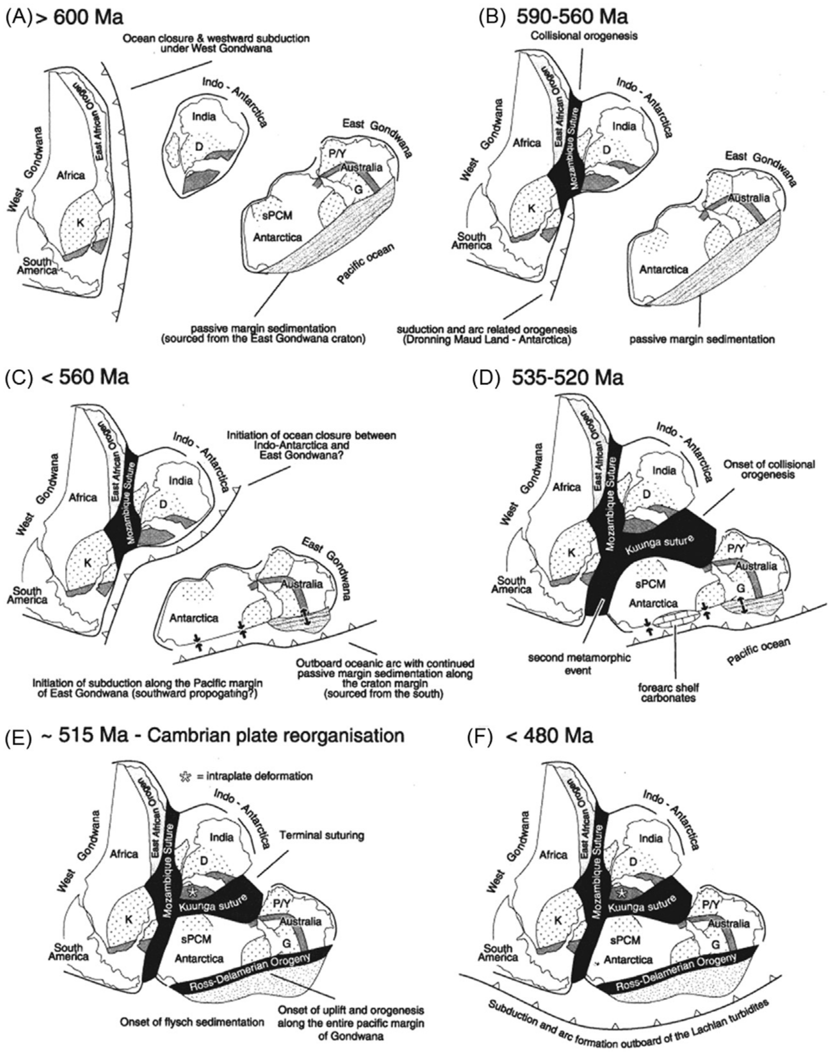


FIGURE 6.5 Main tectonic stages of the amalgamation of Gondwana (after Boger and Miller, 2004). Abbreviated Cratons – *D*, Dwarhai Craton; *G*, Gawler Craton; *K*, Kalahary Craton; *P/Y*, Pilbara/Yilgarn Craton; *sPCM*, Southern Prince Charles Mountains.

region, according to Ruppel et al., 2020 (and ref. therein), a polyphase tectono-metamorphic stage up to HP-HT grade and a coeval sequence of magmatic pulses from c. 650 to 500 Ma define a wide collisional margin of Pan-African age characterized by a fold and thrust tectonics and a protracted cooling pattern. Geophysical data (Mieth and Jokat., 2014) might indicate that the southern continuation of the TOAST terminates against an as yet unidentified craton further inland (the cryptic 'Valkyrie Craton'). Jacobs et al. (2015)

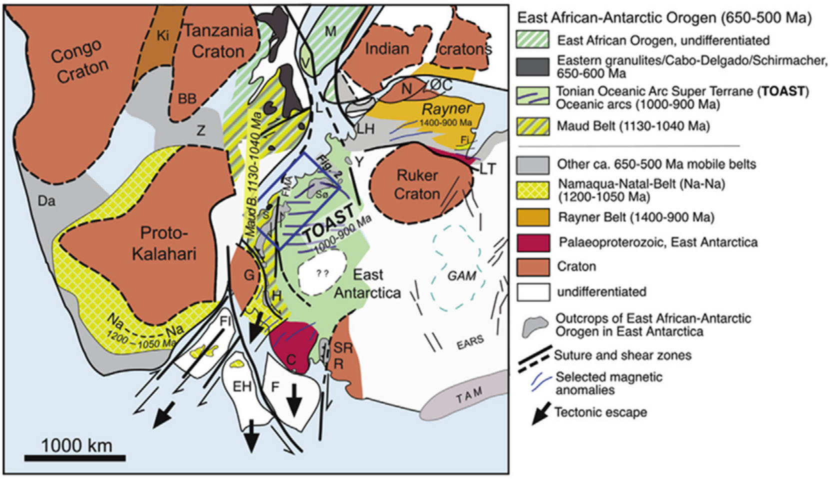


FIGURE 6.6 Geology of the African-Antarctica-India sector of Gondwana (after Jacobs et al., 2020). Abbreviations: *BB*, Bangweulu Block; *C*, Coats Land, *Da*, Damara belt; *EARS*, East Antarctic Rift System; *EH*, Ellsworth-Haag; *F*, Filchner block; *FI*, Falkland Islands, *Fi*, Fisher Terrane; *FMA*, Forster Magnetic Anomaly; G, Grunehogna; *GAM*, Gamburtsev Mts.; *H*, Heimefrontfjella; *Ki*, Kibaran; *L*, Lurio Belt; *LH*, Lützow-Holm Bay; *LT*, Lambert Terrane; *N*, Napier Complex; *Na-Na*, Namaqua-Natal; *M*, Madagascar; *ØC*, Øygarden Complex; *R*, Read Block; *S*, Schirmacher Oasis; *Sø*, Sør Rondane; *SR*, Shackleton Range; *TAM*, Transantarctic Mountains.

included the TOAST intraoceanic juvenile arc in the EAAO, but some features raise doubts on this interpretation: for example, the structural grain of this belt is orthogonal to that of the Dronning Maud Land and similar to other domains of interior Antarctica continent as observed by Mieth and Jokat (2014). A different model appears plausible. The composite 'Crohn Province' (that is the 'Crohn Craton' of Boger, 2011) could be involved in the Neoproterozoic drifting of the Indian block against the Mawson Continent as also proposed by Mulder et al. (2019) along the 'Pinjarra Orogen' in a way similar to that proposed by Markwitz et al. (2016) for western Australia. Alternatively, could the 'India-Napier-Rayner Belt' block have collided with the 'Crohn Province' when already joined to the 'Mawson Continent' in Mesoproterozoic time? This remains an open question.

2. **Indo-Antarctic Sector**. Almost all the authors involved in studies in this sector describe metamorphic events overprinting the Grenvillian basement, often grouped under the term 'reworking'. Inboard in the continent, along the 'Rayner Belt' and also in the Kaiser-Wilhelm Land (KWL) and Princess Elisabeth Land (PEL), a sequence of reworked rocks of high variable metamorphic grade dated c. 620−500 Ma, including many intrusive bodies, testifies a Pan-African event (the 'Kuunga Orogen' s.l.; Arora et al., 2020; Boger et al., 2001; Corvino et al., 2016; Fitzsimons, 1997; Fitzsimons, 2003; Grew et al., 2012; Kelsey et al., 2007;

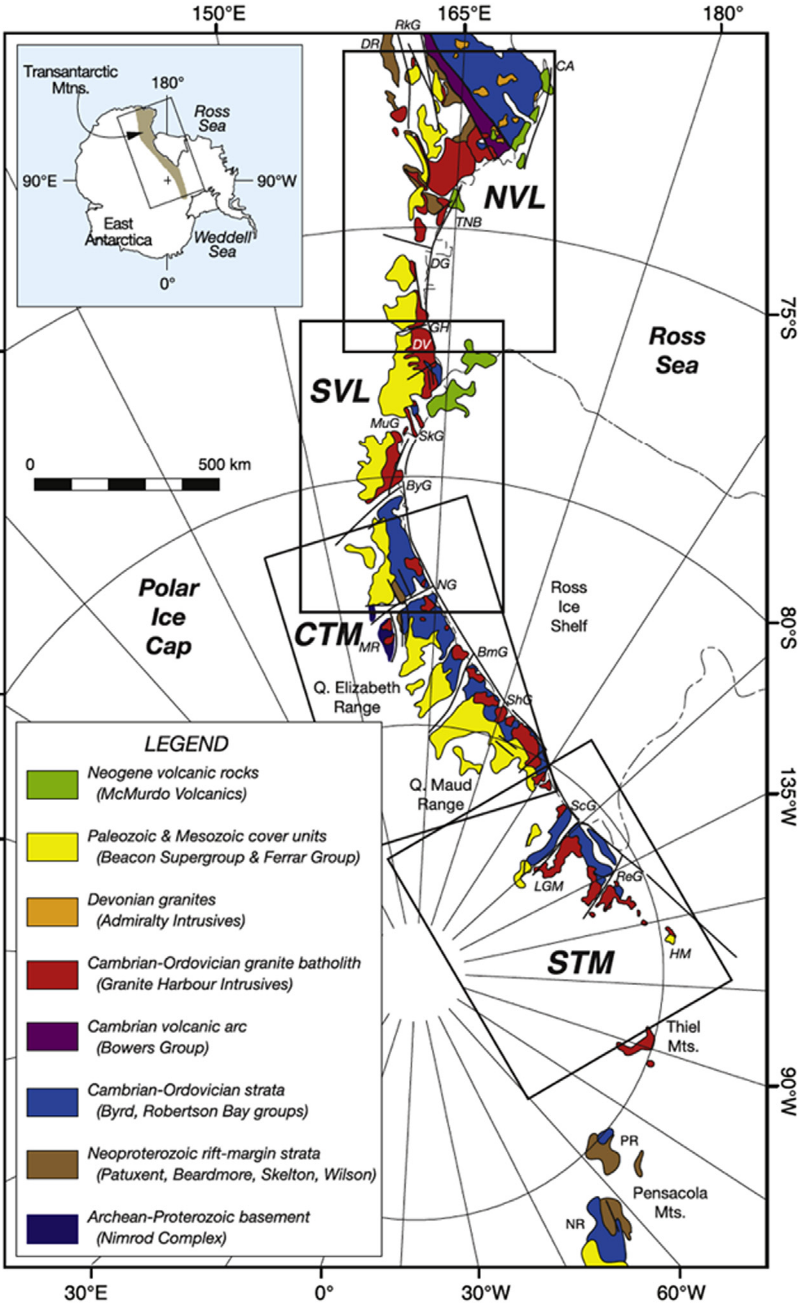
Liu et al., 2013, 2014, 2020; Morrissey et al., 2016; Phillips et al., 2009; Wang et al., 2008) related to the final collision of Indian Cratons and 'Rayner Belt' with a composite block including 'Ruker Craton', and here named the 'Crohn Province' including the TOAST. A hypothetical Pan-African suture has been introduced along the border of the Archaean Ruker Craton (Jacobs et al., 2015; Mulder et al., 2019). Some authors exclude the existence of a Pan-African suture in the Lambert Rift region, suggesting that the Kuunga Orogen represents a belt of intraplate reactivation (Phillips et al., 2006; Wilson et al., 2007). Arora et al. (2020) proposed that these Pan-African events in PEL are correlated to the Eastern Ghat Mobile Belt of India. A sequence of high-grade metamorphic reworking events related to the final assembly of Gondwana in the time range 650–500 Ma are also reported from different localities in the Lützow-Holm Bay region (Dunkley et al., 2020; Takahashi et al., 2018; Tsunogae et al., 2016 and ref. therein).

3. **Wilkes –Albany-Frazer Orogen in the Australo-Antarctic Sector**. In this province few evidence of Pan-African events are described. Sheraton et al. (1992) reported an age of c.515 Ma for outcrops of granites and syenites from David Island and c.500 Ma for alkaline mafic dykes at Bunger Hills. Daczko et al. (2018) reported U-Pb SHRIMP age (lower intercept) of c.533 Ma for a granite orthogneiss from Cape Harrison, interpreted as the age of metamorphism.

6.3.3 The 'Ross Orogen' in the Transantarctic Mountains during the late Precambrian–early Paleozoic evolution of the paleo-Pacific margin of Gondwana (c. 600–450 Ma)

The Transantarctic Mountains (Fig. 6.7), at the margin of the East Antarctica, represent a key element in providing an important but still cryptic record of Proterozoic and Early Paleozoic supercontinent history in the period from c.800 to 450 Ma. The existing dataset, and data that can be provided by future research, can be gathered from sedimentary, plutonic and volcanic assemblages that potentially reflect different tectonic events integral to the Rodinia–Gondwana transformation: from the breakup of Rodinia to the development of a transitional margin and plate-margin activity during Gondwana assembly. In the Transantarctic Mountains, constraints on the timing of Rodinia breakup are provided by siliciclastic sedimentation of Neoproterozoic age from two areas.

In the Skelton Glacier area, basalts interlayered with Skelton Group sediments yielded a Sm–Nd model age of 700–800 Ma (Rowell et al., 1993) and were dated (U–Pb zircon age) at c. 650 Ma by Cooper et al. (2011). In the Nimrod Glacier area (Cotton Plateau), gabbros and basalts interlayered with sediments of the Beardmore Group previously dated at 762 Ma (Sm–Nd isochron age, Borg et al., 1990) are now considered to have been


150°E
165°E
180°
RkG
DR
CA
Transantarctic Mtns.
180°
Ross Sea
90°E
90°W
East Antarctica
Weddell Sea
0°
NVL
TNB
DG
75°S
GH
DV
Ross Sea
SVL
MuG
SkG
0
500 km
ByG
Polar Ice Cap
NG
Ross Ice Shelf
CTM
MR
80°S
BmG
Q. Elizabeth Range
ShG
Q. Maud Range
135°W
ScG
90°E
ReG
LGM
HM
STM
Thiel Mts.
90°W
PR
Pensacola Mts.
NR
30°E
0°
30°W
60°W
LEGEND
Neogene volcanic rocks (McMurdo Volcanics)
Paleozoic & Mesozoic cover units (Beacon Supergroup & Ferrar Group)
Devonian granites (Admiralty Intrusives)
Cambrian-Ordovician granite batholith (Granite Harbour Intrusives)
Cambrian volcanic arc (Bowers Group)
Cambrian-Ordovician strata (Byrd, Robertson Bay groups)
Neoproterozoic rift-margin strata (Patuxent, Beardmore, Skelton, Wilson)
Archean-Proterozoic basement (Nimrod Complex)

(Continued)

emplaced at 668 Ma (U−Pb zircon age, Goodge et al., 2002). Thick siliciclastic successions were long interpreted as deep-water turbidities deposited in Proterozoic time along a rifted margin. Although the older parts of some successions may relate depositionally to the rifting process, recent investigations have demonstrated that some units are nearshore deposits and the depositional age of several major successions must be revised upward. For example, in the central Transantarctic Mountains, sandstones formerly included in the Beardmore Group and considered Neoproterozoic are now assigned to the Middle Cambrian or younger (Goodge et al., 2002). A transformation from drifting to active subduction mode is inferred in the late Neoproterozoic, starting at c. 615−620 Ma and continuing as a protracted contractional tectonic cycle including several discrete tectonic events during the Ross Orogenic cycle, until the Ordovician (Allibone et al., 1993; Armienti et al., 1990; Borg et al., 1987; Goodge, 2020; Goodge et al., 2002, 2010; Talarico et al., 2004 and ref. therein; Hagen-Peter and Cottle, 2016, 2018; Paulsen et al., 2007, 2013; Rocchi et al., 2004, 2011).

As result of these processes the Ross orogenic belt is exposed from Northern Victoria Land at the Pacific end up to the Pensacola Mountains at the Atlantic end. Westernmost Marie Byrd Land (Edward VII Peninsula) has to be considered as part of the same orogenic belt, from which it became isolated only much later, during a major phase of the evolution of the WARS around the end of the Cretaceous. The 'Ross Orogen' is characterised by folds, thrusts, very low- to high-grade metamorphism, granitoids, and flysch- and molasse-type sediments. Remarkable thrusts have been reported from Oates Land, which could be traced into the Australian continuation of the 'Ross Orogen' and the 'Delamerian Orogen' (Flöttmann et al., 1993). The systematic distribution of high- and low-pressure types of metamorphism (e.g., Talarico et al., 2004) and of S- and I-type granitoids (e.g., Vetter and Tessensohn, 1987) led to the model of subduction of the paleo-Pacific beneath East Antarctica. In southern Victoria Land, the older plutons also include a peculiar suite of highly alkaline rocks (nepheline syenites and carbonatites in the 'Koettlitz Glacier Alkaline Province', Cooper et al., 1997) and a stage of Precambrian rift-related magmatism and sedimentation has been documented by Cook (2007) in the Skelton Glacier area and dated at c. 650 Ma by Cooper et al. (2011). In northern Victoria Land, the Ross Orogen is made up

◀ **FIGURE 6.7** Simplified map of the major geologic units underlying the Transantarctic Mountains (after Goodge, 2020). Boxes indicate the main regions: *NVL*, North Victoria Land; *SVL*, South Victoria Land; *CTM*, Central Transantarctic Mountains; *STM*, Southern Transantarctic Mountains. *BmG*, Beardmore Glacier; *ByG*, Byrd Glacier; *CA*, Cape Adare; *DG*, Davis Glacier; *Dr*, Daniels Range; *DV*, Dry Valleys; *GH*, Granite Harbour; *HM*, Horlick Mountains; *LGM*, La Gorce Mountains; *Mr*, Miller Range; *MuG*, Mulock Glacier; *NG*, Nimrod Glacier; *NR*, Neptune Range; *PR*, Patuxent Range; *ReG*, Reedy Glacier; *ScG*, Scott Glacier; *ShG*, Shackleton Glacier; *SkG*, Skelton Glacier; *TNB*, Terra Nova Bay.

of three so-called terranes, as follows: the high- to medium-grade and granite-dominated Wilson Terrane to the west; the low-grade turbiditic Robertson Bay Terrane to the east; and the low-grade and volcanic-rich Bowers Terrane in between. These terranes are considered to be allochthonous or just adjacent paleogeographic domains including an intra-oceanic island arc (Bowers Terrane) and an accretionary wedge (Robertson Bay Terrane) (Tessensohn and Henjes-Kunst, 2005). In northern Victoria Land, a unique occurrence of ultra-high-pressure rocks, including well-preserved mafic eclogites as lenses and pods within metasedimentary gneisses and quartzites, decorates the tectonic boundary between the inboard Wilson Terrane and the Bower Terrane in the Lanterman Range. Geological, petrological and geochronological studies indicate that mafic, ultramafic and felsic host rocks in this region underwent a common metamorphic evolution with an eclogite facies stage about 500 Ma ago at temperatures of up to about 850°C and pressures greater than 2.6 Gpa (Di Vincenzo et al., 1997; Palmeri et al., 2003, 2007). This stage is probably preceded by an eclogitic event at about 530 Ma (Di Vincenzo et al., 2016).

As shown in the timeline reported in Fig. 6.8, a Cryogenian to early Ediacaran stage of extension and subsidence during a pre-orogenic passive-margin phase is marked by bimodal and mafic volcanism and silicoclastic sedimentation between about 670–650 Ma. The onset of the Ross orogenic cycle was marked by c. 600–615 Ma high-grade metamorphism (Hagen-Peter et al., 2016b) and by magmatism at c. 590–565 Ma (Hagen-Peter and Cottle, 2016a) and U-Pb data from glacial granitic clasts (Goodge et al., 2010, 2012). Continuity of magmatism between about 620 and 475 Ma is shown by major peaks in detrital-zircon age spectra from syn-orogenic silicoclastic rocks (Goodge et al., 2002, 2004; Paulsen et al., 2016). Periods of alternating convergent-margin upper plate contraction and extension are inferred from regional structural data and alternating changes in syn-tectonic magma compositions in the Granite Harbour Intrusives (Fig. 6.9), with structural data from northern Victoria Land, Central Transantarctic Mountains, Southern Transantarctic Mountains, and Pensacola Mountains also indicating strain partitioning during sinistral-oblique underflow. The Ordovician upper bound of the Ross orogenic cycle, and establishment of stable post-orogenic crust, is marked at about 470 Ma by an end of magmatism and low-T metamorphic cooling ages in all areas.

The total duration of the Ross cycle is therefore in the range of about 145 Ma and a typical contractional period in the long-lived evolution of the orogenic belt (Fig. 6.10) is commonly modelled based on data from the central Transantarctic Mountains, as the result of relative plate motions and partitioned into both contractional and strike-slip components of deformation. According to this model (Goodge et al., 1993a, 2004a) oceanward retreat of the subduction zone to the east (present-day coordinates) is inferred in order to explain offshore migration of the magmatic axis. The late Neoproterozoic and early Paleozoic tectonic evolution of the East Antarctic margin relate to the long-lived subduction between the East Antarctic shield and proto-Pacific oceanic lithosphere and occurred in four

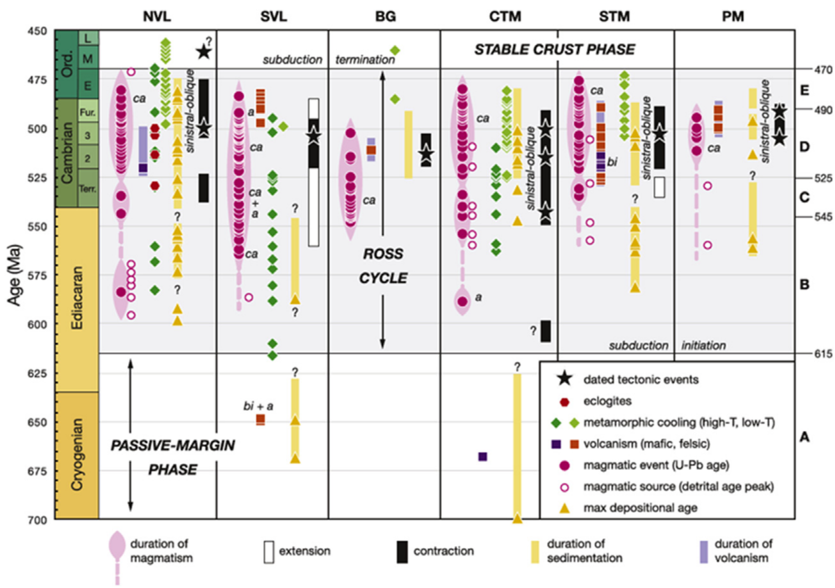


FIGURE 6.8 Timeline showing important tectonic events leading up to and during the Ross Orogenic cycle (after Goodge, 2020). Events with geochronological control are shown by symbols in the legend, and duration of events inferred from stratigraphic, magmatic, and structural relations are shown by colored vertical bars indicated in the diagram. *BG*, Byrd Glacier area; *CNVL*, northern Victoria Land; *CTM*, central Transantarctic Mountains; *PM*, Pensacola Mountains; *STM*, southern Transantarctic Mountains; SVL, southern Victoria Land.

(Goodge et al., 2004a, Fig. 6.11) or five main subsequent stages (synoptic diagram of Fig. 6.8 from Goodge, 2020): 700–615 Ma, a passive-margin stage (A); 615–545 Ma, a platform and incipient arc stage (B); 545-525 Ma, a synorogenic stage (C); 525–490 Ma, a late-orogenic stage (D); 490–470 Ma, a post-orogenic stage (E). The whole tectonomagmatic evolution is confirmed as a complex long-lived active continental margin in an accretionary-type orogenic system.

6.4 West Antarctic Accretionary System

Many papers have been published on the geological events that have modified the original Gondwanian paleo-Pacific margin of the 'Ross Orogeny' up to the present. Detailed recent reviews are those of Goodge (2020), Jordan et al. (2020) and Siddoway (2021). A highly dynamic history of convergence, continental growth and magmatism, alternating with extensional phases and fragmentation-rifting–rotation of microplates, built a complex accretionary system grossly starting from the inboard areas up to the most external ones where a subduction zone is still active in the South Shetland Islands. Such a tectonic evolution is common within accretionary orogens developed at active plate

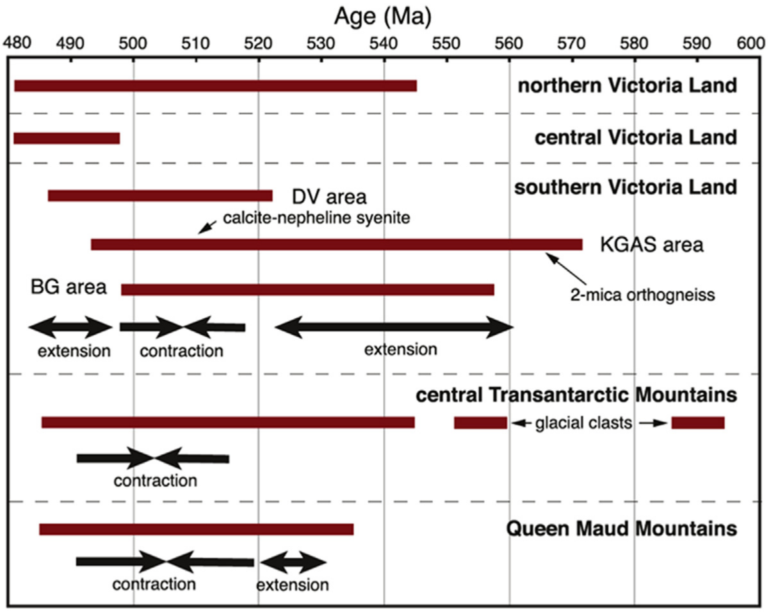


FIGURE 6.9 Timeline summarising zircon U\Pb age data for the Granite Harbour Intrusives along the Ross Orogen and related tectonic events as proposed by Hagen-Peter et al. (2016).

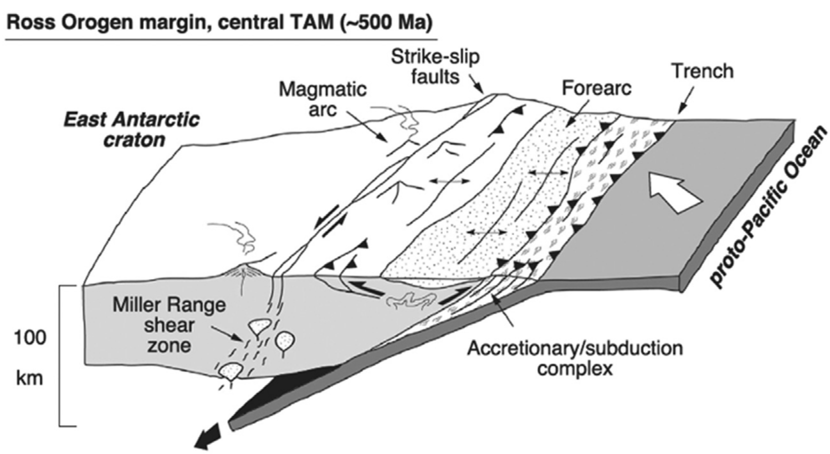


FIGURE 6.10 Schematic diagram of Transantarctic Mountains plate-boundary regime at ~500 Ma (after Goodge, 2020).

margins such as the American Cordillera or the Australian Tasmanides (Cawood and Buchan, 2007; Cawood et al., 2016; Collins, 2002). Convergence along the continental margins is often oblique, so continental fragments of the margin can easily be translated along strike and rotated.

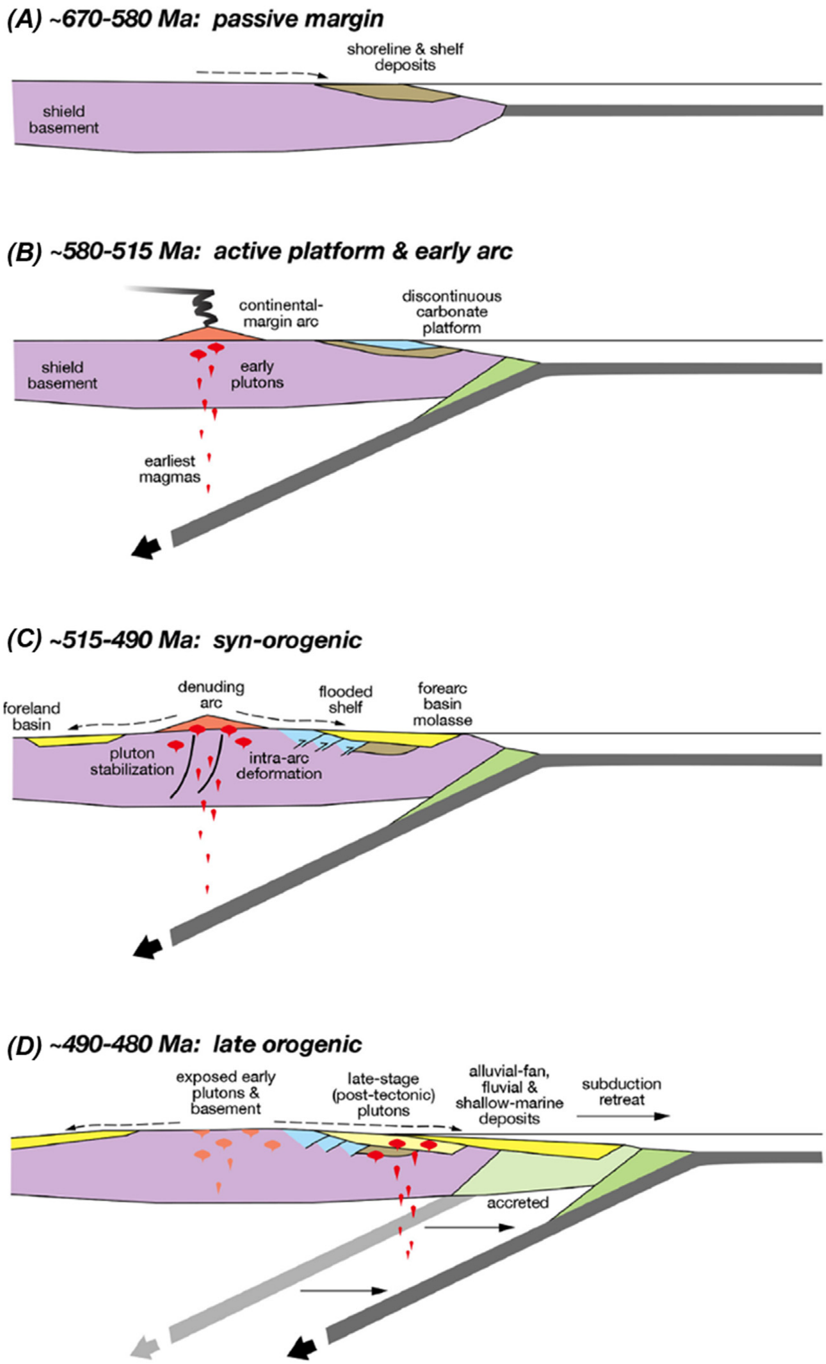


FIGURE 6.11 Model of the late Neoproterozoic and early Paleozoic tectonic evolution of the East Antarctic margin in the central Transantarctic Mountains (after Goodge, 2020). Crustal thicknesses approximately to scale, but sedimentary basins exaggerated in thickness for clarity.

6.4.1 West Antarctica in the Precambrian to Mesozoic (c. 180 Ma) evolution of Gondwana until the middle Jurassic breakup

In this time window, the main geological events so far known in the Antarctic geological record include the formation of a long-lived continental basin where a dominantly fluvial succession, the Beacon Supergroup, was deposited in Devonian to early Jurassic times, and the development of two main magmatic and deformational belts generally known as the middle Paleozoic 'Borchgrevink' and the Permo-Triassic 'Weddell' orogens.

6.4.1.1 Precambrian to Cambrian metamorphic basement

A wide domain of the West Antarctic Rift System, in inboard position adjacent to the Transantarctic Mountains, exposes a lower continental crust composed by crystalline rocks of Cambrian or older Mesoproterozoic age.

The Haag Nunataks crustal block (Jordan et al., 2017) is composed by felsic orthogneisses dated in the range of c. 1240−1060 Ma (emplacement) and deformed at c. 1060 Ma, that is during the Grenvillian Orogeny (Millar and Pankhurst, 1987; Riley et al., 2020). A similar Grenvillian age is proposed for the unexposed basement in the Ellsworth-Mountain crustal block (Castillo et al., 2017).

In the Ross Sea Rift System a thinned continental crustal of Cambro-Ordovician age is commonly assumed but recovered only from the DSDP site 270 and Iselin Bank (Mortimer et al., 2011). In the Ross Province of Mary Byrd Land several outcrops of Cambrian rocks are found. The Swanson Formation is the oldest exposed metasedimentary turbidite sequence related to the late − Ross forearc sedimentation (c. 514−490 Ma) and deformed by the subsequent tectonic events (Siddoway and Fanning, 2009 and ref. therein). At Mount Murphy a granodiorite orthogneiss has been dated at c. 505 Ma and correlated to the late magmatic pulses of the Ross Orogeny (Pankhurst et al., 1998).

6.4.1.2 Devono-Carboniferous arc magmatism ('Borchgrevink Event') (c. 370−350 Ma)

The occurrence in northern Victoria Land of a calc-alkaline magmatic suite of middle Paleozoic age (c. 370−350 Ma; Borg et al., 1986, 1987; GANOVEX Team, 1987; Vetter et al., 1983) is attributed to an orogenic event separated from the waning phase of the Ross Orogeny (Craddock, 1970) by about 80 Ma. Often related to the middle phase of the 'Lachlan fold and thrust Belt' in eastern Australia (Forster and Goscombe, 2013 and ref. therein) the plutonic suite includes dominant granitoids (Admiralty Intrusives) and minor felsic volcanic (Gallipoli Volcanics, Fioretti et al., 2001). The granitoids form several plutons which mainly occur in the Robertson Bay Terrane, although some stitch the tectonic boundary between the Bowers and Wilson Terrane, where the coeval volcanics are also concentrated. All the intrusions are characterised by isotropic fabrics and discordant contacts with respect to the surrounding metasediments suggesting a post-tectonic emplacement. These features, and meagre evidence from radiometric data for concomitant metamorphism and deformation, have so far prevented a comprehensive reconstruction of the tectonic setting and development of an orogenic event called the 'Borchgrevink Orogeny' (Craddock, 1970) in northern Victoria Land. Nevertheless, the continental arc geochemical signature of the Devonian−Carboniferous magmatic suite indicates that the plutonism most likely occurred as the result of a renewed period of subduction activity along the-paleo-Pacific margin (Borg and DePaolo, 1991; Kleinschmidt and Tessensohn, 1987). Similar suites are known in Tasmania, New Zealand and the Campbell Plateau (Bradshaw et al., 1997; Gibson and Ireland, 1996), and in the Ford Ranges of Marie Byrd Land (emplacement age c. 375−345 Ma, Nelson and Cottle, 2018; Yakymchuk et al., 2015 and ref. therein) and in the Antarctic Peninsula (Riley et al., 2016). In Thurston Island granodiorite orthogneisses in the basement are dated at c. 349 Ma (Riley et al., 2016) and 347 Ma (Nelson and Cottle, 2018) suggesting the presence of a rifted crustal micro-plate traslated in to West Antarctica from an inboard position (see Fig. 6.14 from Paulsen et al., 2017).

A correlation between the 'Borchgrevink Event' of northern Victoria Land and the Tasmanian Orogeny was put forward by Findlay (1987). Elsewhere in Antarctica, marine sediments from the Ellsworth Mountains and Pensacola Mountains are considered to have been deposited in the same time window, within an intracontinental basin that, may have extended from the Weddell Sea to the Ross Sea (Elliot, 1975).

6.4.1.3 Beacon Supergroup (Devonian-Permo-Triassic-earliest Jurassic)

The Beacon Supergroup consists of dominantly continental sedimentary deposits forming a generally flat-lying, 0.5−3.5 km thick cover developed over a marked unconformity (Kukri Peneplain) above the Ross orogenic belt throughout most of the Transantarctic Mountains (Elliot, 2013 and ref. therein) (Fig. 6.12). Similar sequences are known in limited outcrops in East Antarctica (e.g., Prince Charles Mountains, Dronning Maud Land and the Ellsworth Mountains) (Barrett, 1991) and as bedrock of the Cenozoic glaciomarine sediments in the Victoria Land Basin (VLB) in the Ross Sea as documented by the CRP-3 drill-hole (Cape Roberts Science Team, 2000). Outside Antarctica, similar sedimentary rocks, collectively called Gondwanian Sequences, occur in South Africa, Australia and South America. The deposition of these sediments in Antarctica started in Devonian time with the Taylor Group, consisting of dominantly quartz-arenites and conglomerates. A product of erosion and fluvial processes under arid and semiarid conditions (Campbell and Claridge, 1987), the Taylor Group accumulated in a series of basins along the paleo-Pacific margin of Gondwana. The deposition of fossiliferous

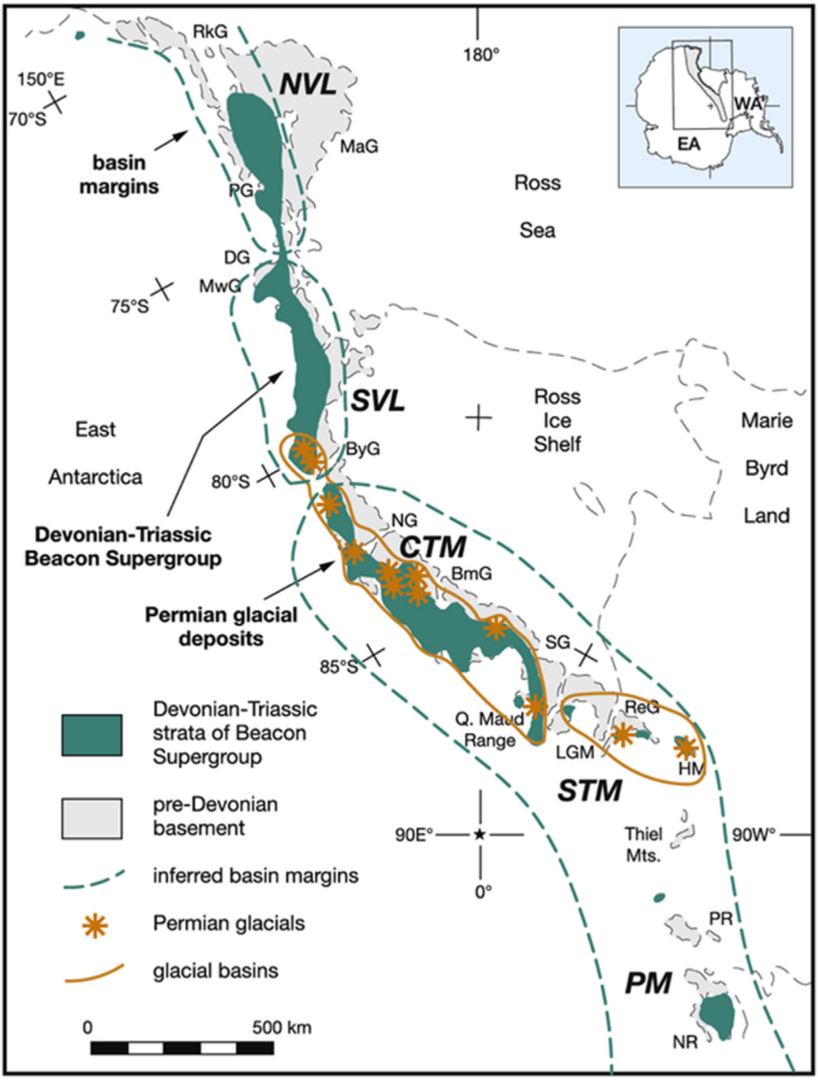


FIGURE 6.12 Distribution of Beacon Supergroup strata in the Transantarctic Mountains (after Elliot, 2013, as modified by Goodge, 2020).

siltites was followed by an erosional phase, probably related to a glacial event which led to the deposition of glaciogenic sediments in late Carboniferous–Early Permian time. In the Transantarctic Mountains, the Carboniferous to Triassic sediments form the Victoria Group, which includes carbonaceous layers and feldspathic sandstones. A summary of the main lithostratigraphic features in the central Transantarctic Mountains, southern Victoria Land and northern Victoria Land is displayed in Fig. 6.13. The low sedimentation rate (c.12.5 m/My) and absence of

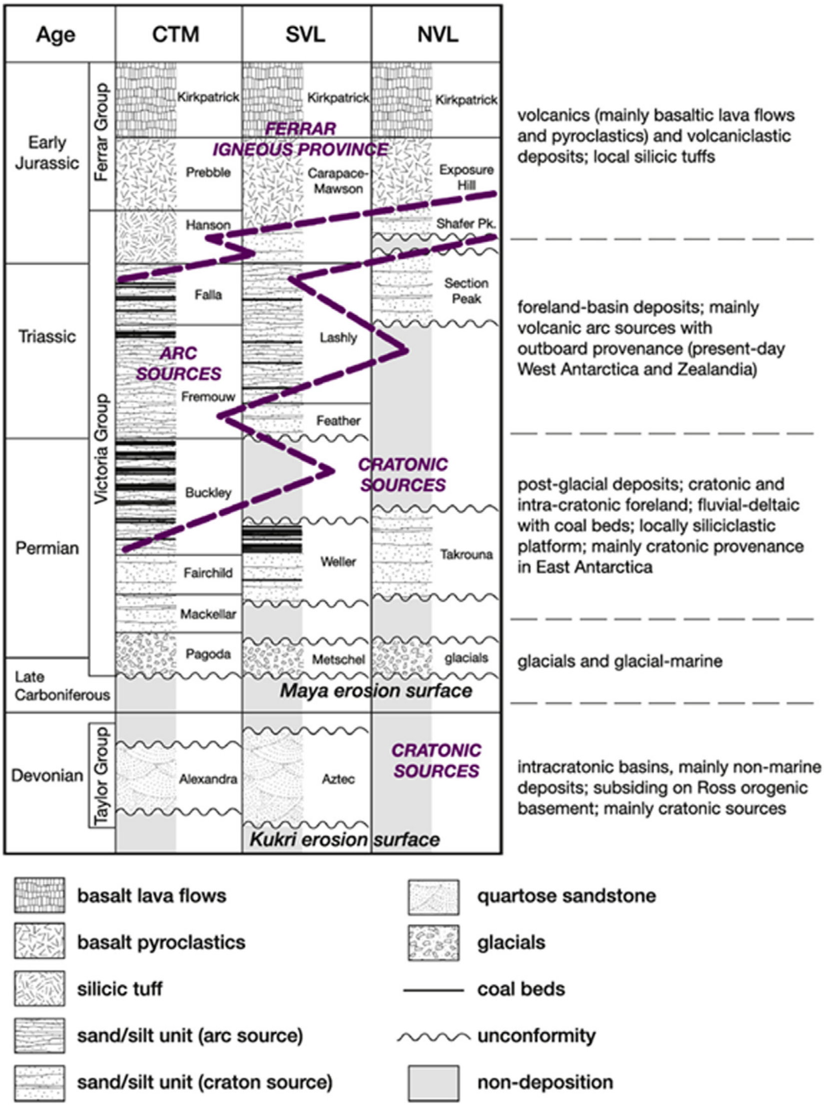


FIGURE 6.13 Beacon Supergroup stratigraphy in the central Transantarctic Mountains, southern Victoria Land and northern Victoria Land, showing major stratigraphic units of the Beacon Supergroup and Ferrar Group. Trends in depositional environments and sediment sources are generalised (after Elliot, 2013 as modified by Goodge, 2020).

concomitant compressional deformation suggest that deposition occurred over a thick continental crust (Barrett, 1991) but different tectonic models have been proposed, including a passive margin (Isbell, 1999), intra-cratonic (Barrett, 1991; Woolfe and Barrett, 1995) or marginal/back-arc (Bradshaw and Webers, 1988).

6.4.1.4 The Ellsworth-Whitmore Mountains Terrane and the Permo-Triassic arc magmatism

The Ellsworth or Weddell Orogeny (or – in a larger context – also known as the Gondwanide Orogeny which spanned parts of southern Africa – Cape Fold Belt – and South America – Sierra de la Ventana Fold Belt) occurred in Permo-Triassic time c. 250–200 Ma. As noted by Cawood (2005), this orogeny overlaps with the end Paleozoic assembly of Pangea (Li and Powell, 2001), through ocean closure and accretion of Gondwana, Laurussia (Laurentia + Baltica) and Siberia, as well as completion of terrane accretion in the Altaids. Stratigraphic and geochronological data (Dalziel, 1982; Dalziel and Elliot, 1982; Johnston, 2000; Storey et al., 1987; Trouw and De Wit, 1999) indicate that Permo-Triassic deformation of variable intensity and distribution occurs throughout West Antarctica and the adjoining Cape Fold Belt of southern Africa. In Antarctica, this orogenic event is well documented in the Ellsworth–Whitmore Mountains and in the Pensacola Mountains, where upright to inclined folds with axial planar cleavage are inferred to have formed in a dextral transpressive environment (Curtis, 1998). The Ellsworth–Pensacola Mountains chain represents the fold belt. It merges with the Ross Orogen in the Pensacola Mountains (Fig. 6.2), where the tectonism partly overprinted the older Ross-aged structures. Elsewhere, deformation is heterogeneously distributed, with Storey et al. (1987) noting that in the Antarctic Peninsula, unconformities previously ascribed to the Gondwanide Orogeny are younger and that the only event related to Gondwanide deformation is the regional metamorphism at 245 Ma. Parts of the Trinity Peninsula Group where affected. The trend of the Ellsworth Orogen is at a high angle to the paleo-Pacific margin of Antarctica and the Ross Orogen. This obliqueness is due to secondary rotation, determined from paleomagnetic investigations by Funaki et al. (1991) and confirmed by Randall and Mac Niocaill (2004).

Triassic magmatism is widespread in the Antarctic Peninsula (Figs. 6.2 and 6.14) and dated c.200–230 Ma (Rb/Sr and U/Pb methods) in Graham Land, eastern Palmer Land and Thurston Island (Bastias et al., 2020; Flowerdew et al., 2006; Pankhurst, 1982; Riley et al., 2012, 2017; Vaughan and Storey, 2000; Weaver et al., 1994) and eastern Mary Byrd Land (Nelson and Cottle, 2018). These magmatic rocks are generally orthogneisses derived by protoliths with a calcalkaline to alkali-calcic affinity and a variable peraluminous character. They testify to the presence of an active Andean-style continental magmatic arc all along the Antarctic margin to the Patagonian belts. Metamorphic age of these intrusives is debated, and referred mainly to a middle Cretaceous contractional event (Vaughan et al., 2012).

6.4.1.5 Ferrar Supergroup and the Gondwana breakup (c. 180 Ma)

The first major tectonic stage in the breakup of Gondwana corresponds to an initial rifting phase that started in the Weddell Sea, initially as a back-arc

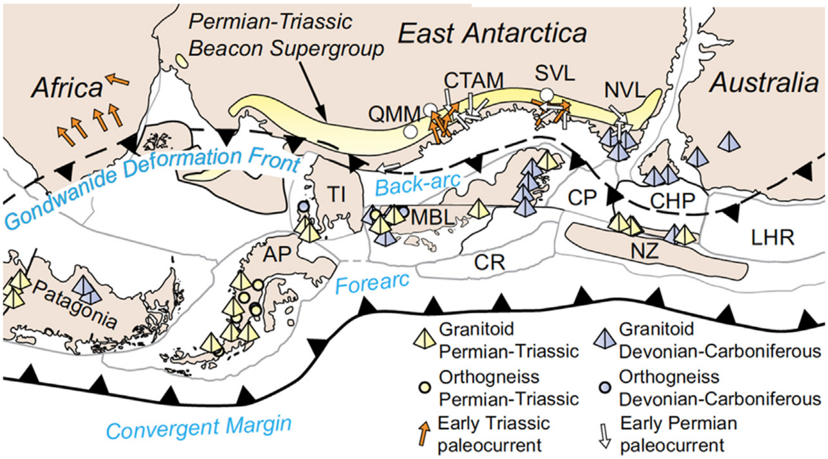
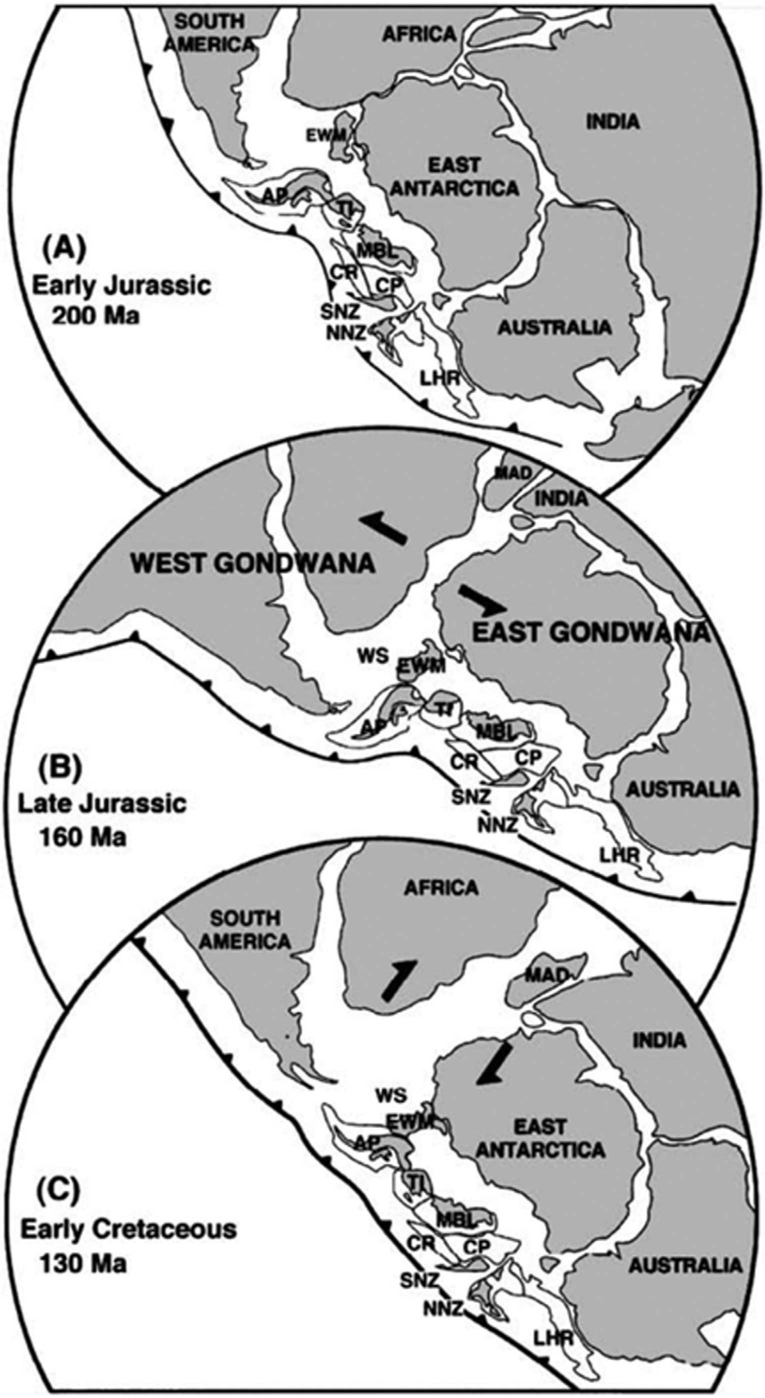


FIGURE 6.14 Permian–Triassic Gondwana reconstruction showing the distribution of Permian– Triassic strata of the Beacon Supergroup with respect to the trace of the Gondwanide deformation front and Permian–Triassic plutonic and metamorphic rocks located in the orogenic hinterland. *AP*, Antarctic Peninsula; *CR*, Chatham Rise; *CP*, Campbell Plateau; *CHP*, Challenger Plateau; *CTAM*, central Transantarctic Mountains; *LHR*, Lord Howe Rise; *MBL*, Marie Byrd Land; *NVL*, north Victoria Land; *NZ*, New Zealand; *QMM*, Queen Maud Mountains; *SVL*, south Victoria Land; *TI*, Thurston Island (after Paulsen et al., 2017).

basin, in the Late Jurassic (Fig. 6.15) (Heimann et al., 1994; Jordan et al., 2017, 2020; Lawver et al., 1991; Riley et al., 2020 and ref. therein). This stage involved right-lateral transtension as East Gondwana (Antarctica, Australia, India and New Zealand) and West Gondwana (South America and Africa) moved apart with stretching beginning in the north and propagating southward (Lawver et al., 1992). Initial breakup involved a complex geodynamic evolution characterised by the rotation and translation of several microplates such as the Ellsworth Mountains block; a displaced part of the Gondwanide fold belt. The original position of the various microplates is still controversial, but the Ellsworth–Whitmore Mountains crustal block most likely originated from the paleopacific margin of the Ross Orogen (Elliot et al., 2016). Rotation of West Antarctic microplates must have been accomplished by c. 165 Ma (the time of opening of the Weddell Sea), and rotation of the Ellsworth–Whitmore Mountains crustal block was finished before translation into its present position by 175 Ma (Grunow et al., 1987).

Rotation of microplates did not apparently involve the production of oceanic crust (Marshall, 1994) but rather a crustal block rotation with controlling faults concealed beneath Mesozoic sedimentary basins in the Weddell Sea (Storey, 1996; Storey et al., 1988). The breakup has been explained by several authors as the result of a hot mega-plume, which, according to Storey and Kyle (1999), could have promoted domal uplift and formation of a triple junction. According to Dalziel et al. (1999), the Gondwana plume


SOUTH AMERICA
AFRICA
INDIA
EWM
EAST ANTARCTICA
AP
TI
(A)
MBL
Early Jurassic
200 Ma
CR
CP
SNZ
NNZ
AUSTRALIA
LHR
MAD
INDIA
WEST GONDWANA
EAST GONDWANA
WS
EWM
TI
AP
MBL
(B)
CR
CP
Late Jurassic
160 Ma
SNZ
NNZ
AUSTRALIA
LHR
AFRICA
SOUTH AMERICA
MAD
INDIA
WS
EWM
EAST ANTARCTICA
AP
TI
(C)
Early Cretaceous
130 Ma
MBL
CR
CP
SNZ
NNZ
AUSTRALIA
LHR

(*Continued*)

may have also caused or expedited formation of the early Mesozoic Gondwanide fold belt, due to the buoyancy of a hot plume acting on the downgoing slab and causing it to flatten. In both cases, plume activity led to the production of magma batches reflecting different degrees of plume–lithosphere interaction, which migrated along crustal shear zones to ultimately form the various large igneous provinces. The plume-related magmatic products are represented by huge within-plate mafic and felsic magmatic provinces in many Gondwana continents as well as Antarctica (Cox, 1988; Storey, 1996; White and MacKenzie, 1989). In the Transantarctic Mountains (LeMasurier and Thomson, 1990), Jurassic mafic rocks are generally known as the Ferrar Supergroup or the Jurassic Ferrar Large Igneous Province (FLIP) (Fig. 6.16). They are divided into a volcanic component, the Kirkpatrick Basalt – preceded by extensive phreatomagmatic volcanoclastic rocks (Elliot et al., 2006; Viereck-Götte et al., 2007) – and the intrusive dolerite (diabase) sills and dikes of the Ferrar Dolerite. Geochemically, the FLIP is unusual with upper crustal-like characteristics such as high-large ion lithophile element concentrations and enriched isotopic signatures ($^{87}Sr/^{86}Sr > 0.709$). The crust-like signature suggests derivation from an enriched lithosphere, possibly connected to subduction along the Pacific margin of Gondwana during the Paleozoic and Mesozoic. The mafic rocks are concentrated within a long linear belt exposed in Tasmania, Antarctica and South Africa (Ivanov et al., 2017). Cox (1988) considered that the linear pattern of the Ferrar and Tasman provinces could not be compatible with classic circular plumes and proposed a hot line rather than a hot spot. A number of rifts that intersected at the Dufek Massif (Elliot, 1992) could have favored the development of zones of weakened lithosphere, which acted as pathways for lateral migration of magmas derived from lithospheric sources. In spite of the still not completely understood tectonic setting, it is important to note that, similarly to other continental flood basalt provinces, all the mafic products formed during a short period of eruption (Tasman province: 175 ± 18 Ma; Ferrar Supergroup: 180–183 Ma; Dronning Maud Land province: 177 ± 2 Ma; Karoo province: 182 ± 2 Ma) (Heimann et al., 1994; Hergt et al., 1989). Coeval felsic intrusions considered to be rift-related are known in West Antarctica (Storey et al., 1988) and southern South America (Chon-Aike or Tobıfera province) (Gust et al., 1985). The

◀ **FIGURE 6.15** (A) Gondwana tight fit reconstruction (Early Jurassic). (B and C) Major initial stage in the break-up of Gondwana: (A) Initial rifting stage (Late Jurassic); (B) Change in the Gondwana break-up stress regime from dominantly North–South between East and West Gondwana to dominantly East–West with the two-plate system being replaced by a multiple-plate system (Early Cretaceous) (after Fitzgerald, 2002; Lawver et al., 1992, 1998). *AP*, Antarctic Peninsula; *TI*, Thurston Island; *MBL*, Marie Byrd Land; *CR*, Chatham Rise; *CP*, Campbell Plateau; *SNZ*, Southern New Zealand; *NNZ*, Northern New Zealand; *LHR*, Lord Howe Rise; *WS*, Weddell Sea. Continent and microplate positions are from Lawver et al. (1992, 1998) with other information from Storey (1996).

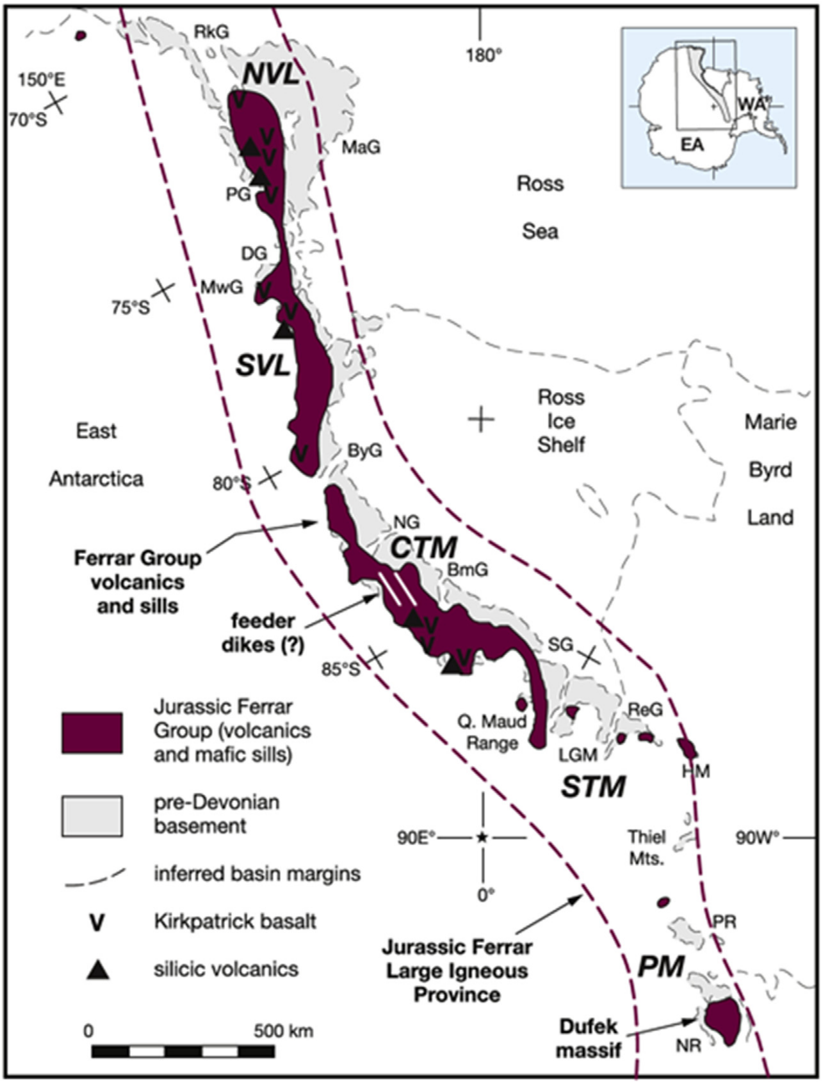


FIGURE 6.16 Distribution of the Ferrar Large Igneous Province in the Transantarctic Mountains (after Elliot, 2013; Elliot and Fanning, 2008 as modified by Goodge, 2020). Location of possible feeder dikes in the Nimrod Glacier area imaged in aeromagnetic data (Goodge and Finn, 2010).

second major geodynamic stage occurred in the Early Cretaceous (Fig. 6.15), as a consequence of the change of the Gondwana breakup stress regime from dominantly north–south between East and West Gondwana to dominantly east–west with the two-plate system being replaced by a multiple-plate system (e.g., Lawver et al., 1992).

Felsic intrusives outcrop in the Whitmore Mountains region (Craddock et al., 2017; Leat et al., 2018) at Pagano Nunatak, Pirrit Hills, Nash–Martin Hills and Linch Nunataks. These granites, emplaced at c. 174–178 Ma, postdate the Ferrar effusive event. They are mildly peraluminous, two-mica granites with chemical composition suggesting a within-plate affinity. While the origin is debated, Craddock et al. (2017) considered that they formed from hybridised magmas between mantle melts and a lower continental crustal contribution.

This modification in stress regime is thought to have induced large-scale ductile deformation concentrated along shear zones in the Antarctic Peninsula (Storey et al., 1996) and thin-skinned deformation and inversion of existing sedimentary basins such as the Latady Basin (Kellogg and Rowley, 1989). By ca. 110 Ma, the microplates of West Antarctica had nearly reached their present location with respect to East Antarctica. In the same time period, separation also began between India and Antarctica (Lawver et al., 1991). Initial stretching between Australia and Antarctica began as early as 125 Ma (Stagg and Willcox, 1992), but sea-floor spreading was delayed to c. 95 Ma (Cande and Mütter, 1982; Royer and Rollet, 1997; Veevers et al., 1990) in the Ross embayment (Elliot, 1992), as well as extension between the Lord Howe Rise and northern New Zealand (Lawver et al., 1992). By the Late Cretaceous, Antarctica had reached its final polar location and configuration, and the final stage of break-up was completed when Zealandia (New Zealand plus Campbell Plateau; Mortimer, 2017) rifted from Marie Byrd Land at 84 Ma (e.g., Lawver et al., 1991; Mortimer et al., 2019; Stock and Molnar, 1987). In this geodynamic context, many particularly large and conspicuous intraplate fracture zones have been investigated. All related to extensive and prolonged extensional regimes spanning in time from Mesozoic to present (Fig. 6.2). These major extensional zones include:

- the Lambert Graben or Lambert Rift (East Antarctica);
- the Graben of Jutulstraumen and Penckmulde (occasionally called Jutul Penck Graben East Antarctica);
- the WARS (West Antarctic Rift System) including the Ross Sea Rift System (RSRS) and the Weddel Sea Rift System (WSRS); and
- the Rennick Graben as the main element of a strike–slip fault system in Victoria and Oates Lands.

The **Lambert Graben**, developed in the East Antarctic Craton, is filled mainly by sediments of the Permo-Triassic Beacon Supergroup and was interpreted by Harrowfield et al. (2015) as an accommodation zone of a wide intracontinental rift that extended from Australia's North West Shelf, between India and Antarctica, to southern Africa. Faulting started during the early Palaeozoic, reached its peak in the Permian and continued to the Early Cretaceous (Hofmann, 1996; Mikhalsky and Leitchenkov, 2018). Possibly, the trench in which subglacial Lake Vostok (Kapitisa et al., 1996; Siegert et al., 2011) sits belongs to the same rift system, but somewhat offset. The continuation of the Lambert Graben in the Indian landmass is the Mahanadi Rift south-west of Calcutta in the state of Orissa (Hofmann, 1996), filled with sediments of the same type and age as the Lambert Graben. The reconstruction of the Gondwanian India–Antarctica fit using these graben systems coincides with reconstructions by Archaean to Early Proterozoic elements for Gondwana.

Many other subglacial topographic basins similar to rift structures have been identified by geophysical methods beneath the thick Antarctic Ice Sheet. These basins, whose surface is often located below sea level, record a multi-phase history of rifting and subsidence and are the depocentres of thick sedimentary sequences (Aitken et al., 2014; Maritati et al., 2016 and ref. therein). Several hundred subglacial lakes have been identified in the topographic depressions (Siegert et al., 2005; Wright and Siegert, 2012).

The **Ross Sea Rift System** is extremely wide (about 1000 km). Extension and subsidence started during the late Mesozoic (about 140 Ma ago) and reached its main activity in the Cretaceous and Early Tertiary. An event at 55-50 Ma produced an enormous relief at its western shoulder, the Transantarctic Mountains. The crustal extension, combined with alkaline intra-continental volcanism, is still active at Mt. Erebus (3794 m) and at Mt. Melbourne (2732 m), both located in Victoria Land (Kyle, 1990a,b; Kyle and Cole, 1974; Tessensohn and Wörner, 1991; Wörner et al., 1989).

The Weddel Sea Rift System.

The **Jutul Penck Graben** of western Dronning Maud Land originated probably around 140 Ma ago or a little bit later (Jacobs and Lisker, 1999). The graben marks the boundary of the Grunehogna Craton towards the south-east which follows an ancestral geological structure.

The Rennick Graben. The strike–slip fault system of Victoria and Oates Lands runs obliquely to the Ross Sea Rift and is cut by it. The principal element of the system is the Rennick Graben, which is presently active, as demonstrated by earthquakes in 1952, 1974 and 1998. The graben contains downfaulted Ferrar volcanics and sediments of the Beacon Supergroup, which have been spectacularly folded and squeezed along the graben shoulders (Rossetti et al., 2003). These structures demonstrate alternating dextral transpression and transtension (with formation of pull-apart basins), within the setting of a complicated strike–slip system (Rossetti et al., 2003). The 1974 earth quake occurred some 120 km to the west at the parallel structure of the Matusevich Glacier. There is an ongoing discussion of the interpretation that the strike–slip fault system represents the continuation of oceanic fracture zones between Australia and Antarctica (e.g., the Tasman Fracture Zone) into the continental crust of Antarctica (Kleinschmidt and Läufer, 2006; Salvini and Storti, 2003; Salvini et al., 1997).

6.4.1.6 The Antarctic Andean Orogen

The orogen of the Antarctic Andes occupies the entire Antarctic Peninsula down to the Walgreen Coast (Fig. 6.2). It formed mainly in three episodes partly involving an older Permo-Triassic basement and may be an even older Paleozoic lower crust (Bastias et al., 2020 and ref. therein): (1) Late Jurassic through Early Cretaceous (150−140 Ma); (2) mid-Cretaceous (c.105 Ma); and (3) Tertiary (c.50 Ma to recent), and is partly still active (e.g., Birkenmajer, 1994; Burton-Johnson and Riley, 2015; Vaughan and Storey, 1997). Thus, it represents the youngest growth zone of the continent. The Antarctic Andes are a typical subduction orogen accompanied by orogenic magmatism in the form of granitic plutons and volcanic rocks. In detail, the deformation and metamorphism are very complicated, because they are polyphase. Folding and thrust faulting is reported mainly from the southern portion of the Antarctic Peninsula (Palmer Land and Alexander Island) and from the extreme north (Trinity Peninsula and eastern South Shetland Islands). The distribution of related metamorphism is also heterogeneous, including high-pressure metamorphism with blueschists characteristic of subduction complexes, e.g., on Elephant Island (Trouw et al., 1991).

The only plate-tectonically more or less active in Antarctica is situated north-west of the Antarctic Peninsula, in the South Shetland Islands (from Snow Island in the south-west up to Elephant Island in the north-east) and the Bransfield Strait (Fig. 6.2). North-west of the South Shetland Islands, a small section of the ocean floor, called the Drake Plate (i.e. the remnant of the older, but largely subducted Phoenix Plate), is being subducted at the South Shetland Trench beneath the Antarctic Plate. Related, mainly andesitic, volcanism forms the island arc of the South Shetland Islands. Parts of the South Shetland Islands (part of Livingston Island and Elephant Island) belong − as does the Peninsula itself − to earlier stages of the Antarctic Andean orogeny and consist of strongly deformed Jurassic trench sediments. The Bransfield Strait is located south of the subduction-related volcanic island arc and forms an active extensional basin accompanied by tholeiitic volcanism, partly submarine, partly as active island volcanoes (e.g., Penguin Island and Deception Island; Smellie and Lopez-Martınez, 2002). The Bransfield Strait often is regarded as a classic example of a back-arc basin, but recently this has been disputed (Gonzales-Casado et al., 2000).

6.5 Mesozoic to Cenozoic Tectonic Evolution of the Transantarctic Mountains

The Transantarctic Mountains, one of the major young mountain chains on Earth, separate East Antarctica from the West Antarctica Rift System and the largely land-based East Antarctic Ice Sheet from the marine-based West Antarctic Ice Sheet over a substantial portion of the Antarctic interior along the margin of the Ross embayment (Fig. 6.1B). The regional structural architecture of the Transantarctic Mountains remains poorly known in most regions because of the extensive ice cover. The East Antarctic Ice Sheet hides the structure of the mountains along the Polar Plateau, preventing the identification of the extent of mountain structures into East Antarctica. In the McMurdo Sound coastal area of the Ross embayment, thin but extensive piedmont glaciers obscure the structural boundary ('Transantarctic Mountains Front') with the off-shore Victoria Land rift basin. The main drainage for the East Antarctic Ice Sheet into the Ross embayment of West Antarctica is through outlet glaciers carved through the mountains. It has long been inferred that these outlet glaciers developed where faults cut transverse to the mountain trend (e.g., Cooper et al., 1991; Davey, 1981; Fitzgerald, 1992; Gould, 1935; Grindley and Laird, 1969; Gunn and Warren, 1962; Tessensohn and Wörner, 1991; Wrenn and Webb, 1982), and the occurrence of pseudotachylites has been proved a valuable tool to constrain the age of some faults (Di Vincenzo et al., 2004).

In most cases, however, there is little direct evidence either for the existence of a fault, or of its age. This is a key issue to address if the structures responsible for localized differential uplift of discrete mountain blocks are to be identified (e.g., Van der Wateren et al., 1999). The results will aid in understanding of localized valley incisions and how much erosion has contributed to mountain uplift. This could provide constraints on the development of the pathways for drainage of the East Antarctic Ice Sheet. Unlike other young mountain belts formed at the convergence plate boundaries (Alpine–Himalayan system and North American and Andean Cordilleras), the Transantarctic Mountains display a tight genetic link between mountain uplift and intra-plate rifting processes within the Antarctic plate.

The Transantarctic Mountains are generally interpreted as a high-relief rift flank uplift (Van der Wateren et al., 1999), formed during the Mesozoic–Cenozoic breakup of the Gondwana supercontinent (Cooper et al., 1987, 1991; Davey and Brancolini, 1995; Fitzgerald and Stump, 1997; Tessensohn and Wörner, 1991). The position of the Transantarctic Mountains in an extensional, rather than a contractional, tectonic regime had been recognised by pioneering Antarctic geologists, who interpreted the mountain chain as a fault-bounded horst block (David and Priestley, 1914; Gould, 1935). More recent structural investigations indicate that the Transantarctic Mountains consist of a linear to curvilinear chain of asymmetric tilt blocks bounded on the West Antarctic edge by a major normal fault zone and subdivided by transverse faults (Fitzgerald, 1992; Fitzgerald and Baldwin, 1997; Fitzgerald et al., 1986; Tessensohn and Wörner, 1991; Tessensohn, 1994a,b). Active rift tectonics and mountain uplift have been inferred from the presence of active volcanism and Neogene–Quaternary age faulting in the western portion of the rift and the Transantarctic Mountains (Behrendt and Cooper, 1991; Davey and Brancolini, 1995; Jones, 1997). The Cenozoic–Cretaceous asymmetric uplift and subsequent erosion exposed basement rock and older sediments along the coastward side of the Transantarctic Mountains, leaving younger sediments only on the inland side. Apatite fission track analysis in the McMurdo Sound area indicates an uplift of c. 6 km since c. 55 Ma. Other sectors of the Transantarctic Mountains record denudation events in the Late Cretaceous (Fitzgerald, 1992, 1995, 2002; Studinger et al., 2004) (Fig. 6.17) and in the Early Cretaceous (e.g., Scott Glacier area: Fitzgerald and Stump, 1997; Stump and Fitzgerald, 1992; Wannamaker et al., 2017 and ref. therein). Based on new apatite fission track data, Lisker and Läufer (2013) postulate a totally vanished Mesozoic basin spanning the Transantarctic Mountains in Victoria Land and the Wilkes Basin.

The cause of Transantarctic Mountains uplift and denudation is the subject of continuing debate, the reconstruction of Cenozoic tectonic processes being complicated by the complex interplay between a number of factors including the regional plate geodynamics, rifting style, erosion rates, subsidence and formation of thick sedimentary layers, the volcanic activity and the glacial processes. The possible mechanisms for the uplift (Fig. 6.18) include thermal buoyancy due to conductive or advective heating from the extended upper mantle of the hotter West Antarctic lithosphere (Stern and ten Brink, 1989; ten Brink and Stern, 1992), simple shear extension (Fitzgerald et al., 1986), isostatic rebound due to stretching of the lithosphere

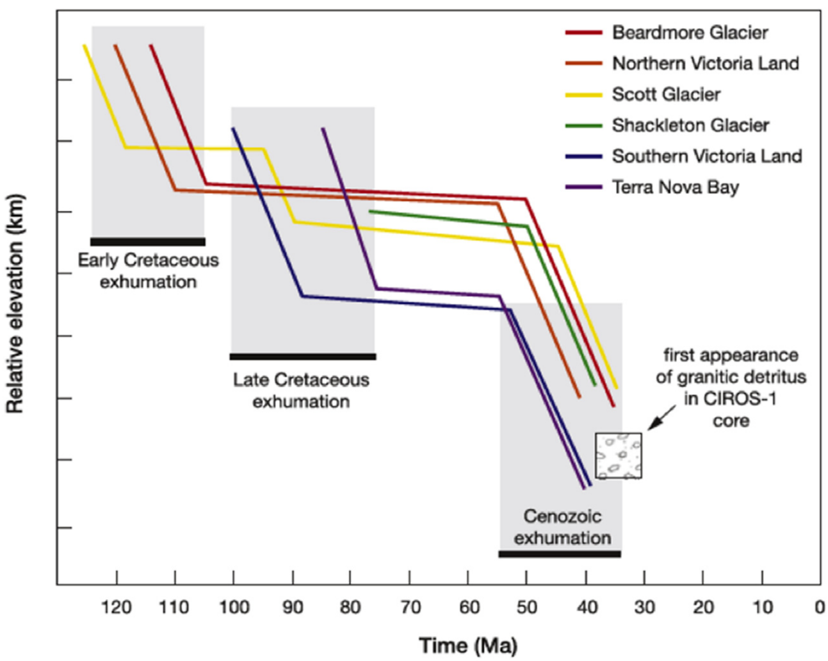


FIGURE 6.17 Summary of exhumation patterns for different areas along the Transantarctic Mountains, obtained from apatite fission track cooling ages (after Fitzgerald, 2002). *BDM*, Beardmore Glacier area; *NVL*, northern Victoria Land; *SCG*, Scott Glacier area; *SHG*, Shackleton Glacier area; *SVL*, southern Victoria Land; *TNB*, Terra Nova Bay.

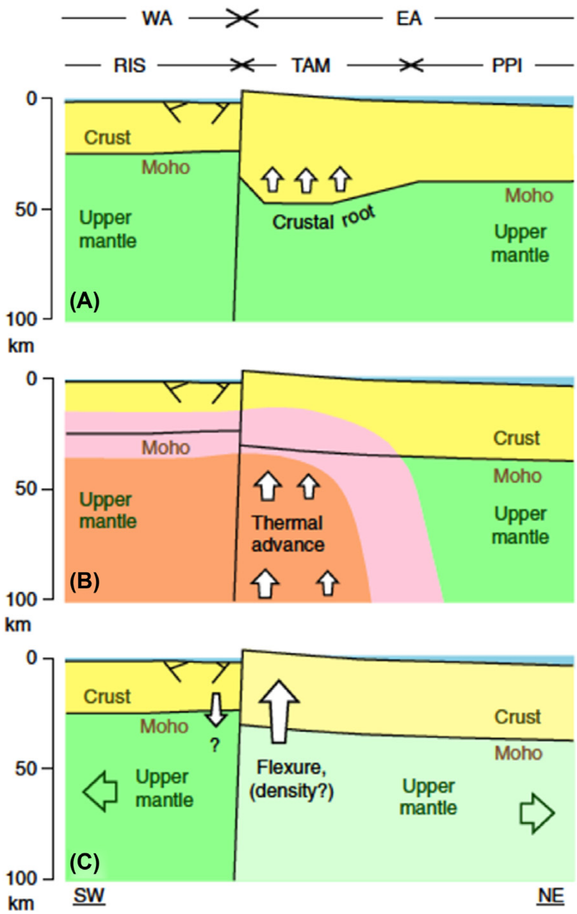


FIGURE 6.18 Hypothetical uplift mechanisms for Transantarctic Mountains rift shoulder (after Wannamaker et al., 2017). These are: (A) buoyant uplift via low-density crustal root; (B) uplift via lateral heating, thermal expansion, and possible melting; and (C) uplift via other mechanisms such as lithospheric cantilevered flexure with or without regional density contrasts. Physiographic regions include West Antarctica (WA), East Antarctica (EA), Ross Ice Shelf (RIS), Transantarctic Mountains (TAM), and Polar Plateau (PPI). Diagram not to scale.

through normal faulting (Bott and Stern, 1992), plastic necking (Chery et al., 1992), elastic necking (Van der Beek et al., 1994) and rebound in response to erosion (Stern and ten Brink, 1989). In the McMurdo Sound area, some of these mechanisms are based on specific assumptions about the crustal and upper mantle structure beneath the 'Transantarctic Mountains Front', as well as about the timing of the rift-related processes in the nearby VLB. Some constraints have already been provided by gravity studies (e.g., Davey and Cooper, 1987; Reitmayr, 1997), and by seismic reflection data (Della Vedova et al., 1997;

O'Connell and Stepp, 1993). Data from the ACRUP seismic experiment indicate thickening crust beneath the Transantarctic Mountains to a depth of 38 km, and quite low P-wave velocities (7.6–7.7 km/s) in the mantle beneath the VLB (Della Vedova et al., 1997), while low S-wave velocities are inferred at 60–160 km depth from surface wave analysis (Bannister et al., 1999), suggesting that the upper mantle is anomalously warm at that depth. More recently, the findings of drilling projects in the McMurdo Sound area (CIROS, Cape Roberts Project) have resolutely constrained the age of the onset of subsidence at the westernmost margin of the Victoria Land Basin. The first direct geological evidence of a major pre-Oligocene uplift phase of the Transantarctic Mountains comes from the oldest strata cored in the CIROS-1 and CRP-3 drill-holes (Barrett et al., 1989, 2001). These include granitic clasts eroded from exposed basement to the west, implying that the Transantarctic Mountains were at least half of their present height by then, as erosion had cut through more than 2000 m of Devonian–Jurassic Gondwana cover beds to basement (Barrett et al., 1989, 2001). In the Cape Roberts drill core, the presence of the Devonian Arena Formation (Beacon Supergroup) as bedrock beneath the Cenozoic sediments indicates that significant uplift and unroofing of the Transantarctic Mountains must have occurred prior to the Oligocene (Barrett et al., 2001).

6.6 Tectonic evolution in the Ross Sea Sector during the Cenozoic

The Ross embayment is one of the most striking morphological expressions of the WARS; a region of thin and, by inference, extended continental crust whose regional boundaries are difficult to define precisely and may have been different for different episodes of extension (Fig. 6.5). Three major episodes of extension have been proposed, with rifting starting in the Middle Jurassic, coincident with the onset of Gondwana breakup and the associated Middle Jurassic magmatism (Ferrar Group), and with subsequent episodes in the Cretaceous and late Cenozoic. The active rifting may continue to the present as suggested by the active volcanoes along the western margin of the Ross Sea. However, most extension and thinning of the Ross Sea crust is considered to have taken place during the Mesozoic rift period (Siddoway, 2008) and the tectonic connection between the Ross and Weddell embayments is largely unclear (Behrendt et al., 1992; LeMasurier, 2007, and ref. therein). In particular, our knowledge of the Jurassic episode is problematic since, although the location of rifting is well constrained in the South Africa – Queen Maud Land region, neither the location nor amount of extension is well known in the Transantarctic Mountains and Ross Sea. By comparison, the Cretaceous episode is better documented (LeMasurier, 2007, and ref. therein) and is considered to be closely related to the stretching and rifting events in the Weddell Sea region, where disruption of the plate margin involved the rotation and translation of the several crustal blocks forming the present West

Antarctica, including the Ellsworth/Whitmore Block, which was translated to the boundary between the Ross and Weddell embayments (Jordan et al., 2013, 2014, 2017, 2020).

Since movement of the West Antarctic crustal blocks was largely completed by 110 Ma, later tectonic activity in the WARS has been restricted to the Ross sector of the Transantarctic Mountains and the margin of the Ellsworth–Whitmore Mountains, with the continuation of the Transantarctic Mountains into the Weddell Sea region considered a remnant Jurassic rift (LeMasurier, 2007; Schmidt and Rowley, 1986). The first direct evidence of rifting in the Ross embayment is documented in Marie Byrd Land where mafic dike arrays (Siddoway et al., 2005; Storey et al., 1999) and A-sub-type granites (‘Byrd Coast Granite’; Wever et al., 1994) were emplaced as early as 115,5 ± 3,7 Ma, leading to increased magmatic activity until 97 ± 2 Ma (Brown et al., 2016; Siddoway et al., 2004). Considering Bradshaw’s (1989) suggestion that a sea-floor spreading centre intersected the subduction margin of Mesozoic Zealandia (within Gondwana) around 105 Ma, the production of the mafic dikes likely reflect regional extension in the overriding plate, followed by backarc felsic magmatism at 102–97 Ma. On the basis of the age of magnetic anomaly 33 (83 Ma) identified off Campbell Plateau, sea-floor spreading commenced between Campbell Plateau and Marie Byrd Land at c. 85 Ma. The reconstruction of the Campbell Plateau against Marie Byrd Land and the remarkable match of the deduced ocean–continent boundary, as well as the alignment of the eastern edge of the Eastern Ross Basin with the northern edge of the Campbell Basin on the Campbell Plateau, are evidence that the entire Eastern Ross Basin experienced extension prior to initiation of the Pacific–Antarctic Ridge at c. 85 Ma (Lawver et al., 1992).

The plate tectonic reconstruction presented by Jordan et al. (2020) (Fig. 6.19), shows the motion of West Antarctic blocks (Marie Byrd Land, Thurston Island, the Antarctic Peninsula and the Ellsworth–Whitmore Mountains) from the Middle Jurassic initiation of Gondwana breakup (Elliot et al., 2015; Jordan et al., 2017), through the separation of Southern Africa (König and Jokat, 2006), the initiation of Marie Byrd Land, Ross Sea and West Antarctic rift system extension (Lawver et al., 1992), and the separation of Zealandia (Mortimer et al., 2019), to the final stages of West Antarctic rift development linked to extension in the Adare Trough (Davey et al., 2016).

As evident in Fig. 6.20 the Pacific–Antarctic spreading centre developed in late Cenozoic time. With its initiation, a triple junction developed with three extensional arms (Australian–Antarctic spreading centre, Australian–Pacific arm and the Pacific–Antarctic spreading). Early seafloor spreading anomalies between Tasmania and the South Tasman Rise provide age control. Key tectonic events include: (1) the Australia–Antarctica spreading centre about magnetic anomaly 30 (c. 65 Ma) jumped to a position between South Tasman Rise and

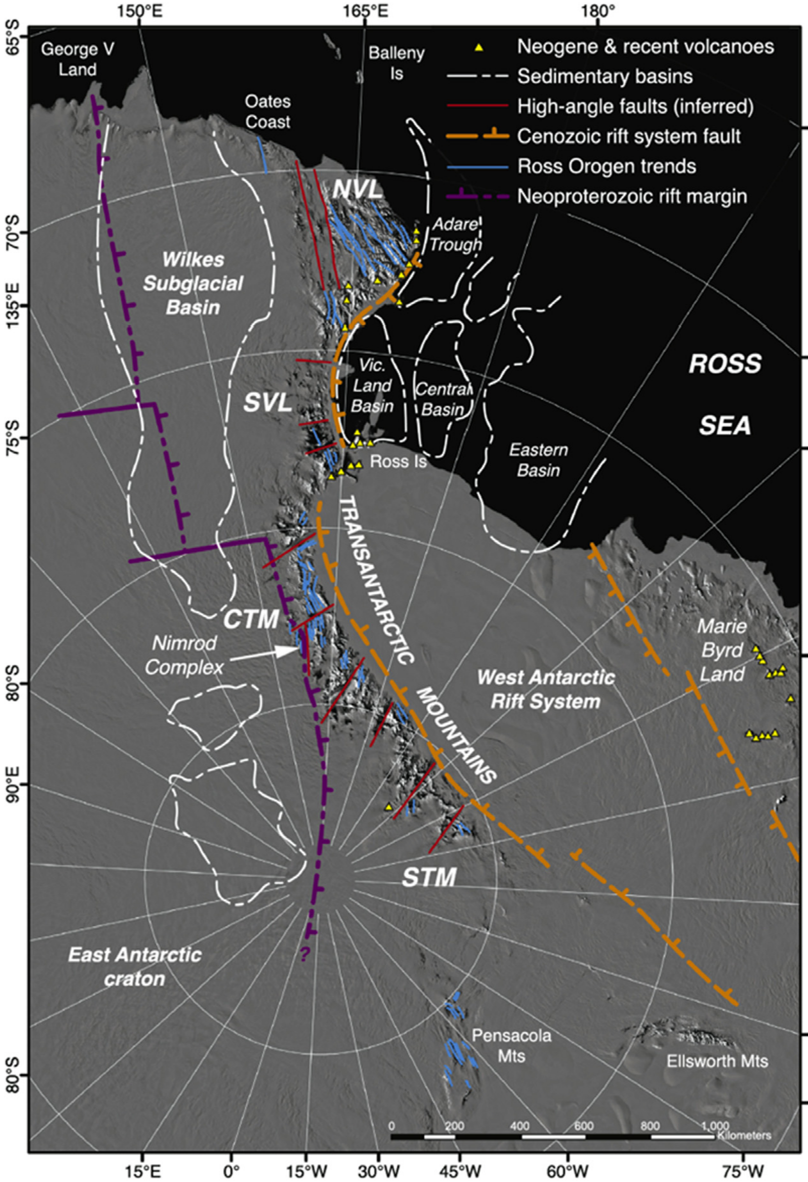


FIGURE 6.19 Tectonic map of the Transantarctic Mountains (TAM) and surrounding areas (after Goodge 2020). Neoproterozoic rift margin boundary is inferred from magnetic, gravity and seismic geophysical data (Goodge and Finn, 2010), and marks the Ross margin of the Precambrian East Antarctic shield. Neogene faults bounding the West Antarctic Rift System (WARS; from Wilson, 1999) also form the TAM front. High-angle faults inferred to underlie the major TAM outlet glaciers are related to movement on the TAM frontal fault system, but they may also be locally reactivated from Neoproterozoic transfer faults. Sedimentary basins in the Ross Sea area are related to opening of the WARS, but the origin of the interior basins such as the Wilkes Subglacial Basin is enigmatic. Neogene and recent volcanoes in the TAM and Marie Byrd Land are related to extension in the WARS, including the active systems on Ross Island.

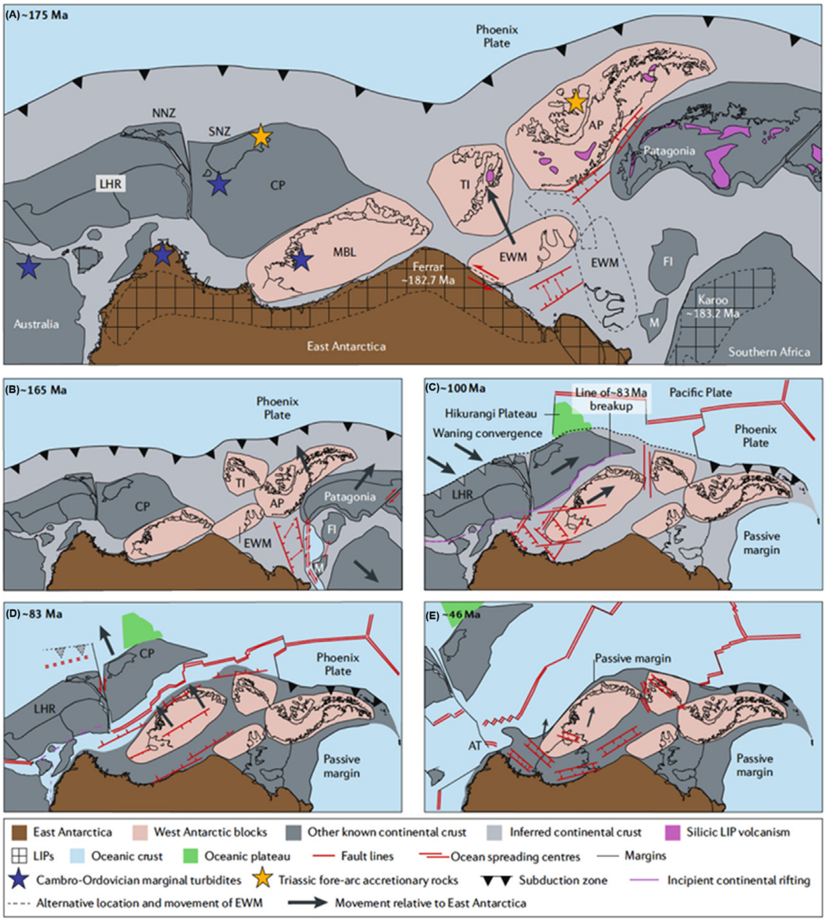


FIGURE 6.20 Western Antarctic plate tectonic reconstruction from 175 to 45 Ma (after Jordan et al., 2020), showing motion in an East Antarctica fixed reference frame. West Antarctic blocks, including Marie Byrd Land (MBL), Thurston Island (TI), the Antarctic Peninsula (AP) and the Ellsworth−Whitmore Mountains (EWM), are depicted as pale pink regions and modern coastlines are shown for reference. Other known present-day continental regions are shown in dark grey. Inferred intervening areas of continental crust are shown in light grey; such regions have been deformed and distorted to such an extent that they cannot be meaningfully traced through time. (A) Middle Jurassic initiation of Gondwana breakup is shown; the hashed regions mark Jurassic mafic large igneous provinces (LIPs) and the dark-pink regions mark silicic LIP volcanism. Cambro-Ordovician marginal turbidites (blue stars) were widespread, whereas Triassic fore-arc accretionary rocks (orange stars) were rare. The dashed outline marks alternative location and rotation of the EWM. (B) Separation of Southern Africa. (C) Initiation of MBL, Ross Sea and West Antarctic rift system extension. (D) Separation of Zealandia. (E) Final stages of West Antarctic rift development linked to extension in the Adare Trough (AT). *CP*, Campbell Plateau; *FI*, Falkland Island Plateau; *LHR*, Lord Howe Rise; *M*, Maurice Ewing Bank; *NNZ*, North Island New Zealand; *SNZ*, South Island New Zealand.

northern Victoria Land; (2) the cessation of sea-floor spreading in the Tasman Sea at magnetic anomaly 24 time (c. 54 Ma) as a consequence of the reformation of the ridge–ridge–ridge triple junction between Campbell Plateau and the Tasman Sea, and (3) the c. 40 km wide Adare Rift opened within 'older' sea-floor north-west of the Ross Sea (after magnetic anomaly 13; 33 Ma) (Cande and Stock, 2006; Cande et al., 2000) and up to 11 Ma (Granot et al., 2018). The Ross Sea rift basins include from E to W (Cooper et al., 1987; Davey and De Santis, 2006; Fielding et al., 2006; Jordan et al., 2020): the Eastern Basin – underlying most of eastern Ross-Sea; the Central Trough – running north–south discontinuously through central Ross Sea; the Northern Basin – underlying the north-eastern Ross Sea margin; and the Victoria Land Basin (VLB) underlying south-western Ross Sea, adjacent to the Transantarctic Mountains. The VLB is one of these four sedimentary basins developed due to rifting of previously thinned continental crust, along pre-existing crustal faults. Localized extension reduced the crustal thickness in some sectors to less than 10 Km. The VLB (the westernmost basin) is marked by high heat flow and alkaline volcanism (Behrendt et al., 1993; Blackman et al., 1987a,b; Cooper et al., 1991; LeMasurier, 2007; White, 1989). Extensive late Cenozoic volcanism is also inferred from aeromagnetic data. Terror Rift is located along the western edge of the VLB in the western Ross Sea (Sauli et al., 2021). The Ross Embayment has been extensively investigated through numerous airborne and marine-based geophysical surveys, highlighting the complex structural setting and tectonic evolution of the western Ross Sea (Davey et al., 2016; Ferraccioli et al., 2009; Fielding et al., 2008; Granot and Dyment, 2018; Granot et al., 2013). There is a lack of appropriate age data for the main seismo-stratigraphical units, inferences on the timing of Cenozoic rifting in the entire Ross Sea remain somewhat speculative. The results of the Cape Roberts Drilling Project (CRP) (Barrett et al., 2001) show that the VLB is mostly late Eocene or younger in age. Records recovered by drilling on the western margin of this basin indicated the onset of subsidence at about 34 Ma, significantly younger than the onset of uplift of the adjacent Transantarctic Mountains (about 55 Ma) (Fitzgerald, 1992) or before (Goodge, 2020). Hence there is a discrepancy in the relationship between uplift of the rift margin mountains and the subsidence of the adjacent rift basin. The fact that the VLB basin extends only from Ross Island to Terra Nova Bay, whereas the Transantarctic Mountains continue further north and south, would indicate that the basin may have originated from some processes other than simple extension. A transtension or 'pull-apart' process has been suggested, with the igneous activity within the 'Polar Three' anomaly (near Coulman Island) providing the transfer mechanism in the north and magnetic anomalies south of the Ross Sea Fault (Bosum et al., 1989).

Another significant geotectonic component in the Ross Sea area is represented by the extensive alkalic McMurdo magmatic province; one of the largest in the world and including two active volcanoes (Mt. Erebus at Ross

Island and Mt. Melbourne on the Ross Sea coast in northern Victoria Land; LeMasurier and Thomson, 1990). Volcanic rocks, either exposed or suggested by aeromagnetic studies in the area under the Ross Sea and West Antarctic Ice Sheet (Behrendt et al., 2002; Bell et al., 2006), occur on either side of the WARS in Marie Byrd Land (West Antarctica) and in the western Ross Sea. Earliest volcanic rocks in the western Ross Sea are Eocene to Oligocene alkali intrusive rocks ('Meander Intrusive'), interpreted to be the eroded remnants of subvolcanic magmatic complexes (Müller et al., 1991; Rocchi et al., 1999, 2002). Elsewhere, the alkaline magmatism is predominantly volcanic and has been subdivided into the informal, but geographic and petrologic distinct, Hallett, Melbourne and Erebus volcanic provinces (Kyle, 1990b). Two small isolated occurrences of basalts dated at 16−20 Ma occur around the head of the Scott Glacier (Stump et al., 1980). In the Erebus volcanic province (Kyle, 1990a), there is a continuous eruptive sequence from 19 Ma to the present day, with the Mt. Erebus crater the site of a persistent anorthoclase phonolite lava lake. The oldest exposed rocks of this province occur on the northern slopes of Mt. Morning where 19 Ma trachyandesite lavas are intruded by 16−18 Ma trachyte dikes. Volcanic ash (tephra) layers within CRP drill cores provide evidence of older Miocene trachytic eruptions. The tectonic setting suggests that the southern extent of the volcanism may be controlled by a major transfer fault, which coincides with the southern boundary of the Terror Rift (Wilson, 1999). Many of the larger volcanic centres have a radial distribution around Mt. Discovery or Mt. Erebus. The radial distribution is interpreted as a result of upwelling of a mantle plume (the Erebus plume), which was located under Mt. Discovery prior to 4 Ma and then migrated to its present position under Mt. Erebus (Kyle, 1990a).

6.7 Concluding remarks, open problems and potential research themes for future geoscience investigations in Antarctica

6.7.1 Persistent challenges for onshore geoscience investigations

In contrast to all other continents, Antarctica and its rocks, and thus its geological structures, are covered by the gigantic Antarctic Ice Sheet, which in places is over 4700 m thick. Less than 2% of the continent is uncovered, and provides the base from which geological knowledge is established, and even these exposed rocks have not been investigated thoroughly due to their remoteness and the environmental challenges faced in deep-field research. On the other hand, more than half of the ice-covered area has been surveyed geophysically, mainly aeromagnetically and gravimetrically. Further investigations are obviously needed, and the image of the Antarctic geological structure and its history will likely change, improve and be completed in the future.

Several mountain ranges in Antarctica have been seldom visited. However, their study would decidedly improve the understanding of geologic and geotectonic connections. It is evident that there are still conspicuous gaps in our knowledge of Marie Byrd Land, the Pensacola Mountains, eastern Dronning Maud Land and East Antarctica between 60° and 120°E. Moreover, the unknown ice-covered interior needs additional studies starting from the more well-known areas. For example, tracing known geological rock complexes from their exposed areas under the ice with the help of suitable geophysical methods. The confidence we can have in geophysical interpretations declines with increasing distance from directly accessible rock complexes. Even a few spot checks of rock samples from isolated deep drill-holes would provide a better reliability of interpretations (i.e. calibration of airborne geophysical data). New data from the RAID (Rapid Access Ice Drill) on key regions of the subglacial geology will be decisive for paleogeodynamic reconstructions (Goodge and Severinghaus, 2016). Another example, as clearly reported in a recent geological review of West Antarctica (Jordan et al., 2020), is to decide whether the region is best conceived as an accreted collection of rigid microcontinental blocks (as commonly depicted) or as a plastically deforming and constantly growing melange of continental fragments and juvenile magmatic regions. Jordan et al. (2020) highlight the importance of new techniques, such as finite-element modelling, and the need to couple models with more detailed geophysical and geological studies. Geophysical data can provide new constraints on the extent of magmatism and the areal extent and geometry of the underlying provinces, and geological observations and dating can provide information about how and when the different components of the system were active.

6.7.2 Antarctica and the Ross Orogen in the Transantarctic Mountains

Other important geological research problems relate to the evolution of Rodinia and Gondwana supercontinents. Since Antarctica, and particularly East Antarctica, had a central position in both Rodinia, between 1300 and 800 million years ago, and Gondwana, between 600 and 200 million years ago (today’s southern continents), geological information archived in this ‘keystone’ region are important not only for a well-founded analysis of local conditions, but also in relation to our understanding of Earth system processes in deep time. Abundant evidence exists for the former connection between Antarctic cratonic areas and geologically similar provinces on neighboring continents.

The study of East Antarctica is a prerequisite for the reconstruction of the assembly and breakup processes of both supercontinents. Geological and paleomagnetic data show that East Antarctica comprises older cratonic fragments and Grenvillian and Pan-African structures in coastal exposures support this notion. The larger part of the so-called 'East Antarctic Craton' continues into India and Australia, while a small fragment (Grunehogna Craton) connects to the Kalahari Craton of Africa. The 'East Antarctic Craton' consists of a number of cratonic nuclei (Boger and Miller, 2004; Fitzsimons, 2000a), confirming the importance of Antarctica for the reconstruction of continental distribution in early Earth history. Although numerous alternative scenarios have been proposed for Rodinia, refined global reconstructions based on paleomagnetic data, supported by geological correlation, show the Transantarctic Mountains margin of East Antarctica, in continuity with the Neoproterozoic margin of eastern Australia, as conjugate to western Laurentia from c.1080 to c.750 Ma (so-called SWEAT model; Dalziel, 1997; Meert and Torsvik 2003; Torsvik, 2003). Recent geological studies, as reviewed by Goodge (2020), appear to confirm the cratonic linkages and Neoproterozoic rift-margin associations proposed earlier by Moores (1991) and Dalziel (1991). As stated by Goodge (2020), despite a more solid understanding of the Neoproterozoic history along the Transantarctic Mountains margin of the Mawson Craton, additional geophysical observations and research are required to resolve uncertain features, including the position of the rift margin, the geometry of rifting, the extent of crustal thinning, the extent of rift margin sedimentation, the location of possible transform offsets and the influence of these structures on younger orogens (Goodge and Finn, 2010).

Significant contributions to the understanding of global plate tectonic and geodynamic processes are also stored in the orogenic belts of Pan-African age (600−500 million years ago), which contain important information about the juxtaposition of West and East Gondwana. In Antarctica, the Shackleton Range and parts of East Antarctica between Dronning Maud Land, Lützow HolmBukta and Prydz Bay belong to these belts (Boger et al., 2002; Buggisch and Kleinschmidt, 2007; Buggisch et al., 1990; Fitzsimons, 2000b; Jacobs et al., 1998; Paech, 2005; Tessensohn et al., 1999). Equivalent rocks have been found in the African Mozambique Belt (Jacobs et al., 1998; Paech, 1985; Paech et al., 2005). So far, little is known about how these fragments continue under the ice and how they can be connected with the better researched mountains in Dronning Maud Land and the Transantarctic Mountains. Equally unknown is the formation, age and relation of subglacial mountains in the East Antarctic interior (e.g., the Gamburtsev Subglacial Mountains).

At approximately the same time as the development of internal Pan-African orogenic belts, the geographic domain corresponding to today's Transantarctic Mountains experienced the Ross Orogeny; the result of dominant accretionary tectonic processes produced by the evolution from a passive, drifting margin to an active, subducting margin. This Pan-African activity was characterised in the Transantarctic Mountains by an accretionary-type convergent plate margin along the southern outer perimeter of Gondwana. Subduction of paleo-Pacific oceanic lithosphere beneath the active Gondwana margin of continental East Antarctica spanned as much as 145 million years between about 615−470 Ma, reflecting a prolonged period of sustained underflow.

There is widespread consensus in considering the end of the Ross orogeny as a nearly synchronous event along the different sectors of the belt (e.g., Encarnacion and Grunow, 1996; Stump, 1995). Voluminous granitoids (e.g., Granite Harbour Intrusive Complex) intruded as dated batholiths at c. 560−470 Ma (Encarnacion and Grunow, 1996; Hagen-Peter and Cottle, 2016) represent a unifying feature throughout the length of the Transantarctic Mountains (Borg et al., 1990; Stump, 1995). But although the general tectonic history of the Ross Orogeny is fairly well known within each of the major segments of the Transantarctic Mountains, significant variations in lithostratigraphic, structural and metamorphic patterns, as well as in granitoid geochemical affinity, are evident between the different segments. Considerable uncertainty remains about the onset of the subduction, the tectonic setting of the early granitoids with variable chemical affinities (from calc-alkaline to alkaline and carbonatite) of southern Victoria Land, the nature of the contact between the orogenic belt and the Mawson Craton, and the relations with the Pan-African structures of the Shackleton Range.

Many other broad first-order questions still wait for future research focused on the Ross Orogeny, as acknowledged by Goodge (2020). Until our knowledge of the relationships of the tectono-metamorphic histories, and of the detailed chronology of the magmatic, tectonic, sedimentary and metamorphic episodes between the various segments of the orogenic belt is understood, a comprehensive tectonic model of the development of the Ross Orogeny remains to be formulated.

6.7.3 Antarctica after Gondwana fragmentation

The importance of Antarctica with regard to the destructive processes of plate tectonics, including the fragmentation of supercontinents is also well recognised. The young continental margins and rift structures of Antarctica, as well as their development, document the breakup of Gondwana. Aside from a small section along the Antarctic Peninsula, the Antarctic continental margins depict the fault structures of Gondwana breakup, leading ultimately to the formation of the present southern continents and oceans.

After fragmentation started and initial drifting began, Antarctica reached and remained at its south polar position since at least the Late Cretaceous (approximately 130 Ma). Its isolation from the neighbouring continents, through the early opening of the Southern Ocean and the formation of the Antarctic Circumpolar Current, also began at this time. The further breakup process led to the full formation of the Southern Ocean. The exact opening processes, and their effects on paleoceanography, have not yet been satisfactorily reconstructed because high-quality magnetic data are lacking in many key areas of the oceans. This is especially the case for the South Pacific and areas between Antarctica and Africa. For an understanding of the plate tectonic development of the Antarctic continent several events must be considered: (1) 130−100 million years ago, the opening of the Weddell, Lazarev and Riiser-Larsen Seas, and of the southern Atlantic and Indian Oceans; (2) 110−80 million years ago, the genesis of the southern Kerguelen Plateau; (3) 80−40 million years ago, the separation of Tasmania/Australia/Zealandia from Antarctica; (4) 90−14 million years ago, the development of the Ross Sea Rift; and (5) 30−20 million years ago when South America separated from Antarctica, and the Drake Passage and Scotia Sea opened, leading to the formation of the Circumpolar Current. In this time context critical data for the Ross Embayment are lacking (Goodge, 2020; Jordan et al., 2020). There is particular need to add details to the bounding structure between the Transantarctic Mountains and WARS provinces, in respect to their geometry and variability with depth. Cryptic evidence suggests the presence of fragments of the Transantarctic Mountains in Marie Byrd Land (Bradshaw, 2007) and/or beneath the Ross Ice Shelf (Tinto et al., 2019). The relationships can be better understood through integration between surface geology data and subsurface geophysical data (see also McKay et al., 2021).

Despite the many questions about the mechanisms and the consequences of these global processes (e.g., of climatic nature) it is certain that the breakup of Gondwana led to the current configuration of continental plates and the isolation of continental Antarctica. Breakup provided the starting point for many current processes and patterns, including the glaciation of the polar regions, the present-day oceanic and atmospheric circulation, climate patterns and the distribution of biota, and the global environmental conditions for human existence. To improve the present state of knowledge, the distinct progression of Gondwana breakup and global ocean circulation should be determined more precisely. The prominent mountains chains represented by the Transantarctic Mountains await further research addressing mechanisms of uplift, and regional variability of paleothermal histories that indicate diverse exhumation mechanisms, and the connection to Cenozoic denudation through tectonic and glacial process.

An increased research effort in these areas is essential to allow a better timing of the Antarctic Circumpolar Current onset (Barker and Thomas, 2004; Barker et al., 1998, 2007), to understand the links between glacial history and Transantarctic Mountains uplift, the feedbacks between glacial erosion and uplift rates, and how surface processes connect to both tectonic and climatic influences. These aspects underlie the central influence of the polar regions in general circulation models (Oglesby, 1999; Sloan et al., 1996). The thick continental ice sheets drive latitudinal gradients and sea-ice formation, leading to the formation of cold Antarctic bottom water, which through deep ocean currents reaches the lower latitudes (Carter et al., 2021), and indeed the other hemisphere. The oxygen isotope record from deep-sea cores, and eustatic changes inferred from sequence stratigraphic records on passive continental margins, have been leading paradigms for the interpretations regarding Antarctic ice sheet history. However, these global proxy records of glacio-eustasy are contradicted by geologic records in Antarctica (Harwood et al., 1991, 1993; Miller and Mabin, 1998; Moriwaki et al., 1992; Wilson, 1995). Because direct data from the Antarctic region are necessary for deep-time climate model-validation, a series of drilling projects has targeted the Antarctic continental margin to retrieve high-resolution stratigraphic records of Antarctica's glacial and climatic history (Barker et al., 1998; Barrett et al., 2000, 2001; Colleoni et al., 2021; Cooper and Webb, 1994; Gohl et al., 2021; Hambrey, 2002; Hambrey et al., 1998; Harwood et al., 2002; McKay et al., 2019; McKay et al., 2021). Paleoclimatic reconstructions based on the results of the CRP in the Ross Sea region for the early Oligocene to early Miocene time show the occurrence of ice but also a warmer climate in the Antarctic than today (Barrett et al., 2000, 2001; Hambrey et al., 1998; Naish et al., 2001). Unresolved, however, is the question of how stable the Antarctic ice sheets were during the last 20 My and the timing of the onset of glaciation. Building upon knowledge from ANDRILL 2006 and 2007 (Harwood et al., 2002), a primary objective of ongoing investigations is to decipher the responses of past ice sheets to climate forcing, including variability at a range of time-scales (e.g., Colleoni et al., 2018 and Colleoni et al., 2021). Both CIROS and CRP pursued this elusive target. Indeed, in CRP-3, Oligocene strata passed via an unconformity into the pre-glacial Devonian–Triassic Beacon Supergroup rocks, proving that sites are to be found that contain Eocene and earlier Cenozoic, or even Cretaceous records. These opportunities, along with others, will likely be identified through future drilling programs that complement Ross Sea data with data from other coastal regions of Antarctica, and will provide a deeper insight into the tectonic and climatic relevance of the Ross Embayment region.

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