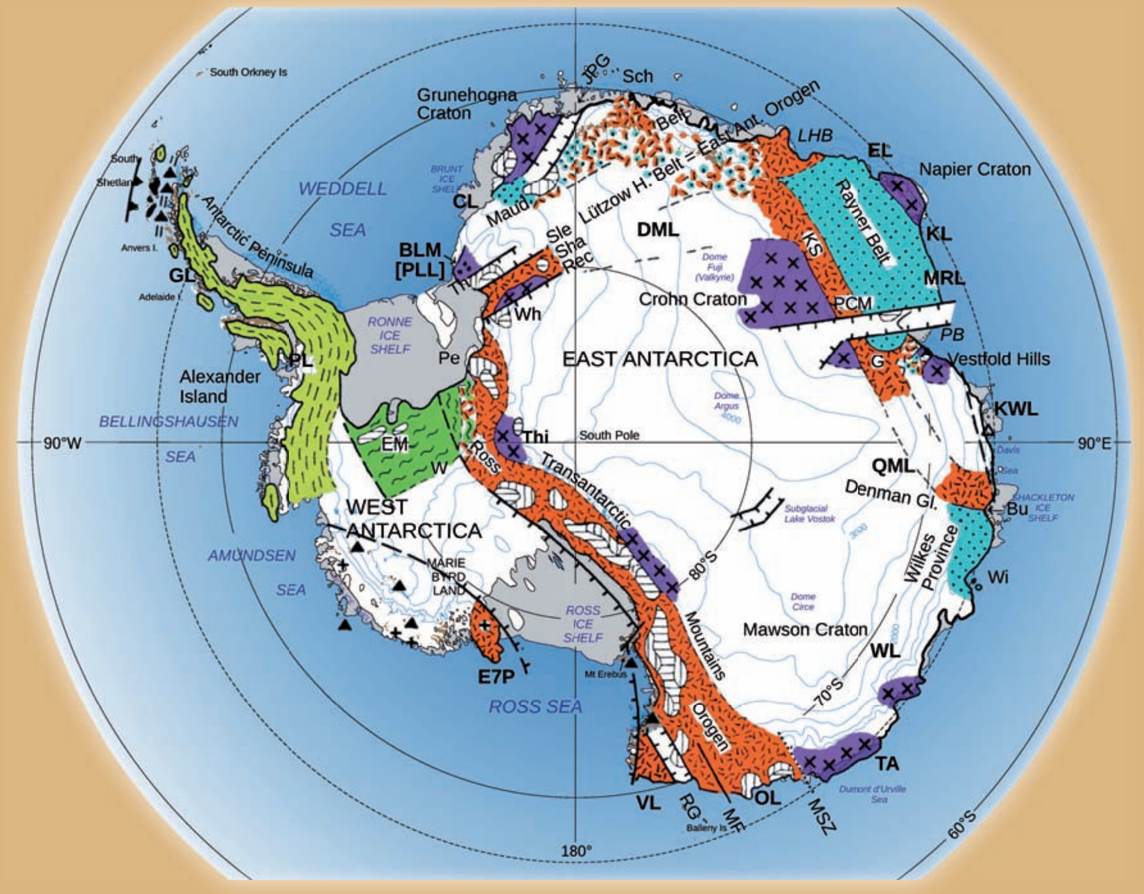
Beiträge zur regionalen Geologie der Erde **33**

Georg Kleinschmidt (Ed.)

The Geology of the Antarctic Continent



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The Geology of the Antarctic Continent

edited by
Georg Kleinschmidt

With 194 figures and 10 tables

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Georg Kleinschmidt (ed.): The Geology of the Antarctic Continent

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We would be pleased to receive your comments on the content of this book: editors@schweizerbart.de

Front cover: Geological sketch map of the Antarctic continent. Carthographic base: modified after Antarctiva.svg, Wikipedia, Westantarctica

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Preface

The idea of this book has its roots in two things:
(i) For more than twenty years, I have given lectures on the geology of Antarctica. However, unfortunately, I have not been able to recommend a suitable textbook for my lectures without reservation.
(ii) When I retired in 2003, I was goaded with: "from now on, you have enough time to write such a book". However, I was aware from the start, that I would never be able to realize that venture just by myself. Therefore, eight specialists were approached and asked for co-authorship, who I regard as the experts of particular Antarctic regions. Some of them are comprehensively experienced, some are still carrying out intense Antarctic fieldwork. Many thanks to all of them for their willingness to collaborate and for completing their contributions pretty much in time, at least almost all. Unfortunately, one of the authors whom we counted on, deserted us at the eleventh hour. Therefore we had to try saving chapter 7 (and indeed the whole book) as a four some. We hope we were successful. I have to point out one of us authors: Michael Thomson gave very useful advice in general and especially, how to handle the different completions of the chapters. Unfortunately, Michael Thomson († January 19, 2020) did not live long enough to see the final completion of this book. Then, we are grateful to Dr. Beverley Tkalcec, Frankfurt, for improving the English of the non-native English-speaking authors. Of course, we authors are as well indebted to the publishers Dr. A. Nägele and Dr. W. Obermiller and their team of Borntraeger Publishers (Stuttgart) for their patience, for accepting special requests and for supporting us in both word and deed. Special thanks is due to all Antarctic field-comrades, especially to the participants in the GANOVEX expeditions, above all to their initiator Franz Tessensohn (1939-2019). To him this book is dedicated.

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Contents

1. Introduction and Overview

Georg Kleinschmidt

1.1. Introduction

1.1.1. Antarctica: definitions, topographic and geographic outline

“The Antarctic” as a whole includes, in addition to the continent Antarctica, the Southern Ocean, or in other words: „the Antarctic“ is the polar region south of the Polar Front (previously called “Antarctic Convergence”), which fluctuates roughly around 50° south. There, the cold Antarctic waters of the Southern Ocean sink beneath the warmer waters of the Atlantic, Indian and Pacific oceans. According to international law, the northern limit of “The Antarctic” is 60° south.

The *continent Antarctica* measures ~14 x 10^6 km^2 with and ~12 x 10^6 km^2 without the ice shelves, and thus it is the third-largest continent, i.e. it is larger than both Australia and Europe.

The larger, rather compact and roughly circular part around 90° E is called East Antarctica (= Greater Antarctica), the somewhat smaller and more indented part around 90° W West Antarctica (= Lesser Antarctica). The boundary between East and West Antarctica does not follow the 0°/180° meridians but, by definition, the eastern shoreline of the Weddell Sea and the western shoreline of the Ross Sea, which correspond very roughly with the meridians of 20° W and 170° E, respectively. Therefore strictly speaking, the Transantarctic Mountains belong to East Antarctica (Fig. 1-1). The consequence of this definition is, that funnily enough East Antarctica is to the west of 180° and West Antarctica is to the east of 180°.

More than 99% of Antarctica is ice covered, only 0.32% is exposed (Clarkson 2007, Vaughan 2007). Exposures are concentrated on the Antarctic Peninsula, along the coastal areas and in the high mountain ranges that protrude through the ice cover, e.g. Ellsworth Mountains, Transantarctic Mountains, Mühlig-Hoffmann-Gebirge/Wohlthatmassiv, Prince Charles Mountains (Fig. 1-1). The highest elevation of Antarctica is the Vinson Massiv of the Ellsworth Mountains (4897 m); and there are eight four-thousand-metre mountains in the Transantarctic Mountains, and the ice plateau of central East Antarctica reaches 4093 m elevation at Dome Argus (81° S/77° E). The geological knowledge concerning Antarctica is based first on these outcrops, second on geophysical data and on interpolations.

The mountain ranges are characterised by alpine glaciers, the largest of which are the largest in the world: the Lambert Glacier (>550 km long, up to 100 km wide), the Rennick Glacier and the Jutulstraumen (length and width of both ~300 km/~40 km respectively). However, only few glaciers flow directly into the Southern Ocean (from the three mentioned just the Rennick Glacier). The Antarctic ice sheet drains away mainly in the form of ice shelves. The largest of them are:

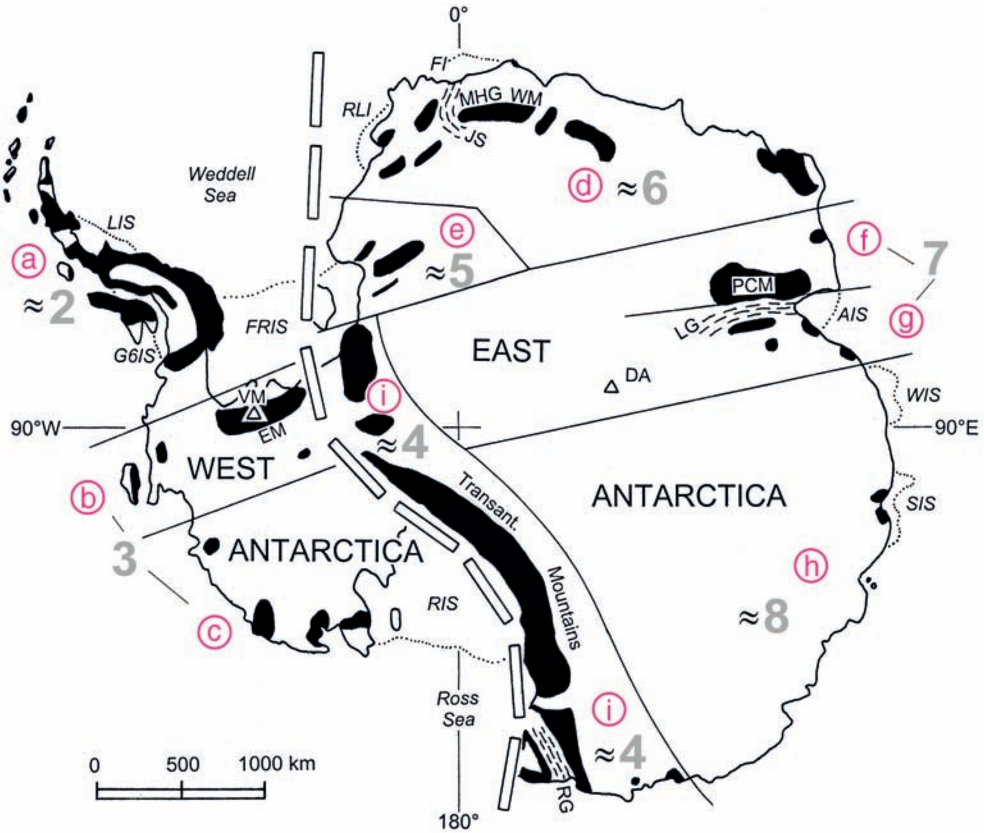


Fig. 1-1. Antarctica: Definitions, physiographic regions, regional distribution of chapters, selection of topographic features. Thick broken line is roughly marking the boundary between West and East Antarctica. Black = areas containing most of the exposures. Physiographic regions: (a) Antarctic Peninsula, (b) Ellsworth Mountains, Ellsworth Land, (c) Marie Byrd Land, (d) Dronning Maud Land, (e) Shackleton Range (+ adjacent areas), (f) Prince Charles Mountains (+ adjacent areas), (g) Lambert Glacier (+ adjacent areas), (h) rest of East Antarctica between ~80° E and ~145° E, (i) Transantarcic Mountains. Grey numerals indicate the **chapters (2-8)**: Region (a) is covered by **chapter 2**; regions (b) and (c) are covered by **chapter 3**; region (d) is covered by **chapter 6**, region (e) by **chapter 5**; regions (f) and (g) are coverd by **chapter 7**; region (h) is covered by **chapter 8** and region (i) by **chapter 4**. Abbreviations of topographic features: Main mountain ranges and mountains: DA = Dome Argus (triangle, 4093 m), EM Ellsworth Mountains, MHG = Mühlig-Hoffmann-Gebirge, PCM = Prince Charles Mountains, VM = Vinson Massif (triangle, 4897 m), WM = Wohlthatmassiv. The three largest glaciers (dashed): JS = Jutulstraumen, LG = Lambert Glacier, RG = Rennick Glacier. Largest ice shelves (dotted): AIS = Amery Ice Shelf, FI = Fimbulisen, FRIS = Filchner-Ronne Ice Shelf, G6IS = George VI Ice Shelf, LIS = Larsen Ice Shelf, RLI = Riiser-Larsenisen, RIS = Ross Ice Shelf, SIS = Shackleton Ice Shelf, WIS = West Ice Shelf.

Ross Ice Shelf	~510 000 km^2
Flichner-Ronne Ice Shelf	~440 000 km^2
Amery Ice Shelf (69° S/72° E)	~63 000 km^2
Larsen Ice Shelf (mainly Larsen C) (67°30’ S/62°30’ W)	~49 000 km^2
Riiser-Larsenisen (73° S/16° W)	~48 000 km^2

Fimbulisen (70°30’/0°)	~41 000 km^2
Shackleton Ice Shelf (66° S/100° E)	~34 000 km^2
George VI Ice Shelf (71°45’ S/68° W)	~24 000 km^2
West Ice Shelf (67° S/85° E)	~16 000 km^2

Others are: Abbot Ice Shelf (72°45’ S/96° W), Getz Ice Shelf (72°15’ S/125° W), Sulzberger Ice Shelf (77° S/148° W), Brunt Ice Shelf (74°45’ S/22°30’ W), Ekströmisen (67° S/8° W).

The ice cover reaches up to 4776 m in thickness (average: 2126 m, 1937 m including ice shelves; Fretwell et al. 2013). The ice volume amounts to 26.92 x 10^6 km^3 (Fretwell et al. 2013), representing about 90% of the earth’s fresh water. This mass of ice pushes the entire continent into the mantle of the earth. As a consequence, the continental shelf of Antarctica is situated ~500 m (400 to 600 m) below sea level, i.e. clearly deeper than the shelves of all other ordinary continents (200 km below sea level; Leier 2007). If the Antarctic ice cover was removed rapidly, the continent would be for the most part under water, – it would be an archipelago.

1.1.2. Previous appropriate books

There are only very few scientific books on the geology of Antarctica, actually just two:

1. Otto Nordenskjöld: “Antarktis”, 29 pp (!), Heidelberg 1913 (!) and
2. Robert J. Tingey: “The geology of Antarctica”, 680 pp, Oxford 1991a.

Many works on Antarctica in general contain chapters dealing with the fundamentals of the Antarctic geology, e.g. the Encyclopedia by B. Riffenburgh (2007), the „Discussion“ by Sir V. Fuchs & R.M. Laws (1977), the climatological volume by F. Florindo & M. Siegert (2009) and the German book by N.W. Roland (2009).

Therefore, since the only comprehensive book, i.e. that by R.J. Tingey and his co-authors, was published more than 20 years ago, we see the necessity to publish a textbook,

– containing the latest development of geological knowledge,
– taking into account the International Polar Year (IPY) 2007-08(-09),
– based on extensive fieldwork of the 9 authors.

1.1.3. The book’s main concept

Antarctica can be subdivided into the following nine physiographical regions (Fig. 1-1):

West Antarctica, consisting of
(a) the Antarctic Peninsula,
(b) the Ellsworth Mountains region (≈ Ellsworth Land),
(c) Marie Byrd Land

East Antarctica, consisting of
(d) Dronning Maud Land,
(e) the Shackleton Range and adjacent areas,
(f) Prince Charles Mountains,

(g) the Lambert Glacier (Lambert Rift, Lambert Graben),
(h) the rest of East Antarctica including its most interior and totally ice-covered part,

and approximately separating both West and East Antarctica:
(i) the Transantarctic Mountains (even if they – strictly speaking – have to be included into East Antarctica).

From that subdivision, the following single chapters of this book emerged almost automatically:

1. Introduction and Overview (by the editor);
2. Antarctic Peninsula (by J.L. Smellie, Leicester, U.K.);
3. West Antarctica (Marie Byrd Land, West Antarctic Rift System, Ellsworth Mts. (by C. Siddoway, Colorado Springs, USA);
4. Transantarctic Mountains (by J.W. Goodge, Duluth, USA);
5. Shackleton Range (by G. Kleinschmidt, Frankfurt a.M., Germany);
6. Dronning Maud Land (by Andreas L. Läufer, Hannover, Germany);
7. Lambert Glacier area (East Antarctica between 45° E and 85° E) (by A.L. Läufer, Hannover, F. Lisker, Bremen, N.W. Roland, Hannover/Burgwedel, and G. Kleinschmidt, Frankfurt, all Germany);
8. East Antarctica (main part, i.e. between 85° E and 145° E) (by P.R. Ménot, Saint Étienne, France);
9. Mineral Resources (by N.W. Roland, Hannover, Germany);
10. Fossil Record (by M.R.A. Thomson, Leeds, U.K.).

Our book is restricted to the Antarctic *continent* including the continental shelf and related islands, but it does not deal with the Southern Ocean – that would end up in another voluminous tome. Thus, the South Orkney Islands, the South Shetland Islands, Adelaide, Alexander, Thurston and Scott Islands will be taken into consideration, but not, or at the most scarcely, South Georgia, the South Sandwich Islands, the Îles Kerguelen, Heard and Macquarie Islands, and the Balleny Islands.

Even though ice is *the* main characteristic of the Antarctic and could be regarded as the most conspicuous rock type of Antarctica and even though the ice shelves are mainly included in outline maps, Antarctic ice or, to put it more precisely, the glaciology of Antarctica will be touched only peripherally in this book.

Furthermore, the book tries to go along with the toponymic guidelines of SCAR (1994). Nevertheless, there may occur a few inconsistencies, e.g. some authors use "Oates Coast" instead of "Oates Land" (or vice versa).

1.2. Geological Overview (Fig. 1-2)

Like all other continents, the Antarctic continent consists of ancient cores – called old cratons, just cratons or nuclei – and of relatively younger mobile belts – orogens. Originally, i.e. in the pre-plate-tectonic phase of geological science, the cratons were thought to be relatively rigid parts of continents and older than Cambrian. Meanwhile, the term has changed somewhat concerning the age: It is used nowadays for continental cores of Archaean age, but in Antarctica cratonic ages down to 1.5 Ga are still common. Moreover, it should be reminded that a craton may consist of an exposed basement called "shield", and of an undeformed, unmetamorphosed, rather old, sedimentary cover called "platform". Anyhow, "craton" is rather a makeshift term, as increasingly older

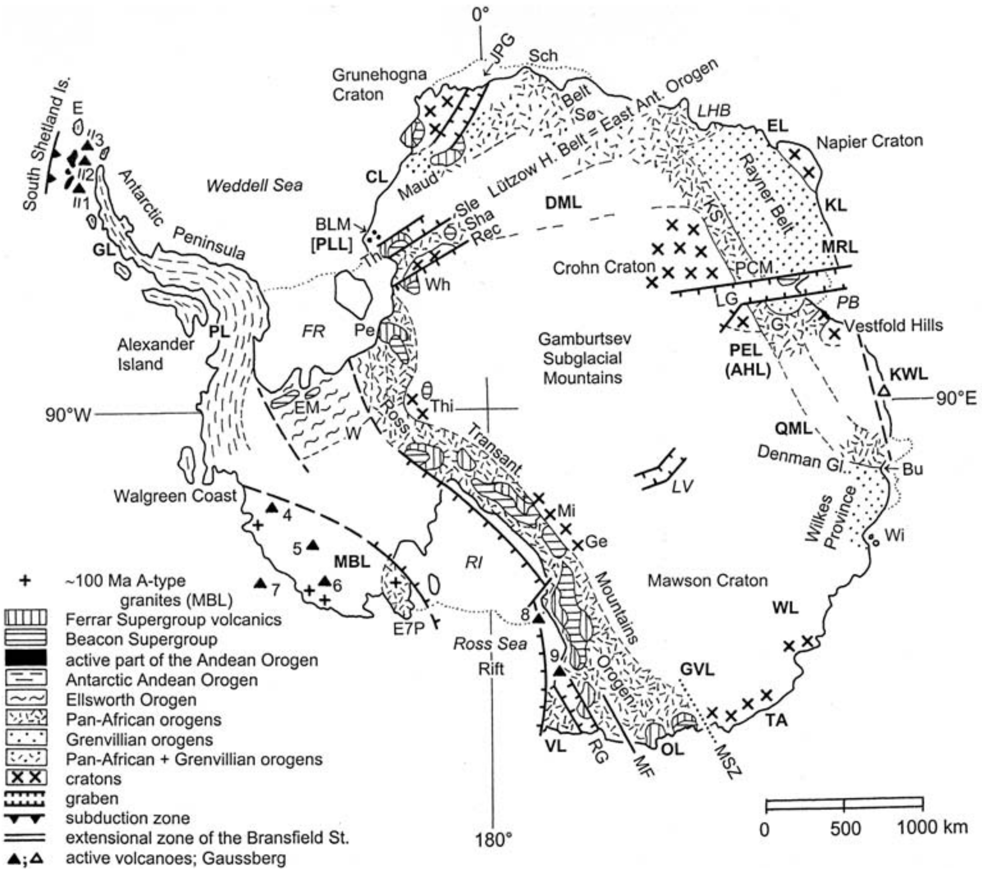


Fig. 1-2. The geological architecture of Antarctica (modified from Kleinschmidt 2014). Abbreviations: **AHL** = American Highland, BLM = Bertrab-, Littlewood-, Moltke-Nunatakker, Bu = Bunger Hills, **CL** = Coats Land, **DML** = Dronning Maud Land, E = Elephant Island, **EL** = Enderby Land, EM = Ellworth Mountains, E7P = Edward VIIth Peninsula, FR = Filchner Ronne Ice Shelf, G = Grove Mountains, Ge = Geologists Range, **GL** = Graham Land, **GVL** = George Vth Land, JPG = Jutul Penck Graben, **KL** = Kemp Land, KS = (Antarctic) Kuunga Suture, **KWL** = Kaiser-Wilhelm-II.-Land, LG = Lambert Graben, LHB = Lützow-Holmbukta, LV = Lake Vostok, **MBL** = Marie Byrd Land, MF = Matusevich strike-slip fault, Mi = Miller Range, **MRL** = Mac.Robertson Land, MSZ = Mertz Shear Zone, **OL** = Oates Land, PB = Prydz Bay, PCM = Prince Charles Mountains, Pe = Pensacola Mountains, **PEL** = Princess Elizabeth Land, **PL** = Palmer Land, **PLL** = Prinzregent-Luitpold-Land, **QML** = Queen Mary Land, Rec = Recovery Glacier (fault), RG = Rennick Graben, RI = Ross Ice Shelf, Sch = Schirmacheroase, Sha = Shackleton Range, Sle = Slessor Glacier (fault), Sø = Sør-Rondane, **TA** = Terre Adélie, Th = Theron Mountains, Thi = Thiel Mountains, **VL** = Victoria Land, W = Whitmore Mountains, Wh = Whichaway Nunataks, Wi = Windmill Islands, **WL** = Wilkes Land. Active volcanoes: 1 Deception Island, 2 Penguin Island, 3 Bridgeman Island, 4 Mt. Takahe, 5 Mt. Sidley, 6 Mt. Berlin, 7 Mt. Siple, 8 Mt. Erebus, 9 Mt. Melbourne.

mobile belts (= orogens) are definable and therefore distinguishable from a craton. Thus, Kimban orogens (about 1.6 Ga old) are just being identified and distinguished from the cratonic areas of Antarctica.

The post-cratonic orogens in Antarctica are: 1. Grenvillian orogens (about 1.1 Ga old); 2. Pan-African orogens (about 600 to 500 Ma old); 3. the Ellsworth Orogen (about 200 Ma old); 4. the Andean Orogen (about 150 to 100 to 40 Ma old, extending into still active plate-tectonic processes).

1.2.1. The Antarctic cratons

Until the nineties, East Antarctica was thought to form one uniform large craton called "East Antarctic Craton" or – meaning the same – "East Antarctic Shield" (e.g. Craddock 1972a, Grikurov & Dibner 1979, Borg et al. 1990). However, a single East Antarctic Craton/East Antarctic Shield does not exist. Instead, East Antarctica consists of several smaller cratons separated by different mobile belts. The most important of these smaller cratons are: 1. the Grunehogna Craton; 2. the Napier Craton; 3. the Mawson Craton; and 4. the Crohn Craton (≈ Ruker Craton). Therefore, the terms "East Antarctic Craton" and "East Antarctic Shield" should no longer be used (although "East Antarctic Shield" may be acceptable just in a geomorphological sense). Nevertheless, some of our authors use the terms "East Antarctic Shield" or even "East Antarctic Craton" for the sake of simplicity.

1.2.1.1. The Grunehogna Craton

The little Grunehogna Craton makes up westernmost Dronning Maud Land. There, mainly 1 Ga old platform sediments are totally flat-lying and undeformed (Fig. 1-3). They cover a 3 Ga old granitic basement, which peeps out just spotlike at Annandagstoppane (ca. 6° W/72°30' S). Only this part of the Grunehogna Craton may be called "shield". There U-Pb zircon ages fix the crystallization of the granite at 3067±8 Ma (Marschall et al. 2010).

Fig. 1-3. The Grunehogna Craton. Cratonic platform: flat lying, undeformed sedimentary rocks more than 1 Ga old at the former South African summer station Grunehogna.

The Grunehogna Craton forms just a little fragment of the Kalahari-/Kaapvaal Craton in southern Africa.

1.2.1.2. The Napier Craton

The Napier Craton (alias Napier Complex) of Enderby Land is a small fragment of the Indian Dharwar Craton. Its highly metamorphosed rocks (ortho- and paragneisses, granulites, charnockites incl. enderbites!) yielded the greatest radiometric ages of Antarctica at all (~3.8, ~3.0, ~2.8, ~2.5 Ga: Harley & Black 1997).

1.2.1.3. The Mawson Craton

The Mawson Craton (alias Mawson Block or Mawson Continent: Fanning et al. 1995, Fitzsimons 2000a) is the largest of the East Antarctic cratonic realms. It is nearly completely covered by ice and is exposed only at the coastlines of Wilkes Land, Terre Adélie and George V Land, i.e. opposite Australia, and at the rear of the Transantarctic Mountains in the Geologists Range, in the Miller Range and in the eastern Thiel Mountains. Its highly metamorphosed shield areas consist of gneisses and granulites, which are 2.5 and 1.7 Ga old. Only the eastern Thiel Mountains form a platform covered by relatively flat-lying Neoproterozoic, Cambrian and Ordovician sedimentary rocks. According to Fanning et al. (1995) and Fitzsimons (2000a), the Mawson Craton includes as well the Australian Gawler Craton. The Antarctic part of the Mawson Craton has been named Terre Adélie Craton by R.-P. Ménot (this vol.).

1.2.1.4. The Crohn Craton (≈ Ruker Craton)

The extent of the Crohn Craton (Boger 2011) of Mac.Robertson Land and Princess Elizabeth Land (≈ American Highland) is even more speculative. There, just the southern Prince Charles Mountains on both sides of the Lambert Glacier are certainly cratonic. The main rock types are orthogneisses showing radiometric ages of just over 3 Ga (Boger et al. 2008). It is unknown how far the Crohn Craton continues into the ice-covered centre of East Antarctica.

Recently, Jacobs et al. (2015) use the term “Ruker Craton” instead of Crohn Craton and restricted to the Prince Charles Mountains. However, according to Boger (2011) the Crohn Craton consists in addition to the Archaean basement complex called “Ruker Complex” or “Ruker Terrane” (Boger 2011, Boger et al. 2006, Mikhalsky et al. 2006a, Phillips et al. 2006) of younger cover sequences (Palaeoproterozoic until Tonian).

1.2.1.5. Other Cratons of East Antarctica

The southern sector of the *Shackleton Range* consists partly of a ~1.7 to ~1.8 Ga old basement (Will et al. 2009) unconformably overlain by Neoproterozoic strata. These are unmetamorphosed and internally undeformed beginning with the formation of an autochthonous palaeosoil (Buggisch et al. 1994). Thus, the cratonal area of the southern Shackleton Range shows both shield and platform. It may belong to the Mawson or even the Crohn Craton. However, the latter speculation is contradictory to aeromagnetic data in southern Dronning Maud Land (Mieth & Jokat 2014a).

The rather small cratonal area of the *Vestfold Hills* (mainly gneisses and migmatites) showing ages of ca. 2.5 Ga (Snape et al. 1997) also doesn't properly match the Crohn Craton.

Orthogneiss complexes of ~30 x 8 km^2 and ~2.5 x 0.3 km^2 showing ages of ~2.6 and ~3.0 to 2.9 Ga, respectively, occur on both sides of the mouth of the *Denman Glacier* in Queen Mary Land (Black et al. 1992). Both cratonal "islands" have been included rather speculatively into the Crohn Craton (Boger 2011). However, it seems to be nonsense to classify such small basement bodies as independent cratons.

The tiny isolated outcrops of the *Bertrab-*, *Littlewood- and Moltke nunataks* (BLM) in Prinzregent-Luitpold-Land (≈ Coats Land) may be parts of another ice-covered craton as discussed by Jacobs & Thomas (2004), Kleinschmidt & Boger (2009), Loewy et al. (2011) and Mieth & Jokat (2014a). The at least 1.1 Ga old undeformed rhyolitic rocks of Bertrab- and Littlewood nunataks (Gose et al. 1997) and the aeromagnetic pattern of the region (Mieth & Jokat 2014a) are the main arguments for it. However, the final interpretation depends on dating the inaccessible Moltke-Nunatakker (Kleinschmidt & Boger 2009). Incidentally, it's just possible, that a "BLM-Craton" (≈ Coats Land Block) extends to the northern edge of the Shackleton Range (in the Högbom Nappe, see chapter 5.2.7.) or in the basement of the Theron Mountains (see chapter 5.6.).

The cratonic areas of East Antarctica would be even more reduced and segmented, if Proterozoic orogens were or could be much better distinguished. This is underway with the ~1.7 Ga old Kimban Orogen (e.g. Boger 2011). Starting positions are parts of Terre Adélie and the Mertz Shear Zone of George V Land forming the boundary between the 2.4 Ga old basement of the Mawson Craton and 500 Ma old granites of the Ross Orogen (see chapter 8), where the Kimban Orogen is assumed to go along the rear of the Transantarctic Mountains and finally join the southern section of the Shackleton Range.

1.2.2. Grenvillian Orogens

The meaning of the term "Grenvillian Orogeny" has been generalised during the recent decades from a rather regional to a global event which is responsible for the formation of the supercontinent Rodinia. There are three such orogenetic belts in Antarctica (Fitzsimons 2000a): (1) the Maud Belt, (2) the Rayner Belt, and (3) the Wilkes Province.

The *Maud Belt* (Groenewald et al. 1995) forms the welded seam of the Grunehogna Craton on the one hand and of the Crohn Craton and parts of the Shackleton Range on the other hand. It runs from western to eastern Dronning Maud Land and comprises Heimefrontfjella, Kirwanveggen, H.U. Sverdrupfjella, Mühlig-Hoffmann-Gebirge, Wohlthatmassiv, Schirmacheroase, Sør Rondane, Belgica and Yamato Mountains. All these show additionally an obvious Pan-African overprint (see below), particularly the last six.

The *Rayner Belt* (northern Prince Charles Mountains and the Rayner Complex; Yoshida & Kizaki 1983) welds together the Napier Craton and the Crohn Craton in Enderby-, Kemp- and Mac.Robertson Lands.

The *Wilkes Province* (Fitzsimons 2000a) of Wilkes Land is exposed in the Bunger Hills and on the Windmill Islands. It welds together the Mawson Craton with conceivable parts of the Crohn Craton extending into Queen Mary Land or with the cratonic section of the Vestfold Hills (see above). The Wilkes Province continues as the Albany Fraser Belt into the Australian vis-à-vis and there it marks the soldered joint of the Mawson Craton with the Yilgarn Craton.

The Grenvillian provinces of Antarctica were erroneously considered in the 80s and early 90s as just one, very extended orogen following the Antarctic coastline as a 250 km wide strip from Coats Land in the west up to George V Land in the east. It was even given its own name: "Circum East Antarctic Mobile Belt" (Yoshida 1992). Such a circum-Antarctic orogen doesn't exist. The extension towards the east up to Terre Adélie and George V Land is incorrect, and towards the west up to Prinzregent-Luitpold-Land/Coats Land (BLM, see above) is unproven.

1.2.3. Pan-African Orogens

World-wide, Pan-African Orogenies took place around 650 to 500 Ma and largely resulted in the formation of the next supercontinent: Gondwana. Before this time – very roughly around 750 Ma – Rodinia disintegrated into several continental fragments, and between them oceans opened, e.g. the Mozambique Ocean between today's Africa and today's India. This process concerned also today's East Antarctica (West Antarctica didn't exist at that time). The southern extension of the Mozambique Ocean opened between the Grunehogna Craton (beeing part of the Kalahari Craton) and the rest of East Antarctica, as shown by oceanic relics – so-called ophiolites – in the northern Shackleton Range (chapter 5). Perhaps one arm of this ocean branched off in southern Enderby Land towards Prince Charles Mountains and Princess Elizabeth Land. Around the same time, the edge of the Paleo-Pacific Ocean formed in the area of today's Transantarctic Mountains. The subduction of this Paleo-Pacific Ocean led to the *Ross Orogen* 600 to 500 Ma ago. Thus the Ross Orogeny is of Pan-African age, but did not contribute to the formation of Gondwana, in contrast to the *other similar-aged orogens* of Antarctica; products of closures of the Mozambique Ocean and arms: *East Antarctic Orogen* (= *Lützow Holm Belt*) and *Kuunga Suture*.

1.2.3.1. The Ross Orogen

The Ross Orogen approximately follows the present Transantarctic Mountains and extends some 3400 km from northern Victoria Land, i.e. the Pacific end of the Transantarctic Mountains, up to its quasi-Atlantic end, i.e. the Pensacola Mountains. Westernmost Marie Byrd Land (Edward VII Peninsula) also belongs to the Ross Orogen; its separation occurred only much later near the end of the Cretaceous.

The Ross Orogen is characterised by folds (Fig. 1-4), thrusts, metamorphism, abundant granites, terranes, flysch- and molasse-like sediments. Thrust systems symmetrically west- and east-directed have been discovered in Oates Land, which are traceable into the Australian "Ross Orogen" known as the Delamerian Orogen (Flöttmann et al. 1993a). The systematic distribution of different types of metamorphism (HP v. LP; e.g. Talarico et al. 2004) and granites (S-type v. I-type; e.g. Vetter & Tessensohn 1987) led to the subduction model of the Palaeo-Pacific towards East Antarctica. In northern Victoria Land, the Ross Orogen consists of three terranes: the high-grade metamorphosed and granite-rich Wilson Terrane to the west, the very low-grade metamorphosed, turbiditic Robertson Bay Terrane to the east, and the likewise very low-grade metamorphosed, volcanic-rich Bowers Terrane inbetween. There is some discussion whether these terranes are allochthonous or represent an island arc (Bowers Terrane) or an accretionary wedge (Robertson Bay Terrane). All results argue for the Ross Orogen being a subduction orogen (so-called Andean type).

Fig. 1-4. Ross Orogen: Cambrian turbidites, folded around 500 Ma, northern Victoria Land, Transantarctic Mountains, near Robertson Bay (from Kleinschmidt 2001).

1.2.3.2. Other Pan-African Orogens

The *East Antarctic Orogen* (originally named "East Antarctic mobile belt" by Jacobs et al. 1998) comprises the chain Shackleton Range – Dronning Maud Land – Sør Rondane – Lützow-Holmbukta. Fitzsimons (2000b) proposed the name *Lützow Holm Belt* for the same orogen. The orogenetic structure northern Prince Charles Mountains – Grove Mountains – southeastern Prydz Bay or Denman Glacier has been described as [Antarctic] *Kuunga Suture* (Boger et al. 2002). The exact course of both the East Antarctic Orogen/Lützow Holm Belt and the [Antarctic] Kuunga Suture is uncertain because of the extensive ice-cover.

The *East Antarctic Orogen/Lützow Holm Belt* is characterised by thrust- and nappe tectonics (Buggisch & Kleinschmidt 2007, *cum lit*.), by abundant late-orogenetic collapse structures (Kleinschmidt & Brommer 1997), by thick molasse formations in the Shackleton Range and Kirwanveggen (Buggisch et al. 1999, Helferich et al. 2004, respectively) and by syn- and post-orogenetic magmatism in Dronning Maud Land (Roland 2004), i.e., by typical features of a collisional orogen. However, a direct link between the Shackleton Range and the rest of the orogen is inconsistent with recently discovered NW-SE trending aeromagnetic subglacial structures in southern Dronning Maud Land (Mieth & Jokat 2014a).

The base of the [Antarctic] *Kuunga Suture* comprises the northern Prince Charles and Grove Mountains. Its eastward continuation is rather speculative. It either runs into the Prydz Bay area as favoured by Boger & Miller (2004) or into the Pan-African portion of the Denman Glacier region (Black et al. 1992) as favoured by Boger in 2011.

Both orogenetic belts, the East Antarctic Orogen/Lützow Holm Belt and the Kuunga Suture, attest to the amalgamation of West Gondwana (South America, Africa, Grunehogna Craton) with East Gondwana (India, Australia, main East Antarctica). This could have happened in two steps according to Boger et al. (2001, 2002): (i) shortly before 550 Ma; (ii) shortly after 550 Ma. The first step joined West Gondwana and Indo-Antarctica (i.e. India, northern Prince Charles Mountains and Napier Complex) together, manifested as Mozambique Belt and its continuation into Antarctica, the main (= northern) part of the East Antarctic Orogen/Lützow Holm Belt. The second step added the rest of East Gondwana, i.e. Australia and the rest of East Antarctica forming the Kuunga Suture. This idea could explain the "double-track" of the East Antarctic Orogen/Lützow Holm Belt: In Dronning Maud Land it is mainly slightly older than 550 Ma and thus product of step (i). The Shackleton Range would belong to step (ii), i.e. mainly younger than 550 Ma. In this way, the existence of the old "alien element" between the two "tracks" (the BLM province south of Coats Land) would be plausible; it would have been added as a mini-craton during step (i) (Kleinschmidt & Boger 2009). Kleinhans et al. (2013) provided an alternative interpretation assigning the younger ages (e.g. from Sør Rondane) to the post-/late-orogenetic collapse. Ruppel et al. (2015) present a quite different plausible geodynamic concept: They interpret the area with the NE-SE trending aeromagetic anomalies by Mieth & Jokat (2014a) as indenter pushing apart the Pan-African belt section Sør Rondane/Lützow-Holmbukta to the east and the Pan-African belt section western Dronning Maud Land/Shackleton Range to the west (see chapters 5 and 6).

1.2.4. The lifetime of Gondwana

The Pan-African orogenies led to the formation of the supercontinent Gondwana. It existed from ca. 500 to 180 Ma ago. Its central part and heart was Antarctica. Testimonies of Gondwana's huge landmass are terrestrial sediments, the Antarctic parts of which are subsumed under the term Beacon Supergroup unconformably overlying the earlier formations of the cratons and of the Grenvillian and Pan-African orogens.

The Beacon Supergroup occurs in the Transantarctic Mountains from Victoria Land to the Pensacola Mountains, in George V Land, at the eastern edge of the Shackleton Range, north and south of the Shackleton Range (Theron Mountains and Whichaway Nunataks, respectively), in western Dronning Maud Land, a little bit modified und specially named in Mac.Roberston Land (Lambert Glacier area) and in the Ellsworth Mountains. Main rock types are fluvial sandstones confirming the predominantly terrestrial environment for the period Devonian–earliest Jurassic. The most spectacular strata are Permo-Carboniferous tillites and diamictites combined with glacial stria on the one hand and Permian coal seams on the other hand. This indicates a rather quick climatic change from glaciation to dense woodlands, i.e. from cold to humid conditions with rather mild temperatures (e.g. Collinson et al. 1994). The Beacon Supergroup is equivalent to the lower Gondwana Supergroup of India and South America, the Karoo Supergroup of southern Africa and Permo-Triassic analogues in Australia.

1.2.5. The disintegration of Gondwana

The desintegration of Gondwana started with the cracking of Gondwana's crust around 200 to 180 Ma ago. These cracks served in the early Jurassic (180 Ma) for extensive effusion of volcanic magma. This volcanism is referred to under the generic term Ferrar Supergroup. Its mainly subalkaline basic volcanics are widespread in western Dronning Maud Land, in the vicinity of the Shackleton Range (Theron Mountains and Whichawy Nunataks), in the entire Transantarctic Mountains from Pensacola Mountains up to Victoria Land and in George V Land. The volcanics form extensive flood basalt blankets

Fig. 1-5. Sill of Ferrar Dolerite: Horn Bluff, George Vth Land, height of exposure ca. 50 m.

("Kirkpatrick Basalts") and even more sills ("Ferrar Dolerites") which ± horizontally intruded earlier formations (basement, Beacon Sandstone; Fig. 1-5). They reach several hundred metres of thickness. Although their chemistry is not alkaline, they are commonly regarded as an indicator of crustal extension and thus for the starting desintegration of Gondwana (e.g. Elliot 1992). However, also a connection with subduction processes during the Ellsworth Orogeny (next paragraph) is taken into consideration. Totally analogous volcanics are widespread in southern Africa and in southeastern Australia (mainly Tasmania).

Locally, the desintegration of Gondwana is accompanied by plutonism, namely in the case of the separation of New Zealand from Antarctic Marie Byrd Land during the Cretaceous (ca. 100 Ma ago). Accordingly, anorogenetic A-type granites occur in both fragments of Gondwana.

Some of the initial cracks in the Gondwanan crust finally expanded into the present oceans. Gondwana's disintegration ended with the complete isolation of Antarctica, i. e. the opening of the deep-sea gateways between Victoria Land and Tasmania 33 Ma ago and between the Antarctic Peninsula and South America (Drake Passage) 32 Ma ago (Bell 2008). That this is one of the principal causes for the complete, extensive Antarctic glaciation 33.5 Ma ago, was one of the main outcomes of the International Polar Year 2007/08 (Miller et al. 2008).

1.2.6. The Ellsworth Orogen = Weddell Orogen

The Palaeo-Pacific edge of Antarctica and, thus, of entire Gondwana has existed since the assembly of Gondwana throughout Gondwana's lifetime. At this edge after the Ross Orogeny, another growth zone was added ca. 250 to 200 Ma ago, called Ellsworth Orogen, occasionally Weddell Orogen, and – in a broader sense – Gondwanides Orogen. The Gondwanides include extra-Antarctic continuations in southern Africa (Cape Fold Belt) and in South America (fold belt of the Sierra de la Ventana). The Ellsworth Orogen

Fig. 1-6. Ellsworth Orogen: folded Permian tillites („Whiteout Conglomerate") in the Ellsworth Mountains (phot. W. Buggisch; from Kleinschmidt 2014).

extends from the Ellsworth Mountains to the Pensacola Mountains and contains the Whitmore Mountains further south. The line Ellsworth Mountains–Pensacola Mountains represent the fold belt (Fig. 1-6), whereas the Whitmore Mountains are the magmatic arc of the orogen.

In the Pensacola Mountains, the structures of the Ellsworth Orogeny overprint the earlier structures of the Ross Orogeny. Both are parallel. However, the Ellsworth Orogen of the Ellsworth and Whitmore Mountains trends nearly at right angles to the Ross Orogen. According to palaeomagnetic data, this transverse position is due to a sinistral rotation of ca. 90° (Randall & Mac Niocaill 2004). However, the transition Ellsworth Mountains–Pensacola Mountains is covered by ice and therefore problematic. The same concerns the relationship with the Antarctic Peninsula. Both problem areas are most probably characterised by large fault zones, which is supported by the subglacial morphology (Bedmap2: Fretwell et al. 2013).

1.2.7. The Antarctic Andean Orogen

The Antarctic Andean Orogen occupies the entire Antarctic Peninsula roughly up to the Walgreen Coast. It formed mainly in three steps: (1.) Upper Jurassic to Lower Cretaceous (150–140 Ma); (2.) Mid Cretaceous (ca. 100 Ma); (3.) Palaeogene to Quaternary (ca. 50 Ma – present) (e.g. Birkenmajer 1994a, Vaughan & Storey 1997).

It is still under discussion whether locally discernible Permo-Triassic deformation and metamorphism have to be included into the Gondwanides or not. The Antarctic Andes are the youngest "growth ring" of the continent, and their orogenetic processes are partly still active (see below). Like the South American Andes, the Antarctic Andes represent a typical subduction orogen with masses of orogenetic magmatism in form of granitic to dioritic plutons and volcanoes and their remnants. In detail, the deformational and

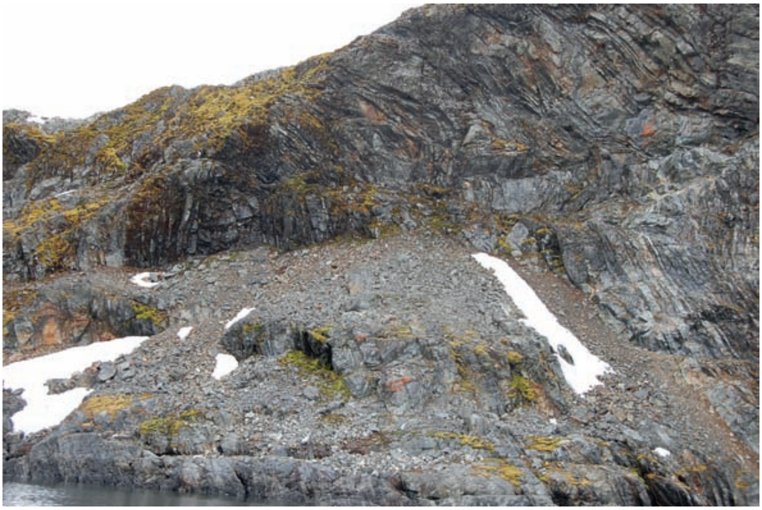


Fig. 1-7. Antarctic Andean Orogen: folding of Triassic (??) turbidites at Paradise Bay, Antarctic Peninsula.

metamorphic history is very complicated. In spite of the dominating magmatism, folding is abundant, e.g. in southernmost Palmer Land and at the northern end of Graham Land (Fig. 1-7). Palmer Land and Alexander Island are characterised by thrust tectonics with stacking of three main units ("Western, Central and Eastern Domains"; Vaughan & Storey 2000, Vaughan et al. 2002a). These findings are confirmed by the distribution of metamophic rocks (Wendt et al. 2008), including blueschists, typical of high pressure metamorphism in subduction complexes, e.g. on Elephant Island (Trouw et al. 1991, Trouw et al. 1998). However, the idea of a plain and simple subduction orogen is out-of-date. The current database suggests a much more complex interpretation with docking of terranes (see Smellie this vol., chapter 2).

1.2.7.1. The active part of the Antarctic Andean Orogen

The plate-tectonically youngest and the only more or less active part of Antarctica is situated northwest of the Antarctic Peninsula and is just the most Pacific-ward part of the Antarctic Andean Orogen. It comprises most of the South Shetland Islands and the Bransfield Strait.

Northwest of the South Shetland Islands, a small part of the Pacific Ocean floor has until recently undergone, and may still be a little bit undergoing, subduction under the Antarctic Plate at the South Shetland Trench. The rest of the almost entirely subducted, purely oceanic Phoenix Plate is occasionally called Drake Plate. The related melt-products form the dominantly andesitic-basaltic chain of volcanoes of the central South Shetland Islands (= island arc). Nearly all of these volcanoes are extinct and date from the Upper Createaceous up to the Eocene period (e.g. Nawrocki et al. 2010). Just one can be classified as active – at least geologically: Penguin Island, situated not quite 1 km south of King George Island, erupted last some 100 years ago. However, Penguin Island is mostly thought to be connected with the opening process of the Bransfield Strait.

(Some of the South Shetland Islands – Elephant Island with neighbours and Smith Island – belong like the Antarctic Peninsula to earlier stages of the Antarctic Andean Orogeny and consist of strongly folded and metamorphosed Mesozoic trench-filling).

The Bransfield Strait is situated south of the island arc and forms an embryonic oceanic basin with an active spreading zone running not in the centre but near the northern edge of the strait. It opens ca. 11 mm per year (Dietrich et al. 1998). It is accompanied by a mainly tholeiitic volcanism, partly submarine, partly as active island volcanoes (Deception and Bridgeman Islands). The Bransfield Strait has been regarded as an example of a back-arc basin some time ago. Meanwhile, this neat but simplified model has been given up (for details see Smellie this vol., chapter 2).

1.2.8. Large fault systems

The architecture of Antarctica, consisting of cratons and younger “growth rings”, i.e. the orogens mentioned, is truncated by relatively young, mostly extensional faulting tectonics. Particularly large or remarkable fault structures are:

- the West Antarctic Rift System (WARS) and its main part, the Ross Sea Graben (= Ross Sea Rift), situated in Antarctic’s Pacific sector,
- the Lambert Graben (= Lambert Rift), situated in East Antarctica,
- the graben Jutulstraumen-Penckmulde (in short: Jutul Penck Graben) situated in the Atlantic sector,
- the Rennick Graben as the main element of the strike-slip fault system of Victoria Land and Oates Land,
- the fault systems around the Shackleton Range,
- and a possible fault system in Kaiser-Wilhelm-II.-Land running ± coast-parallel.

The Ross Sea Graben/Rift is extremely wide (up to 900 km). Its subsidence started in the late Mesozoic (around 140 Ma), reached its main activity in the Palaeogene (around 40 Ma) and led to an enormous relief at its western shoulder. The displacement Transantarctic Mountains – Ross Sea amounts to more than 10 km. The crustal extension is accompanied by typical intracontinental, i.e. alkali, volcanism. It is still active in the form of the stratovolcanoes Mt. Erebus (3794 m) and Mt. Melbourne (2732 m) of Victoria Land, similarly at Mt. Berlin (3478 m), Mt. Sidley (4181 m), Mt. Siple (3110 m) and Mt. Takahe (3460 m) of Marie Byrd Land.

The Lambert Graben cuts across the Rayner Belt, the Kuunga Suture and the Crohn Craton and is at least partly filled with Permo-Triassic sedimentary rocks of the Beacon Supergroup. The fault tectonics started already in the early Palaeozoic, reached their peak in the Permian and lasted up to the Lower Cretaceous (Hofmann 1996). Possibly the subglacial Lake Vostok is located in a somewhat staggered continuation of a branch of the fracture system. The continuation of the Lambert Graben in India is called Mahanadi Rift, is situated in the state of Orissa southwest of Calcutta, and is filled with sediments of similar type and of the same age as in the Lambert Graben. Therefore the Gondwana sector of India–Antarctica can be very precisely reconstructed using the Lambert-Mahanadi rift system (Hofmann 1996).

The Jutul Penck Graben of western Dronning Maud Land is about as large as the Rhine Graben. It subsided probably around 140 Ma ago or somewhat later (Jacobs & Lisker 1999) and marks the southeastern boundary of the Grunehogna Craton using a much older structure. The continuation of the Jutul Penck Graben into the still active East African graben system is under discussion (Grantham & Hunter 1991). The idea is plausible, as sections of the East African graben system were active already in the Jurassic.

The extended strike-slip fault system of Victoria Land and Oates Land runs obliquely to and is cut off by the Ross Sea Graben. The fault system still shows a certain activity as verified by earthquakes in 1952, 1974 and 1998 (Reading 2002). The quakes in 1974 and 1998 are located in the main element of the system, the Rennick Graben. There, Ferrar volcanics and sedimentary rocks of the Beacon Supergroup are downthrown. These rocks are spectacularly folded, squeezed and thrust over the shoulders in some graben segments. This means a complicated development of the overall dextral strike-slip system. Additionally, compressive and extensional components alternate from place to place leading to the formation of transpressional and transtensional structures (the folds and thrusts or pull-apart basins, respectively; Rosetti et al. 2003a). The earthquake in 1952 happened ca. 120 west of the Rennick Graben in the course of the parallel structure of the Matusevich Glacier. It is under ongoing discussion, whether the strike-slip system of Victoria Land and Oates Land could represent continuations of oceanic fracture zones between Australia and Antarctica – e.g. of the Tasman Fracture Zone – into the Antarctic continental crust.

The mainly Pan-African Shackleton Range forms a horst uplifted at post-Jurassic fracture systems paralleling both the Slessor- and Revovery Glaciers. Accordingly, the Theron Mountains and the Whichaway nunataks – both made up of rocks of the Beacon and Ferrar Supergroups – are downfaulted to the north and to the south, respectively. This is clearly underlined by bedmap2 (Fretwell et al. 2013).

Additionally, bedmap2 suggests an 800 km long fault zone in Kaiser-Wilhelm-II.-Land running parallel to the coast up to 50 km inland. It is associated with the roughly 56.000 year old volcano Gaussberg (370 m; Tingey et al. 1983). It represents an intraplate volcano composed of alkali rocks (leucitites), similar to the volcanoes of the WARS.

1.2.9. The present plate-tectonic situation of Antarctica

The Antarctic Plate is the fourth largest of the Earth's present plates. It is larger than the North American and a little bit smaller than the Eurasian Plate. It comprises the continent and large portions of the Southern Ocean and is nearly completely surrounded by mid-oceanic spreading zones (Fig. 1-8). The continental part of the Antarctic Plate is nearly entirely bounded by passive continental margins, i.e., by oceanic portions of the Antarctic Plate. There is just one relatively short section of a more or less active continental margin concerning the Antarctic Peninsula between ca. 62° W and ca. 50° W with subrecent subduction. There, oceanic crust of the Pacific – actually of the former Phoenix Plate (= Drake Plate) – is or has been subducted underneath the Antarctic Plate (see paragraph 1.2.7.1.). Active subduction occurs as well close to the Antarctic plate boundary at the South Sandwich Trench. There, oceanic crust of the South American Plate is subducted westward under the oceanic crust of the Scotia Plate. This leads to the island arc volcanism of the South Sandwich Islands.

1.2.10. Appended paragraph on Antarctic meteorites

The occurrence of meteorites in the Antarctic is famous for both its frequency and its processes of concentration (Cassidy et al. 1992, Cassidy 2003, Schultz 2007, Roland 2009, Righter et al. 2014). 65–70% of all meteorites have been collected in the Antarctic (Quilty 2007)! Meteorites may contribute to decipher the geological evolution of certain Antarctic regions. Only in these cases, they are mentioned in the appropriate chapters.

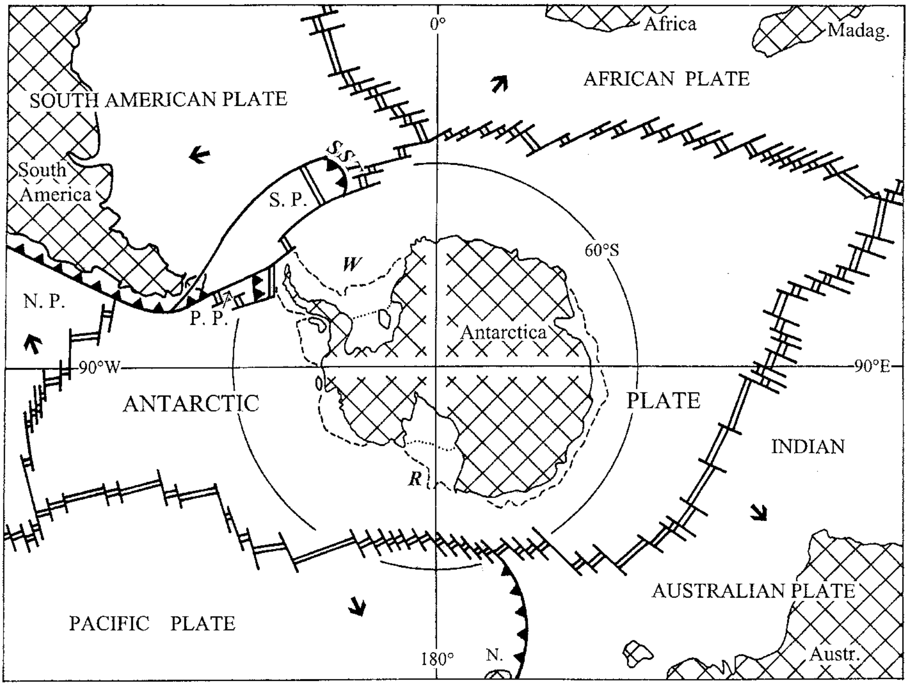


Fig. 1-8. The Antarctic Plate. Present plate-tectonic situation. Arrows indicate relative plate motions from the Antarctic point of view. Broken line: margin of continental shelf of Antarctica, dotted line: margin of largest ice shelves, double line: spreading zones, toothed line: subduction, single line: transform fault. N. = New Zealand, N.P. = Nazca Plate, P.P. = Phoenix Plate (≈ Drake Plate), R = Ross Sea, S.P. = Scotia Plate, SST = South Sandwich Trench, W = Weddell Sea. (Modified from Kleinschmidt 2001).

Chapter 1 completed: April 2015

2. Antarctic Peninsula – Geology and Dynamic Development

John L. Smellie

2.1. Introduction – Scope of this Chapter

More than a quarter century has elapsed since the geology of the Antarctic Peninsula region was last described in a dedicated volume (Barker et al. 1991). Allowing for some overlap, the overriding purpose of this chapter is to bring up to date the many recent advances in our understanding of the geology and dynamic evolution of the Antarctic Peninsula that have occurred since that volume was published. Accordingly, the major focus of this new comprehensive study is on the results of investigations that have been published during the past 25–30 years and the bibliography will reflect that. The previous review, by Barker et al. (1991), was focused principally on tectonic and geophysical aspects. By contrast, this chapter deliberately emphasizes the distribution, stratigraphy, geological characteristics and development of the many geological units, and they are described within an updated unifying tectonostratigraphical framework.

2.2. The Antarctic Peninsula: Physiography, Climate and Ice Cover

The Antarctic Peninsula is a well-defined physiographic province that extends over 16 degrees of latitude (c. 1700 km, from 61° to almost 77° S; Fig. 2-1). It encompasses Graham Land in the north and Palmer Land in the south. Previously, the southernmost part of the Peninsula was called eastern Ellsworth Land but recent changes in geographical naming have extended Palmer Land to encompass the former eastern Ellsworth Land. The Peninsula forms a long, northerly-narrowing, dissected landmass with an elevation that increases from >1000 m to >2000 m a.s.l. in a southerly direction. It contains more exposed rock than any other part of Antarctica but is extensively draped by the Antarctic Peninsula Ice Sheet, which merges with the West Antarctic Ice Sheet in the south. The west (Pacific) coast is characterised by numerous deep fjords, islands and small ice shelves whilst extensive ice shelves fringe much of the east (Weddell) coast, although several prominent ice shelf collapses have occurred recently on both flanks. Graham Land is dominated by an alpine glacier system that drains the narrow central plateau in steep, fast-flowing, mainly marine tidewater glaciers. By contrast, Palmer Land is draped by a more extensive ice sheet. The permanent ice cover is relatively thin today (typically 300–500 m; Lythe et al. 2001).

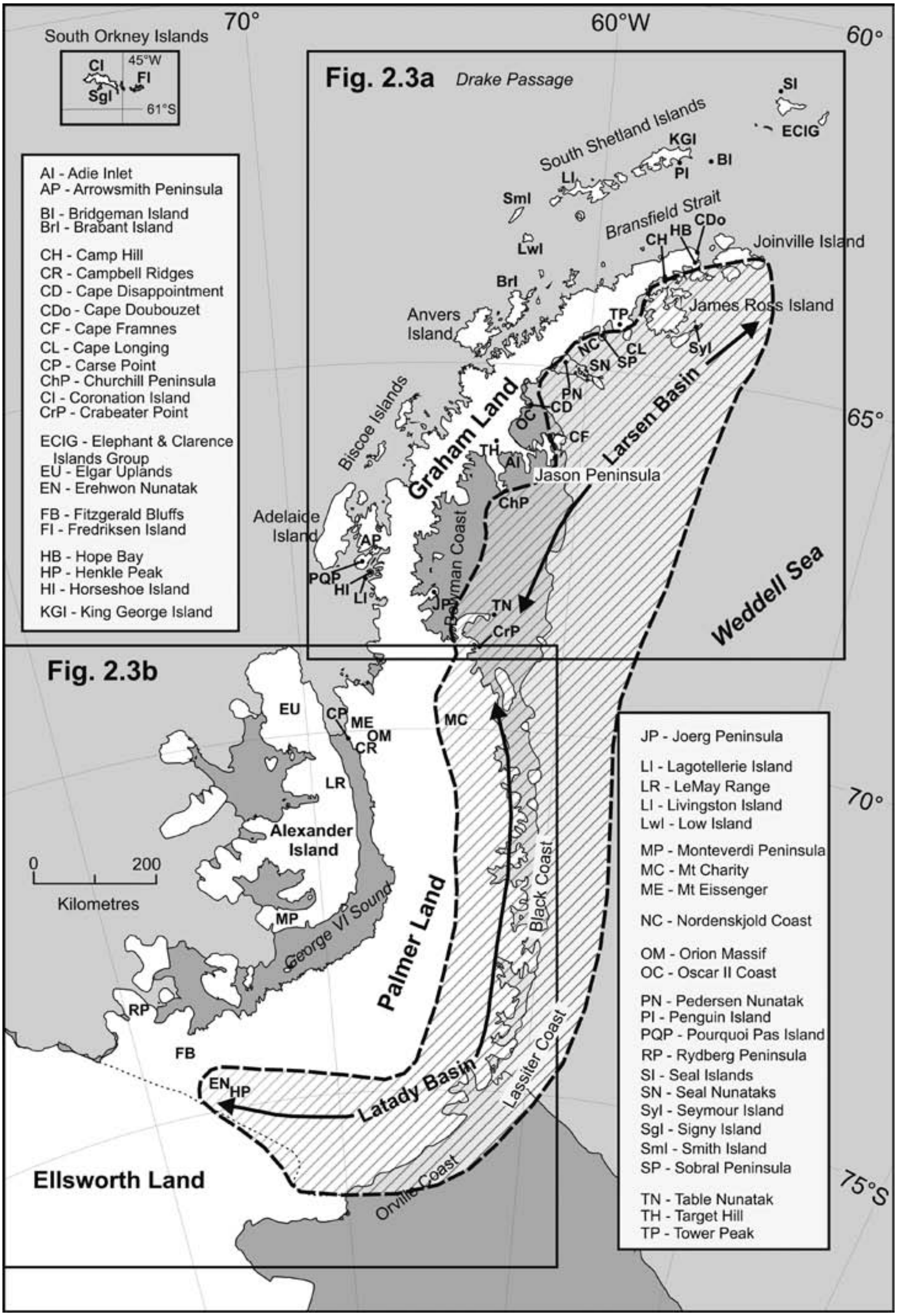


Fig. 2-1. Map of the Antarctic Peninsula showing the places mentioned in the text. The extent of the Larsen (Graham Land; mainly Cretaceous–Tertiary) and Latady (Palmer Land; Jurassic) sedimentary basins is also shown.

2.3. Tectonic Setting

The Antarctic Peninsula has been a continental margin throughout its known geological history. It is one of the key crustal blocks in West Antarctica and a major component of the proto-Pacific margin of Gondwana. It has traditionally been regarded as an ensialic Andean-type magmatic arc, active since early Mesozoic times (200 Ma), at least, formed as a result of eastward subduction of proto-Pacific and Pacific oceanic crust (e.g. Dalziel & Elliot 1973, Smellie & Clarkson 1975, Suárez 1976, Smellie 1981, Pankhurst 1982, Thomson et al. 1983, Storey & Garrett 1985, Harrison & Loske 1988, Milne & Millar 1989, Kellogg & Rowley 1991, Leat et al. 1995, Storey et al. 1996a). All the major tectono-stratigraphical elements of an arc–trench system have been recognised, including accretionary complexes, fore-arc and back-arc basins and the magmatic terrane (Fig. 2-2). A large number of stratigraphical units has been described (Figs. 2-3, 2-4). The arc system ultimately ceased activity following a series of ridge crest–trench collisions from Palaeogene time (Herron & Tucholke 1976, Barker 1982, Larter & Barker 1991). Subduction continues today only at the northernmost last-remaining segment, the South Shetland trench, albeit at a very slow rate (Robertson Maurice et al. 2003). A small marginal basin opened up in Bransfield Strait (e.g. Barker 1982) and a large alkaline basaltic stratovolcano and associated satellite volcanic centres formed in the James Ross Island region, in a back-arc position relative to continuing subduction at the South Shetland trench whilst, elsewhere in the Peninsula, post-subduction alkali basalts were erupted from several monogenetic volcanic fields (e.g. Hole et al. 1995).

These established views of the tectonic evolution of the Antarctic Peninsula as an autochthonous magmatic arc prevailed until the mid–late 1990's when interpretation of the tectonic history of the Antarctic Peninsula was significantly revitalised by two important new models, involving (1) terrane accretion and (2) pan-continental Gondwana break-up processes. In the terrane accretion model, the Peninsula is divided into three tectonic domains, of which the two westerly ones (considered 'exotic') are postulated to have docked against the third (the parautochthonous leading edge of Gondwana) during mid-Cretaceous times (Fig. 2-5; Vaughan & Storey 2000, Vaughan et al. 2002a, b). The Central Domain might be a composite feature composed of two magmatic arcs conjoined along a postulated >1500 km-long suture zone (Ferraccioli et al. 2006). The suturing event is thought to have caused a deformational tectonic episode known as the Palmer Land orogenic event, which deformed pre-mid Cretaceous volcanic and sedimentary rocks in Palmer Land and associated syn-tectonic plutons. Although the putative terranes are relatively well defined in Palmer Land, only one terrane boundary is exposed and the tectonic model becomes progressively harder to apply south through Palmer Land and, particularly, northward through Graham Land. On a much smaller scale, a less well-defined terrane model was also proposed independently for the South Shetland Islands, for volcanic sequences of Palaeogene age that might have been juxtaposed during the Neogene construction of King George Island (e.g. Birkenmajer 1997, Birkenmajer 2003). No mechanism for the postulated South Shetland Islands terrane accretion is yet known and it remains speculative. Moreover, the terrane accretion model for the Antarctic Peninsula has been re-evaluated, with a reversion essentially to the status quo ante (autochthonous continental arc) suggested (Burton-Johnson & Riley 2015). The distinction between the two models is not trival. It is integral to deciding between models of crustal development and whether Phanerozoic crustal growth took place at continental margins dominantly by in situ processes or by arc and terrane accretion.

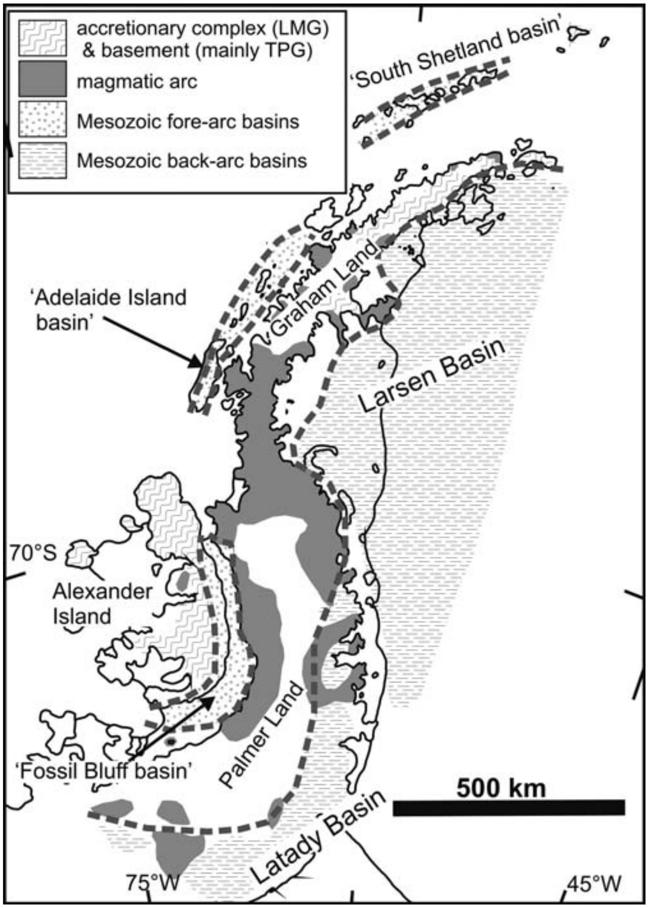


Fig. 2-2. Map showing the distribution of the principal tectono-stratigraphical elements comprising the Mesozoic Antarctic Peninsula arc–trench system (modified after Macdonald & Butterworth 1990). This configuration of the main tectono-stratigraphic elements remained essentially stable throughout the Cretaceous but broke down during the Cenozoic, as oceanic ridge segments approached and collided with the Peninsula trench, causing magmatism and accretion to move trenchward and cease. LMG – LeMay Group (Alexander Island); TPG – Trinity Peninsula Group (northern Graham Land).

In the Gondwana break-up model, it was postulated that regionally extensive outcrops of Jurassic volcanic sequences in the Peninsula, formerly regarded as 'normal' elements of a subduction-related magmatic arc (e.g. Weaver et al. 1982, Moyes & Smellie 1994), were generated during widespread crustal melting and voluminous eruption of rhyolites following the impingement of a very large mantle plume (e.g. Storey et al. 1992a). Together with a possible role for subduction-related plate-boundary forces, the plume might have been largely responsible for the break-up of the Gondwana continent. The plume has been linked to the early Jurassic eruption and emplacement of the voluminous Karoo and Ferrar large igneous (basaltic) provinces in South Africa and the Transantarctic Mountains, respectively (Storey et al. 1992a, Storey 1995, Storey & Kyle 1996).

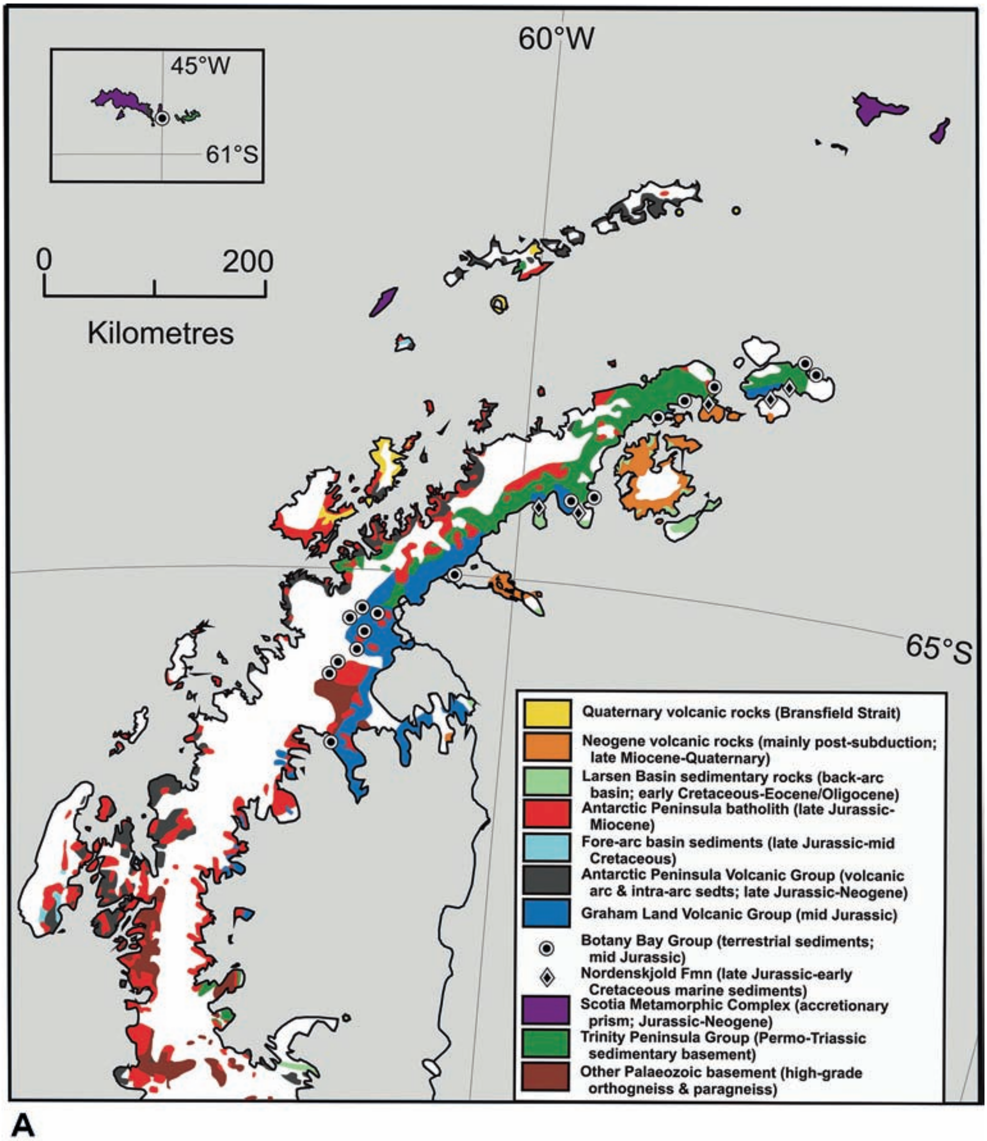
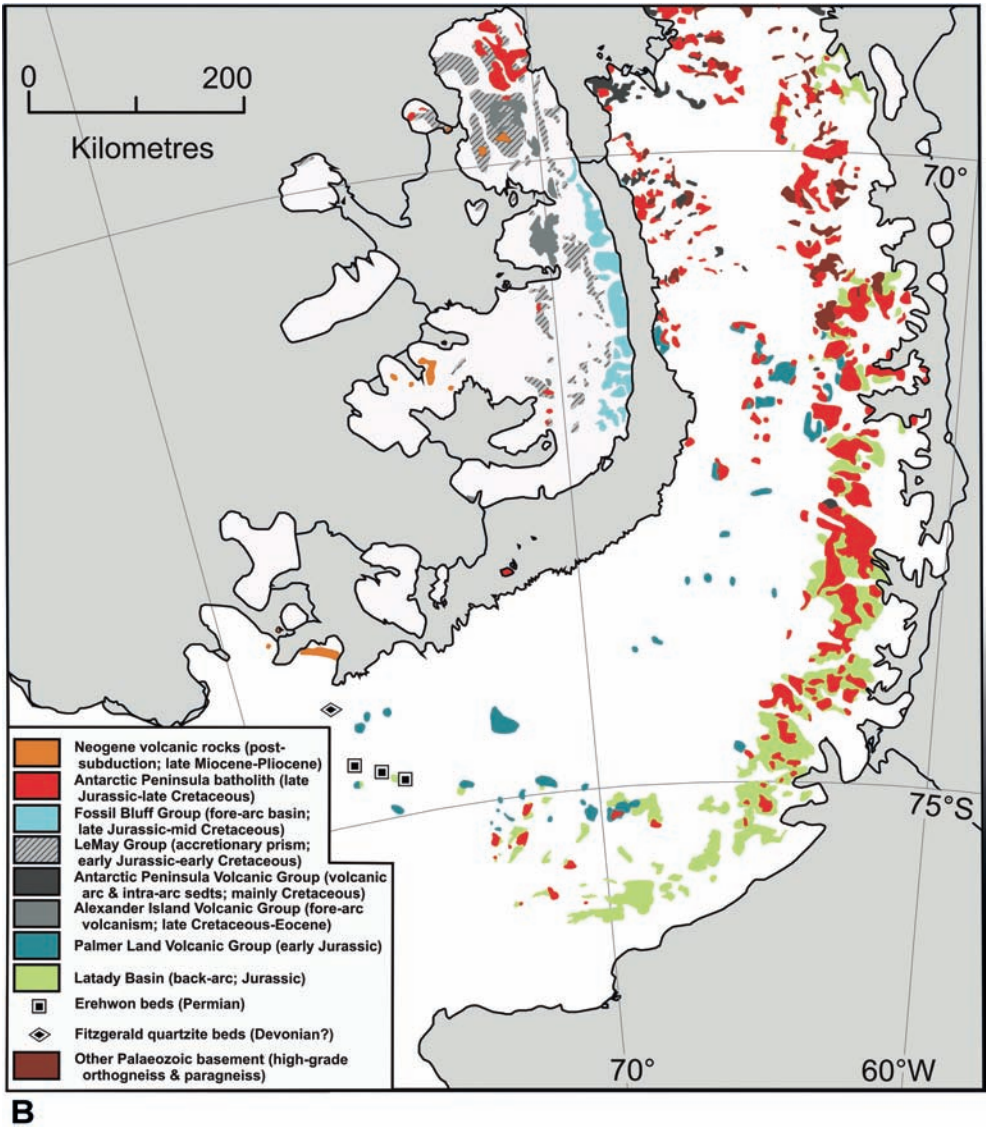


Fig. 2-3. Geological map of the Antarctic Peninsula. A. Graham Land (modified after Hathway 2000). B. Palmer Land and Alexander Island.

2.4. Pre-Middle Jurassic: Gondwana Margin "Basement"

Basement in the Antarctic Peninsula, defined here empirically as any pre-Middle Jurassic (i.e. pre- c. 180 Ma) outcrop, comprises three principal groups: (1) Gneissic outcrops, mainly granitic orthogneiss and migmatite but including less common layered paragneiss, calc-silicate and marble, found sparsely throughout the Antarctic Peninsula; (2) Three



groups of sedimentary strata, namely the Fitzgerald quartzite beds and the Erehwon beds in southern Palmer Land, and the Trinity Peninsula Group in Graham Land and its correlatives in the South Shetland and South Orkney islands; and (3) Schistose metasedimentary rocks grouped together as 'Terrane B', part of the Scotia Metamorphic Complex in the South Shetland and South Orkney islands.

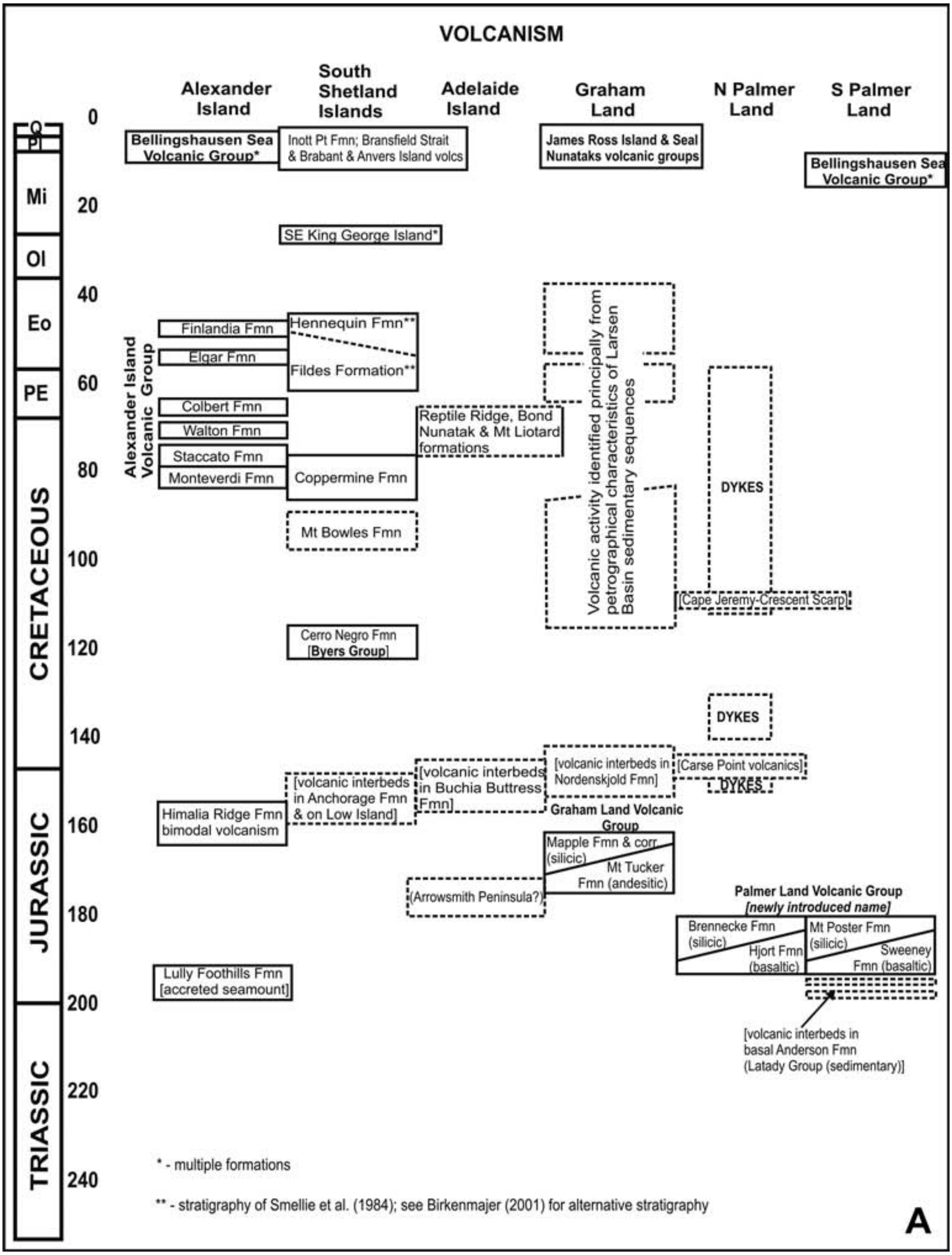
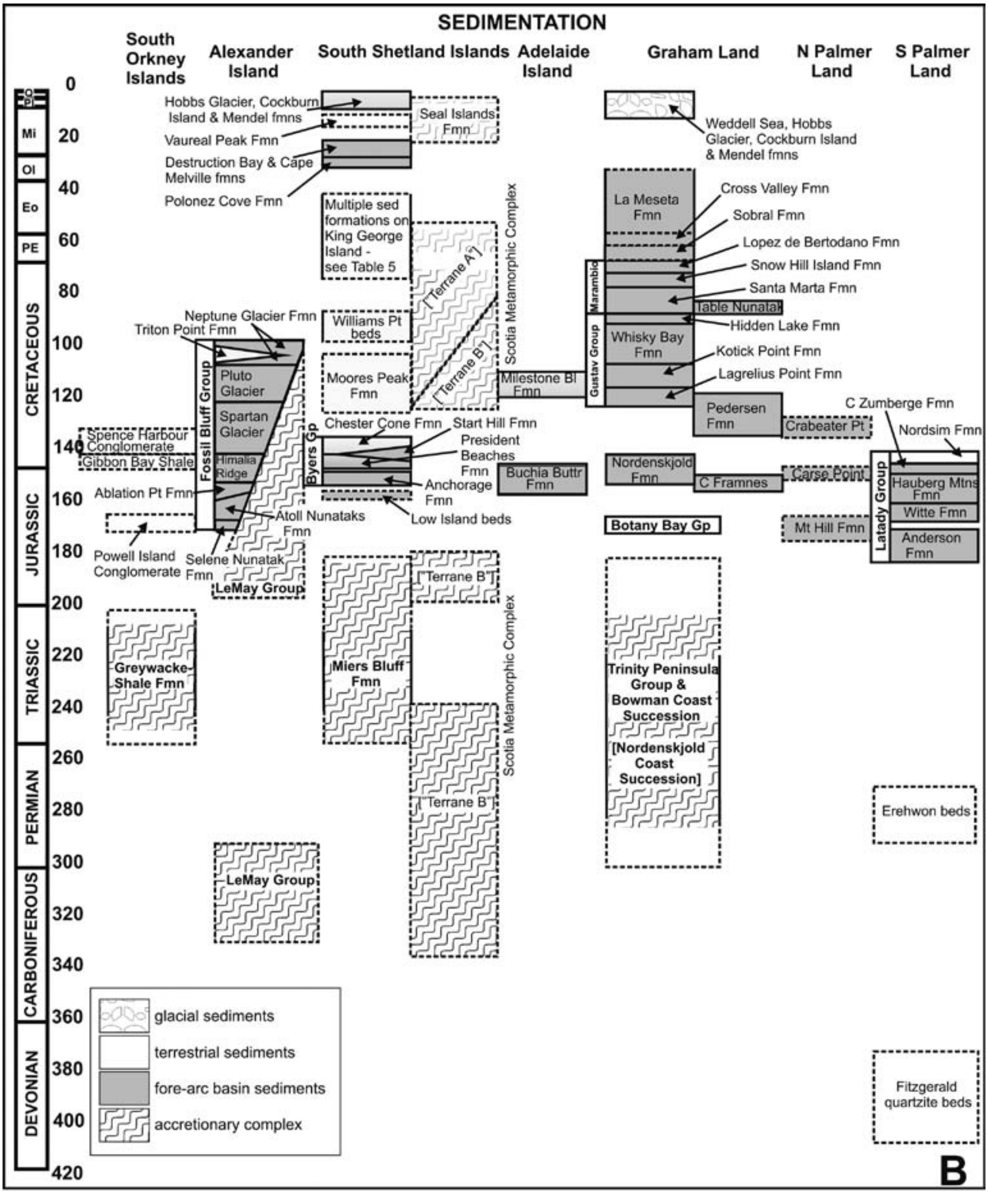
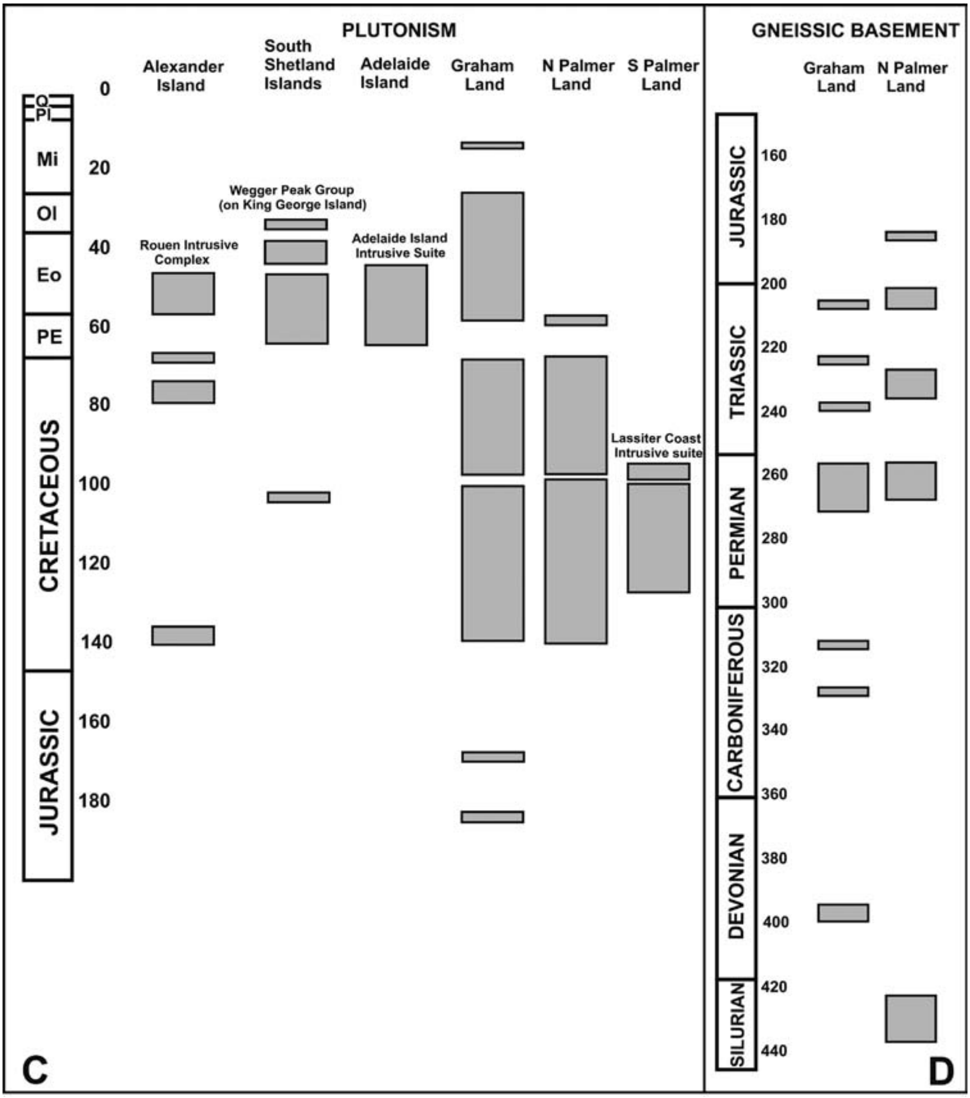


Fig. 2-4. Summary stratigraphy of the Antarctic Peninsula. A. Volcanism. B. Sedimentation. C. Plutonism. D. Gneissic basement. Note in B, age range for metasedimentary basement (zigzag ornament) includes polyphase deformation and metamorphism. Shaded boxes in B indicate marine sedimentary sequences.



2.4.1. Palaeozoic gneissic basement

Ortho- and paragneissic basement rocks are widely distributed but poorly exposed in the Antarctic Peninsula, largely hidden beneath ice and pervasively intruded by plutons of the Antarctic Peninsula 'Andean' batholith, which have often metamorphosed and deformed their country rock. Distinction between foliated Andean granitoids and orthogneisses is subjective and correlation between geographically widely separated outcrops is often impossible. Wide reliance has therefore had to be placed on high-precision isotopic geochronologies, principally by the Rb-Sr whole-rock method. More recently, U-Pb ages



and Hf isotopes obtained on zircons have significantly revised our understanding of the distribution of basement gneisses and many outcrops previously thought to be basement have turned out to be Cretaceous in age (i.e. part of the 'Andean' batholith) or Jurassic (associated with Gondwana break-up) and deformed due to emplacement-related strain effects rather than during orogeny (e.g. Pankhurst 1983, Millar et al. 2002, Flowerdew et al. 2006, Leat et al. 2009).

Table 2-1 and Fig. 2-6 summarise isotopic dating evidence for igneous basement in the Antarctic Peninsula. Despite model ages suggesting Precambrian events, there is no direct evidence in the Antarctic Peninsula for in situ basement older than Silurian (e.g. Loske et al. 1990, Millar et al. 2002, Flowerdew et al. 2005, Flowerdew et al. 2006). However, earlier events are represented by enclaves in plutons, clasts and detrital and inherited zircons, with ages ranging from 1870 to 460 Ma (Loske et al. 1990, Millar et al. 2002, Flowerdew et al. 2006). An abundance of zircons with ages of c. 520

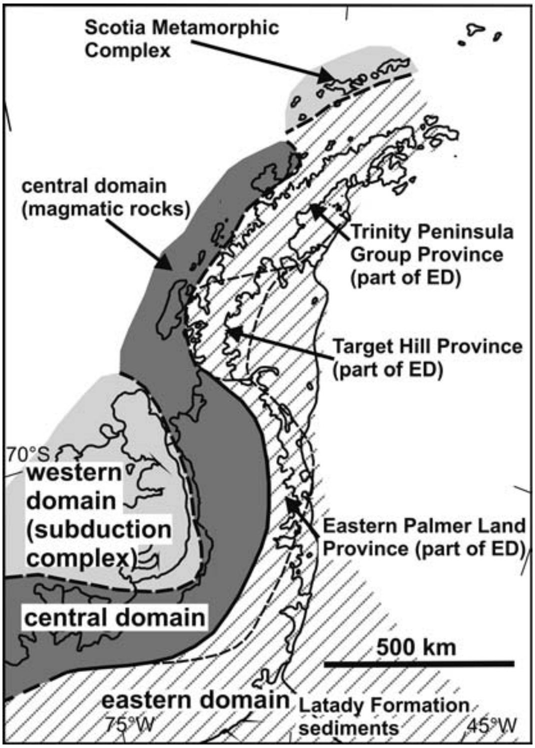


Fig. 2-5 Distribution of proposed tectonostratigraphical domains (terranes) in the Antarctic Peninsula (from Vaughan & Storey 2000). The Western and Central domains are thought to be exotic and docked by colliding against the parautochthonous Eastern Domain (representing the margin of the Gondwana continent). ED – eastern domain.

Ma in paragneiss from Adie Inlet (E Graham Land) suggests derivation from a provenance dominated by Cambrian granitoids, whilst orthogneiss at Cambell Ridges (west Palmer Land) contains inherited zircons with ages of 1000, 530 and 460 Ma. The oldest in situ basement is Silurian in age, comprising orthogneisses at Mt Eissenger (north-west Palmer Land; 435–422 Ma), which are similar in age to a conglomerate clast from Horseshoe Island (Tangeman et al. 1996). Orthogneisses at Target Hill (E Graham Land) have igneous crystallization and metamorphism ages of Devonian and Carboniferous (397 + 327; and 311 Ma, respectively). An important period of Permian (c. 270–258 Ma) and Triassic (c. 237–c. 227 Ma) granite emplacement has also been identified from migmatites, granites, tonalite and orthogneisses at Cape Dubouzet, Adie Inlet, Joerg Peninsula and Horseshoe Island (north, east and south-west Graham Land, respectively), and several localities in north-west and west Palmer Land. Geothermobarometry from migmatised paragnesises associated with the latter suggests conditions ranging between 800 °C at 5–6 kbar and 600 °C at 2 kbar (Piercy & Harrison 1991). The Permian–Triassic granites have high initial $^{87}Sr/^{86}Sr$ ratios and are 'S-types' (Wever et al. 1994, Millar et al. 2001). They represent a significant prolonged period of high crustal heat flow and anatexis. Palmer Land also has evidence for pre-Late Triassic and possibly lower Palaeozoic ages for protoliths of marble and paragneisses (at Auriga Nunataks; Vaughan et al. 1999). The origin and tectonic significance of the magmatism are far from clear. It might be related to either crustal extension or early subduction at the proto-Pacific margin (e.g. Leat et al. 1995, Millar et al. 2001, Millar et al. 2002, Flowerdew et al. 2006, Burton-Johnson & Riley 2015).

Table 2-1. Ages of pre-late Jurassic basement rocks in the Antarctic Peninsula (modified after Hervé et al. (1996), Millar et al. (2002), Flowerdew et al. (2006)).

Locality	Region	Lithology	Interpretation	Age (Ma)	2-σ error (Ma)
Mt Eissenger	NW Palmer Land	orthogneiss	Metamorphism	c. 202	
Auriga Nunataks	W Palmer Land	granodiorite	Igneous crystallization	205	?
Horseshoe Island	SW Graham Land	orthogneiss	Mylonitization	206	4
Orion Massif	NW Palmer Land	grey gneiss	Anatexis/metamorphism	c. 206	
Cape Doubouzet	N Graham Land	dacite porphyry	Anatexis	224	+24/-31
Pegasus Mountains	Central-W Palmer Land	orthogneiss	Igneous crystallization	228	6
Mt Eissenger	NW Palmer Land	orthogneiss	Anatexis/crystallization	228	3
Pegasus Mountains	Central-W Palmer Land	orthogneiss	Igneous crystallization	229	7
Pegasus Mountains	Central-W Palmer Land	orthogneiss	Igneous crystallization	230	8
Pegasus Mountains	Central-W Palmer Land	orthogneiss	Igneous crystallization	231	9
Pegasus Mountains	Central-W Palmer Land	orthogneiss	Igneous crystallization	232	10
Fomalhaut Nunatak	Central-W Palmer Land	granite	Igneous crystallization	c. 233	
Joerg Peninsula	SE Graham Land	orthogneiss	Igneous crystallization	238; 237; 236	5; 2; 7
Sirius Cliffs	NW Palmer Land	granite	Igneous crystallization	c. 233	
Adie Inlet	Central-E Graham Land	paragneiss	Migmatization	258	3
Orion Massif	NW Palmer Land	grey gneiss	Igneous crystallization	258	2
Mt Charity	N Palmer Land	porphyritic granite	Igneous crystallization	267; 259	3; 5
Horseshoe Island	SW Graham Land	orthogneiss	Igneous protolith	c. 270	
Target Hill	Central-E Graham Land	leucogranite	Metamorphism	311	8
Target Hill	Central-E Graham Land	leucogranite	Igneous crystallization	327	9
Target Hill	Central-E Graham Land	orthogneiss	Igneous crystallization	393; 397	1; 8
Mt Eissenger	NW Palmer Land	orthogneiss	Igneous protolith	422	18
Horseshoe Island	SW Graham Land	cobble	Igneous protolith	431	12
Mt Eissenger	NW Palmer Land	grey gneiss	Igneous protolith	435	8

2.4.2. Devonian(?) craton and passive margin sedimentation [Fitzgerald quartzite beds]

The Fitzgerald quartzite beds (informal name) comprise c. 300 m of quartzose metasandstones intruded by Cretaceous granite and hornfelsed to quartzite (Laudon et al. 1987, Laudon 1991). They are only known from Fitzgerald Bluffs, in southern Palmer Land (Fig. 2-3b), and their age is unknown but probably Devonian. Glacial erratics of similar-looking cross-bedded quartzite have been recorded on Rydberg Peninsula and as clasts in Mesozoic

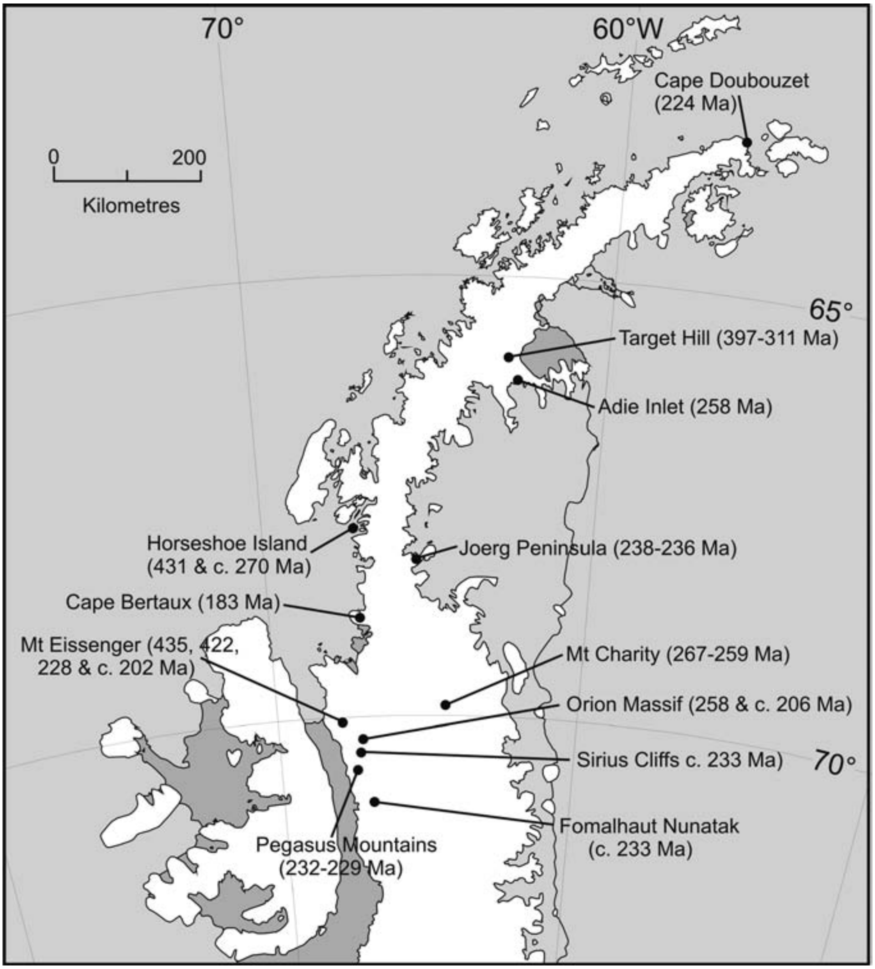


Fig. 2-6. Map showing the distribution and ages of dated pre-late Jurassic gneissic basement rocks in the Antarctic Peninsula.

conglomerates at several places in southern Palmer Land, suggesting that the Fitzgerald quartzite beds might have a large outcrop (Laudon 1991). Most of the sequence comprises quartz sandstone but there are also minor interbeds of siltstone and mudstone. The outcrop is largely inaccessible and the description is based principally on about 100 boulders fallen from a cliff. The sandstones are cross bedded and cross laminated, grey-green and vitreous, formed almost entirely (96%) of quartz grains and mostly lacking detrital feldspar. Based on detrital sand grain modal petrography, the sandstones have a continental interior (cratonic) provenance interpreted to reflect a passive margin tectonic environment that probably pre-dated any active-margin subduction processes in the region, an inference also suggested by whole-rock geochemical analyses (Laudon 1991, Laudon & Craddock 1992, Laudon & Ford 1997). The closest petrological (and therefore likely age) similarities are with the Devonian Crashsite Quartzite of the Ellsworth Mountains and the Beacon Supergroup in the Transantarctic Mountains (Devonian Taylor Group).

2.4.3. Permian sedimentation and transition to a subducting plate margin [Erehwon beds]

The Erehwon beds (informal name) are restricted to a small cluster of tiny isolated nunataks in southern Palmer Land, at Erehwon Nunatak (just 6 m high), 'Sobaco Nunatak' and small outcrops on the NE side of Henkle Peak (Fig. 2-3b; Laudon et al. 1987, Laudon 1991). Thicknesses of exposed strata range from 2 m at Erehwon Nunatak to 28 m at 'Sobaco Nunatak'. They are composed of interbedded dark grey thinly laminated mudstone, siltstone and fine-grained volcanogenic sandstone. Some of the siltstones resemble devitrified tuffs. One of the muddy beds, 10 cm thick, contains abundant intact *Glossopteris*, *Phyllotheca* and *Equisetum* leaves that resemble a Late(?) Permian Gondwana flora probably deposited in a lake or quiet-water coastal setting (Gee 1989a). Detrital sand-grain modes are characterised by abundant modal quartz, feldspar and lithic fragments. Many lithic grains are unrecognisable due to alteration and recrystallization but volcanic grains are the most abundant type, and there are rare micaceous metamorphic grains. A dissected and/or transitional magmatic arc provenance is suggested (i.e. consuming plate margin), in contrast to the pre-subduction setting inferred for the Fitzgerald quartzite beds (Laudon 1991, Laudon & Craddock 1992, Laudon & Ford 1997). Geochemical analyses suggest passive margin to active continental margin settings (Laudon & Ford 1997). The closest petrographical similarities are with the Permian Polarstar Formation in the Ellsworth Mountains. The Erehwon beds might have been deposited on the proto-Pacific side of the same sedimentary basin. They might therefore represent products of an early phase of arc magmatism along the margin of Gondwana (Rowley et al. 1991a). Deposition of the Erehwon beds was essentially coincident with Permian 'S-type' granitoid emplacement in Graham Land and Palmer Land (see above).

2.4.4. Permian–Triassic deep-sea fan sedimentation at a passive margin evolving to subducting margin [Trinity Peninsula Group and correlatives]

The Trinity Peninsula Group consists of polyphase deformed low-grade (mainly prehnite-pumpellyite facies) metasedimentary sequences (Dalziel 1984, Smellie 1991a, Smellie & Millar 1995, Smellie et al. 1996, Bradshaw et al. 2012). It is the most widespread and best exposed tectono-stratigraphical basement unit in the Antarctic Peninsula and crops out extensively in northern Graham Land (Fig. 2-3a). The presence of distal outliers in the Bowman Coast, south-east Graham Land (Bowman Coast Succession (BCS): Flowerdew 2008), suggests that it formerly had a much greater outcrop extent. The BCS differs from other Trinity Peninsula Group outcrops in having waterlain tuffs and intermediate and silicic lavas in its uppermost part. The Trinity Peninsula Group in Graham Land is at least 3 km thick (possibly as much as c. 5–10 km) and is divided stratigraphically into multiple formations (Hope Bay, Legoupil, View Point, Bahia Charlotte and Paradise Harbour formations; Alarçón et al. 1976, Hyden & Tanner 1981, Birkenmajer 1992a, see also del Valle et al. 2007). Distinction is based on a combination of sedimentological characteristics and geographical separation but the extent, age and stratigraphical relationships between the named formations are unknown due to a lack of distinctive marker beds and fossils. Sandstone detrital modes and whole-rock geochemistry have also been use to subdivide the sedimentologically monotonous

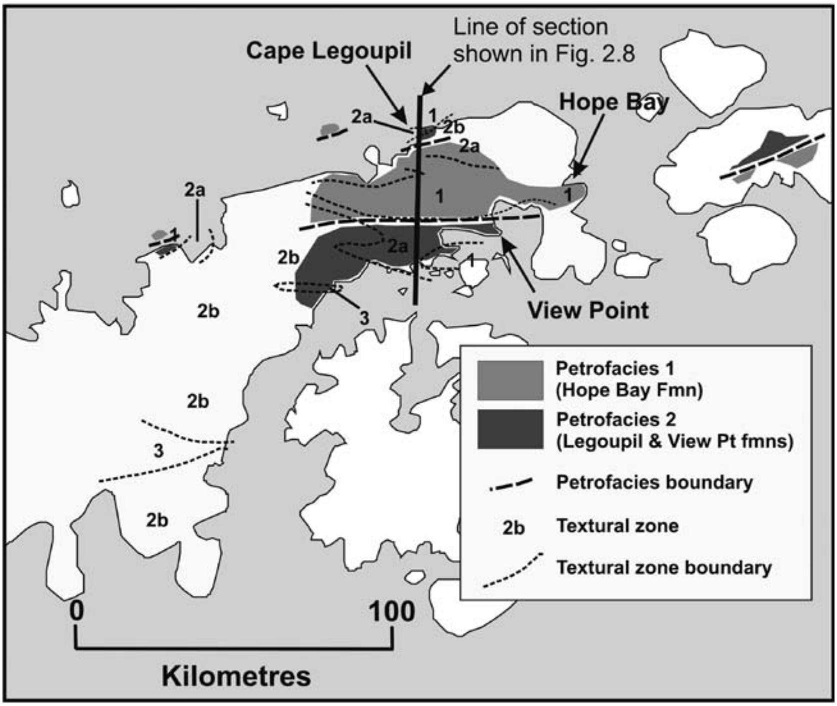
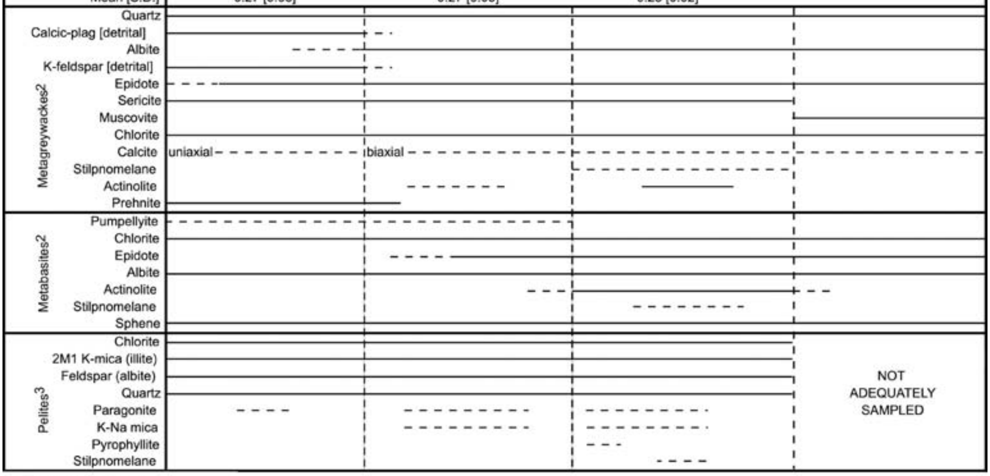


Fig. 2-7. Map of northern Graham Land showing the distribution of three major sand detrital mode petrofacies (coloured in greys) and textural zones (numbered). After Smellie (1991a) and Smellie et al. (1996).

succession in Graham Land and might prove to be the most practical method for establishing a regional stratigraphy (Smellie 1987a, Smellie 1991a; see also Arche et al. 1992a, Andreis et al. 1997, Willan 2003, Castillo et al. 2015). At least three distinctive petrofacies are distinguished. One petrofacies is composed of thick beds of feldspathic arkosic sandstones and corresponds stratigraphically to the Hope Bay Formation, whilst the second encompasses both the Legoupil and View Point formations, composed of thinner quartzose sandstone beds and a greater proportion of mudstones. The third petrofacies is less well defined but shows intermediate characteristics and might correspond to the Bahia Charlotte Formation, in part at least. The distribution of the three petrofacies also broadly coincides with textural zones based on metamorphic and microstructural characteristics (Smellie 1991a; Fig. 2-7; Table 2-2). A conceptual stratigraphical order was suggested by Smellie (1991a) corresponding to: the Hope Bay Formation younger than the Bahia Charlotte Formation younger than the combined Cape Legoupil and View Point formations. Latest Carboniferous (302 Ma) U-Pb ages obtained on zircons in the View Point Formation suggest that it might be older than the Legoupil Formation, which may be Early or Middle Triassic (<250 Ma; Thomson 1975, Bradshaw et al. 2012). Smellie (1991a) suggested that regional-scale strike-slip faults might separate the Hope Bay Formation from the View Point and Legoupil formations (Fig. 2-8). In addition, an un-named stratigraphical unit metamorphosed to greenschist facies crops out in the Nordenskjöld Coast. It yielded K-Ar ages for metamorphism extending back to the latest Permian indicating a Permian or older (>249 Ma, probably <300 Ma) depositional age. It thus might be one of the oldest and deepest structural parts of the Trinity Peninsula

	1	2a	2b	3
Textural zones in greywacke sandstones (after Smellie, 1991)	Essentially diagenetic textures; grain boundaries distinct and unmodified	Slightly recrystallized; grains show marginal recrystallization; clastic textures but some samples show faint tectonic foliation	Moderate-extensive recrystallization; relict clastic textures modified by weak-strong tectonic foliation; mortar texture in most modified sandstones	Completely recrystallized strongly foliated schists
Textural zones in mudrocks (Aitkenhead, 1975; and this paper)	Detrital textures; weak foliation[1] only in clay-rich laminations; recrystallization only of finest detrital grains; clay crystals mainly < 2 microns, up to 7 microns	Relict detrital textures (sand grains, bedding); weak to moderate bed-parallel foliation and spaced crenulation cleavage, mainly in fold hinges; clay crystal sizes as zone 1[1]	Extensive recrystallization; detrital textures poorly preserved; even largest clasts often shape modified and recrystallized; strong mica foliation (not alway bed-parallel) and crenulation cleavage; micas < 5 microns, up to 10 microns	Completely recrystallized, strongly foliated phyllites and schists; no detrital textures; mica crystals commonly up to 50 microns (Smellie & Millar, 1995)
Metamorphic grade	prehnite-pumpellyite facies; anchizone		pumpellyite-actinolite facies; epizone	greenschist facies
Kubler Indices: Range / Mean [S.D.]	0.21-0.41 / 0.27 [0.06]	0.24-0.32 / 0.27 [0.03]	0.18-0.25 / 0.23 [0.02]	(0.23, 0.24)



1. May be detrital fabric
2. Optically determined (Smellie 1991)
3. Determined by XRD (this paper)

Table 2-2. Summary metamorphic and textural characteristics of the Trinity Peninsula Group in northern Graham Land (from Smellie et al. 1996).

Group or else it is an unrelated older formation (Smellie & Millar 1995). Other correlatives and constituent formations of the Trinity Peninsula Group comprise the Miers Bluff Formation and Greywacke-Shale Formation, in the South Shetland and South Orkney islands, respectively (Dalziel 1984, Arche et al. 1992b, Doktor et al. 1994, Smellie et al. 1995). In its sandstone modes, the Greywacke-Shale Formation closely resembles the Legoupil/View Point combined formation, whereas the Miers Bluff Formation resembles the third petrofacies unit/Bahia Charlotte Formation (Smellie 1991a). The sequences are dominantly formed of mudstones and greywacke sandstones of turbiditic origin. There are also rare beds of limestone and volcanic rocks, including deformed dykes (Fleet 1968, Hyden & Tanner 1981, Smellie 1991a, Flowerdew 2008). Deposition of the entire Trinity Peninsula Group is typically assigned to submarine fans in shelf and/or slope environments (e.g. Hyden & Tanner 1981, Dalziel 1984, Bradshaw et al. 2012) but Birkenmajer (1992b) suggested shallow-marine deltaic deposition for his Scar Hills Member of the Hope Bay Formation. The moderate to steeply dipping strata are deformed by at least two main deformation phases, including early large-scale isoclines (nappes; Dalziel 1984). The fold axes strike dominantly to the NE and consistently verge to the SE (Fig. 2-8).

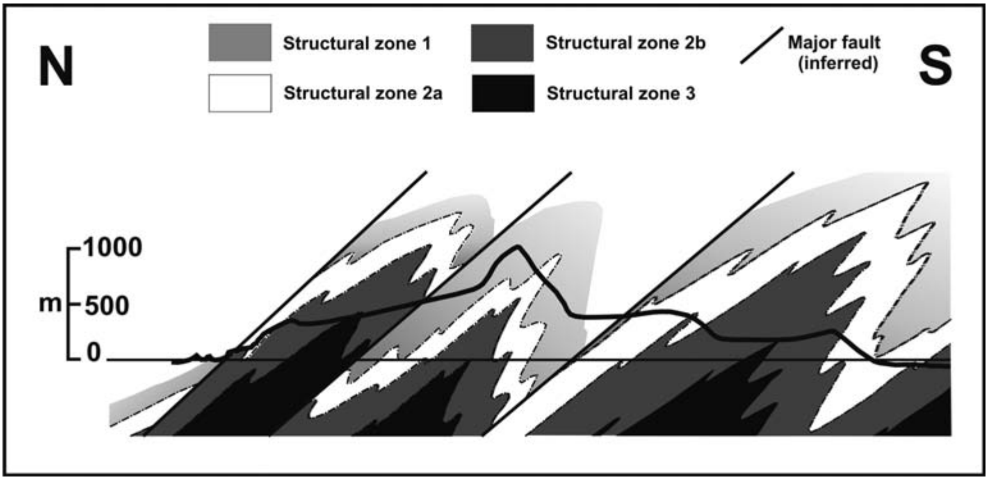


Fig. 2-8. Conceptual north–south cross section of northern Graham Land depicting large-scale deformation of the Trinity Peninsula Group, based on the distribution of textural zones and bedding attitudes. Line of section shown in Fig. 2-7. About 12 times vertical exaggeration.

The age of the Trinity Peninsula Group is not well established. Cobbles and detrital zircons in the View Point, Legoupil and Miers Bluff formations indicate Archaean to Permo-Carboniferous provenance ages that reflect the crystallization of source region granitoids (Loske et al. 1988, Hervé et al. 1991, Millar et al. 2002; see also Barbeau et al. 2010, Bradshaw et al. 2012). The abundance of Ordovician granite clasts (c. 466–487 Ma) in the View Point Formation suggests that they might have a relatively local provenance (Millar et al. 2002, Bradshaw et al. 2012). A minimum age is provided by the presence of undeformed sediments of the overlying Botany Bay Group (169 Ma; Hunter et al. 2005). The Bowman Coast Succession is probably Triassic (c. 238 Ma? Flowerdew et al. 2006). Rb/Sr dating of Miers Bluff Formation sediments also yielded a wide range of mainly Triassic ages (281–197 Ma; Dalziel 1972, Pankhurst 1983, Hervé et al. 1991, Willan et al. 1994, Trouw et al. 1997) commonly interpreted as depositional or diagenetic ages although they may relate to deformation or incipient metamorphism. They may even be a composite age from combining Permian with younger Mesozoic detritus – the presence of scarce early Middle Jurassic zircons in the Miers Bluff Formation and Trinity Peninsula Group suggests that at least parts of the sequence might be younger, perhaps as young as Middle Jurassic (Bajocian) for the Miers Bluff Formation (Hervé et al. 2006, Barbeau et al. 2010). Metamorphism of the Nordenskjöld Coast greenschists at c. 249±7 Ma (Permo-Triassic boundary), if truly part of the Trinity Peninsula Group, would be the first unequivocal evidence for pre-Triassic deposition and for a poly-metamorphic as well as polydeformational history of the group (Smellie & Millar 1995). Fossil remains are rare and comprise Early or Middle Triassic fossils in the Legoupil Formation (Thomson 1975), Late Triassic radiolaria in the Greywacke-Shale Formation (Dalziel et al. 1981), and stratigraphically undiagnostic carbonized plant fragments, bivalves, jellyfish impressions, trace fossils and algal mats (Schopf 1973, Dalziel et al. 1981, Birkenmajer 1992b, del Valle et al. 2007). The early reported presence of Carboniferous spores in the Hope Bay Formation has been discredited (Askin & Elliot 1982). Shu et al. (2000) described a rich Late Triassic palynoflora in the Miers Bluff Formation, whilst Deng et al. (2002) reported the presence of a Late Triassic (possibly

Norian–Rhaetian) palynoflora in the basal Miers Bluff Formation. By contrast, Pimpirev et al. (2000, 2006) have described a Tithonian ammonite in a loose block believed to be derived from the Miers Bluff Formation, and Campanian–late Maastrichtian or possibly Palaeocene calcareous nannofossils from throughout the Miers Bluff Formation. The Late Cretaceous ages were apparently confirmed by Stoykova et al. (2002). However, whilst the Tithonian age for the ammonite is within the possible depositional age range indicated by the U-Pb isotopic ages, the presence of possible Late Cretaceous nannofossils is enigmatic (cf. Shu et al. 2000, Hervé et al. 2006, Barbeau et al. 2010). Moreover, the conspicuous polyphase large- and small-scale deformation present in the Trinity Peninsula Group and its correlatives contrasts with that of unmetamorphosed and slightly to undeformed Early Cretaceous volcanic and volcaniclastic strata on Livingston Island (e.g. Smellie et al. 1984, Hervé et al. 2006), and with undeformed Middle Jurassic (c. 170 Ma) strata of the Botany Bay Group, which unconformably overlies the Trinity Peninsula Group in Graham Land (see later). Finally, a minimum age for the Miers Bluff Formation is indicated by the age of an intruding undeformed pluton dated as 137 Ma (earliest Valanginian; Hervé et al. 2006).

The tectonic setting of the Trinity Peninsula Group is contentious. The detrital modes suggest an early (Carboniferous–Lower or Middle Triassic) passive margin setting that gave way to an eroded continental margin magmatic arc (Upper Triassic–Middle Jurassic). This is similar to the possible tectonic relationships between the Fitzgerald quartzite beds and Erehwon beds, although the relative timing for the different parts of the Trinity Peninsula Group is uncertain and the precise structural position of the sedimentary basin relative to a possible coeval arc remains uncertain (Storey & Garrett 1985, Smellie 1987a, Smellie 1991a, Trouw et al. 1997). Most authors agree that it was probably in a fore-arc position, either as an accretionary prism or upper-slope basin (Hyden & Tanner 1981, Smellie 1981, Smellie 1987a, Smellie 1991a, Tanner et al. 1982, Dalziel 1984, Storey & Garrett 1985, Andreis et al. 1997, Millar et al. 2002, Bradshaw et al. 2012). The location of the contemporaneous arc is unknown, although assumed to be to the south-east (present coordinates). Evidence for coeval volcanic activity is scarce but includes rare tephra layers, ignimbrite, peperite, hyaloclastite, pillow lava and autochthonous deformed dykes in Graham Land (Smellie 1991a, Willan 2003, del Valle et al. 2007), and intermediate and silicic lavas in the Bowman Coast Succession (Flowerdew et al. 2006). With recent dating indicating deposition of the Miers Bluff Formation at least as young as Middle Jurassic (Bajocian), the suggestion has arisen that the Miers Bluff Formation may be the youngest part of the Trinity Peninsula Group, and it was deposited to the north-west (present coordinates) of the Trinity Peninsula Group depocentre and the coeval continental margin (Hervé et al. 2006). Also, major large-scale structures appear to verge to the south-east (i.e. arcward in reconstructions), the dykes and some volcanics have *alkaline* compositions, and the metamorphic assemblages suggest a medium pressure gradient, which are characteristics more akin to an arc-rear or foreland position (Smellie et al. 1996). Moreover, the distribution of basement gneiss outcrops (Fig. 2-6) suggests the presence of extensive Palaeozoic continental crust underlying the Trinity Peninsula Group in Graham Land. The most recent study, by Castillo et al. (2015), concluded that the Trinity Peninsula Group was associated with an active continental margin, following a passive margin setting for the oldest units (View Point Formation), an interpretation similar to most other authors (e.g. Smellie 1981, Smellie 1987a, Smellie 1991a, Dalziel 1984, Storey & Garrett 1985). By contrast, Hervé et al. (2006) suggested that deposition of the Miers Bluff Formation (and therefore presumably also, by implication, the older Trinity Peninsula Group) may have pre-dated the unequivocal initiation of easterly-directed subduction that was associated with the continental arc.

2.4.5. Pre-Jurassic deposition and accretion [Scotia Metamorphic Complex (Terrane B)]

The Scotia Metamorphic Complex crops out widely in the South Orkney and South Shetland islands. Two contrasting terrains are present, known as Terrane A and Terrane B, which are in tectonic or metamorphic contact on Elephant Island (Tanner et al. 1982, Dalziel 1984). Trouw et al. (1991) mapped gradational transitions in metamorphic grade that cut obliquely across the previously described tectonic/metamorphic boundary on Elephant Island and cast doubt on its significance. Terrane B, which is the older of the two terrains, grades structurally and in metamorphic grade up into highly deformed but relatively unmetamorphosed sediments of the Greywacke-Shale Formation (Trinity Peninsula Group correlative) on Fredriksen and Powell islands (South Orkney Islands), a contact known as the Powell Island transition zone (Dalziel 1984). Only Terrane B is considered here. It crops out on Coronation, Powell and Signy islands (South Orkney Islands) and on southern Elephant, Gibbs, Aspland, Eadie and O'Brien islands (Elephant and Clarence Islands Group, South Shetland Islands). It comprises albite-epidote-amphibolite and amphibolite facies assemblages that were metamorphosed under pressures approaching 7 kb and temperatures up to 500 °C (Tanner et al. 1982). Less common lithologies include marble on Signy Island and ultramafic rock on Gibbs Island (de Wit et al. 1977, Storey & Meneilly 1985). Each major outcrop area is characterised by a complex polyphase structural history and penetrative fabrics, involving up to three generations of sub-isoclinal to isoclinal folds and related planar and linear tectonite fabrics that have generally been ascribed to formation in a subduction complex (e.g. Smellie 1981, Tanner et al. 1982, Dalziel 1984, Storey & Garrett 1985).

Isotopic ages for Terrane B in the South Orkney Islands range from 180 to 281±56 Ma (Tanner et al. 1982, Hervé et al. 1984, Hervé et al. 1991). Tanner et al. (1982) interpreted a well-defined event at c. 188±5 Ma as the main period of mineral growth following an early intense deformation (i.e. cooling soon after the main prograde metamorphism) rather than subsequent thermal resetting of older rocks. By contrast, Terrane B on Elephant Island yielded Rb-Sr ages of 102–96 Ma and K-Ar ages of 118–80 Ma, with a much younger reset age (29 Ma), suggesting Cretaceous metamorphism and subduction. However, a Rb-Sr age of 287±48 Ma (Permo-Carboniferous) for Gibbs Island schists suggests a Late Palaeozoic metamorphism (Hervé et al. 1991). Moreover, the high initial $^{87}Sr/^{86}Sr$ ratios suggest a more extended crustal history for the metamorphic rocks from both regions, as sediments probably deposited no earlier than the Permian, although it could also be due to incorporation of sedimentary detritus derived from an old crustal provenance. Overprinting of pre-mid Jurassic events in Terrane B during accretion associated with the Peninsula 'Andean' arc (see later) probably yielded the Cretaceous and younger ages, although subduction and accretionary prism growth might have been continuous between Permian and Cretaceous times (and younger, into Terrane A; see later).

2.5. Jurassic–Early Cretaceous: Gondwana Break-up Sequences

2.5.1. Jurassic terrestrial sedimentation in grabens [Botany Bay Group]

The Botany Bay Group comprises several geographically widely separated outcrops of terrestrial sedimentary rocks, mainly poorly bedded conglomerates (Fig. 2-3a). Although only known in Graham Land, there is no reason why other outcrops should not occur in Palmer Land, probably with a slightly older age (early Jurassic rather than mid Jurassic; see below). They rest on and are derived from erosion of deformed Trinity Peninsula Group metasediments and its stratigraphical equivalents and they are overlain by volcanic rocks of the Graham Land Volcanic Group. Successions in northern Graham Land comprise the Mt Flora Formation (at Hope Bay), View Point Beds, Camp Hill Formation, Downham Peak Beds and Tower Peak Formation (Farquharson 1982a, Farquharson 1982b, Farquharson 1983, Farquharson 1984, Elliot & Gracanin 1983, Birkenmajer 1993). There are also several less well known but locally extensive outcrops in east-central Graham Land (hinterlands of Cape Disappointment and Jason Peninsula; Hathway 2000, Riley & Leat 1999, Riley et al. 2010), Joinville Island and in the South Orkney Islands (Powell Island Conglomerate; Elliot & Wells 1982, Wells 1984, Cantrill 2000). The group is best known because of the presence of a historically important diverse fossil flora at Hope Bay that also contains bivalves, fish scales and beetles (e.g. Rees & Cleal 2004). The Camp Hill Formation contains abundant well-preserved leaves, logs, a tree in growth position, and thin coals (Farquharson 1984). The Powell Island conglomerate also contains a macroflora, although less diverse than elsewhere in the Botany Bay Group (Thomson 1973, Cantrill 2000). The northern Graham Land sequences vary in thickness from 124m (Tower Peak Formation) to 780 m (Camp Hill Formation), whilst the View Point and Downham Peak beds are just a few metres thick. Elsewhere, the sequence in the Cape Disappointment hinterland (the most extensive in the Botany Bay Group but least well known) is up to 600 m thick (Riley & Leat 1999) and the Powell Island Conglomerate is over 500 m. Rapid lateral thinning of some successions suggests that they are preserved in steep-sided hollows (graben?) on a land surface formed by the Trinity Peninsula Group and its correlatives.

The age of the Botany Bay Group has been contentious yet is critical for understanding the early history of the break-up of Gondwana prior to the development of the Larsen Basin. The floras at Hope Bay and Camp Hill have been assigned various ages ranging between Early Jurassic and Early Cretaceous (e.g. Stipanicic & Bonetti 1970, Rees & Cleal 2004) whilst that in the Powell Island Conglomerate is considered Early to Middle Jurassic by comparison with the Mt Flora Formation (Cantrill 2000). Conceptual arguments based on the tectono-stratigraphical development of the region have variously assumed an earliest Jurassic or latest Jurassic–earliest Cretaceous age (Macdonald et al. 1988, Farquharson 1982b, Farquharson 1984, Hathway 2000). However, direct U-Pb dating of volcanic zircons in a tuffaceous bed in the Camp Hill Formation and detrital zircons in conglomerate in the Tower Peak Formation has resolved the debate by yielding ages of 167 ±1 and 168.9±1.3 Ma, respectively, confirming a Middle Jurassic age for the Botany Bay Group in Graham Land (Hunter et al. 2005).

The individual sequences are dominated by basement-derived clast-supported conglomerates deposited from debris flows and braided streams on alluvial fans in fault-bounded basins (Fig. 2-9; Elliot & Wells 1982, Farquharson 1982a, Farquharson 1984, Elliot & Gracanin 1983, Wells 1984, Cantrill 2000). The Powell Island Conglomerate also contains landslide blocks of basement up to 30 m across (Elliot & Wells 1982, Wells 1984). The outcrops inland of Cape Disappointment are finer grained and might be lacustrine (Hathway 2000). Plant-bearing sandstones and mudstones are locally prominent, especially in the Mt Flora and Camp Hill formations, and are either sheet-flood or braided stream deposits. The Camp Hill Formation is unusually varied and contains a thick floodplain sequence with fluvial, overbank and lacustrine deposits (Farquharson 1982a). The evolution of the Hope Bay and Camp Hill formations, from coarse proximal deposits to more distal finer-grained sheet-flood and floodplain deposits suggests either a retreat of the bounding scarps against which the alluvial fans were banked or else a reduction in the relief of the source terrain with time (Elliot & Gracanin 1983). The Mt Flora and Tower Peak formations also contain volcanic interbeds (a variety of tuffs and 'agglomerate' (probably volcanic breccia)) mainly towards the top of the individual sequences, and the sediments contain detrital volcanic-derived garnet similar to garnetiferous lavas in the overlying volcanic sequences of the Middle Jurassic Mt Tucker Formation (Hamer & Moyes 1982, Riley et al. 2010) suggesting that the later stages of sedimentation overlapped with the initiation of volcanism in the Graham Land Volcanic Group (Riley & Leat 1999, Riley et al. 2010; see next section).

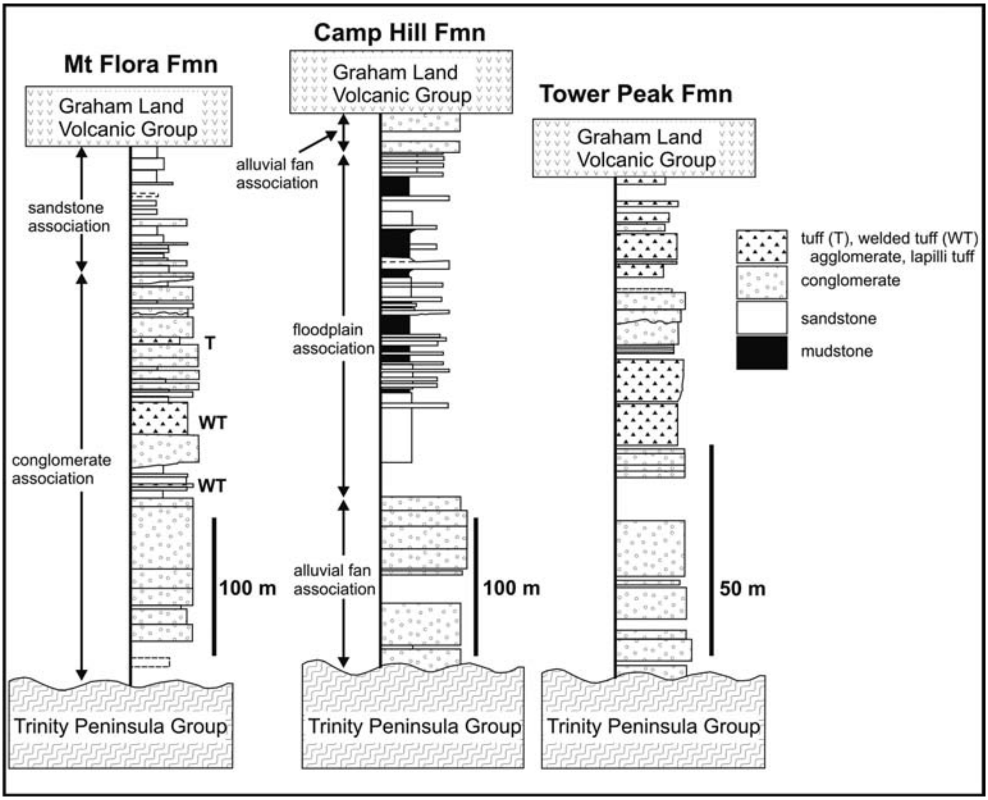


Fig. 2-9. Summary sedimentary logs for the Mt Flora, Camp Hill and Tower Peak formations in the Botany Bay Group (modified after Farquharson 1984).

2.5.2. Jurassic large-volume plume-related bimodal volcanism [Palmer Land Volcanic Group (new name) and Graham Land Volcanic Group]

A series of petrogenetically related but geographically widely separated Early-Middle Jurassic bimodal volcanic rocks crops out along the length of the Antarctic Peninsula (Fig. 2-3a, b). They comprise the Ellsworth and Graham Land volcanic groups, and the Brennecke and Hjort formations in Palmer Land (Rowley et al. 1982, Riley & Leat 1999, Hunter et al. 2006, Riley et al. 2010). With the repositioning of the geographical boundary between Palmer Land and Ellsworth Land, the name Ellsworth Land Volcanic Group has become anachronistic. It is therefore appropriate to rename the Ellsworth Land Volcanic Group as the Palmer Land Volcanic Group, and redefine it to include not only the previously assigned formations of the Ellsworth Land Volcanic Group but also the Brenneke and Hjort Formations (Fig. 2-4a). The age of the volcanism shows a well-defined step-change, from c. 189–183 Ma in the south (Palmer Land Volcanic Group) to c. 172–162 Ma in the north (Graham Land Volcanic Group). Together, they form part of the Jurassic Chon Aike large igneous province that also crops out extensively in southern South America (Pankhurst et al. 2000, Riley et al. 2001, Riley et al. 2018). The volcanism is regarded as a volcanic ‘flare-up’ that resulted in the emplacement of c. 0.5 million km^3 of silicic igneous volcanic and subvolcanic complexes, with three peaks of activity identified at c. 185, 170 and 155 Ma. The combined volcanism forms one of the most voluminous silicic large igneous provinces in the world (Riley & Leat 1999). Its origin has been attributed to lithospheric extension and crustal anatexis caused by the underplating and heating effects of mafic magmas linked to the progressively migrating front of a major mantle plume head (e.g. Wever & Storey 1992, Pankhurst et al. 2000, Riley & Knight 2001; Fig. 2-10). Plume migration effects might also have triggered a putative Jurassic failed rift identified in the south-western Weddell Sea (Jokat & Herter 2016, cf. Willan & Hunter 2005), and which may have affected the geological evolution of the south-eastern margin of the Antarctic Peninsula (Riley et al. 2018).

The Palmer Land Volcanic Group crops out adjacent to and within the western margin of the Latady Basin in southern Antarctic Peninsula (Hunter et al. 2006; Figs. 2-1, 2-3b). In southern Palmer Land it comprises the Mount Poster Formation, composed largely of silicic volcanic rocks, together with smaller outcrops of basaltic lavas and associated sedimentary lithofacies collectively called the Sweeney Formation (Rowley et al. 1982, Riley & Leat 1999, Hunter et al. 2006). The Sweeney Formation is locally at least 1 km thick and is dominated by abundant dark green to black basalt lava with distinctive trails of white amygdales and rare pillows. The basalt resembles more pervasively deformed and recrystallised amygdaloidal basalts forming the Hjort Formation in the Black Coast (E Palmer Land) described by Wever & Storey (1992; see also Storey et al. 1987) and the two may be time equivalent. Other minor volcanic lithofacies include ignimbrites, hyaloclastite, lapilli tuffs and tuffs. The sedimentary beds mainly comprise thinly bedded and laminated, dark-coloured, red-weathering fine sandstone and mudstone showing planar and ripple lamination and containing sparse but ubiquitous plant material, including silicified wood. The depositional setting for the Sweeney Formation was probably lacustrine for the sedimentary beds, and the presence of hyaloclastite, pillow lava and other characteristics (e.g. ripple cross laminated tuffs) suggest a similar environment for much of the associated volcanism (Hunter et al. 2006). The Mount Poster Formation is at least 500 m thick, although it may be as much as c. 2 km (Rowley et al. 1982, Hunter et al. 2006). It is dominated by poorly to densely welded crystal-rich silicic ignimbrite

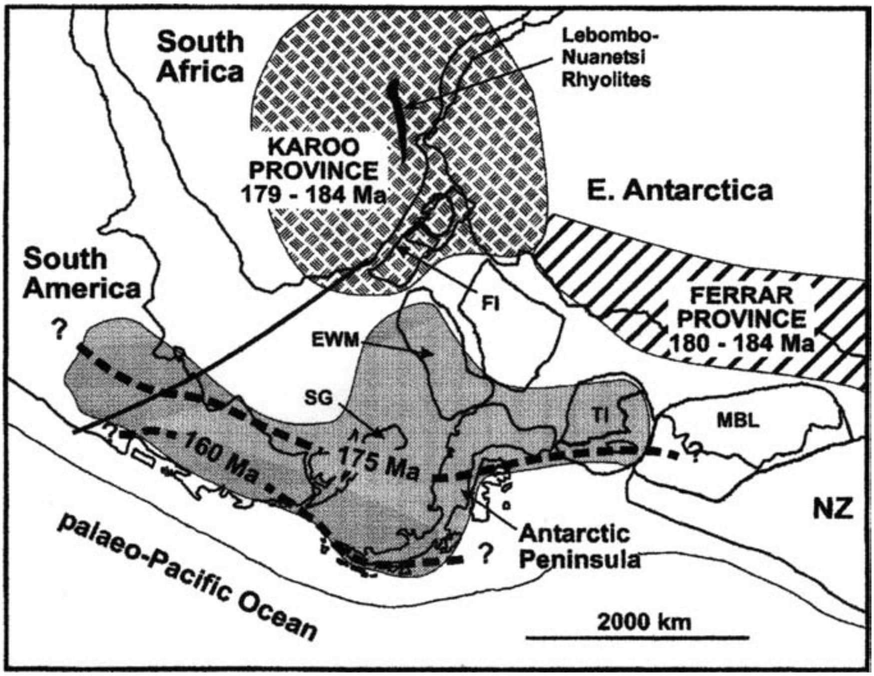


Fig. 2-10. Reconstruction showing large igneous provinces in southern Gondwana during the Jurassic (from Pankhurst et al. 2000). The heavy dashed timelines show how silicic flare-up volcanism in the Antarctic Peninsula (Palmer Land and Graham Land Volcanic groups) and southern South America migrated trenchwards with time.

variably coloured blue-grey, green or purple and showing distinctive ubiquitous small flecks of red mudstone, together with minor tuff and lavas. From their thickness, homogeneity, welding and faulted contacts with the Sweeney Formation, they are believed to have accumulated in an intracaldera setting. Several separate eruptive centres are postulated, and the outcrops may comprise overlapping or nested calderas. Rare high-aspect ratio (i.e. ratio of lateral extent to thickness) ignimbrites are interpreted as distal outflow (extra-caldera) units (Hunter et al. 2006).

In northern Palmer Land, the Palmer Land Volcanic Group is represented by the silicic Brennecke Formation and the mafic Hjort Formation (Wever & Storey 1992). The Brennecke Formation consists of dacite and rhyolite lavas, welded pyroclastic rocks (probably ignimbrites) and rare tuffs and black shale. They are variably deformed and locally occur as laminated schists and tectonites. The Hjort Formation is >150 m thick and comprises deformed metamorphosed mafic lava flows and/or sills now occurring as foliated amphibolitic greenstones. They contain conspicuous white amygdale trails (cf. Sweeney Formation, above). Also included in the Hjort Formation is metamorphosed andesitic lava interlayered with foliated tuffaceous breccia and laminated tuff, which are present at one locality (Kamenev Nunatak; Storey et al. 1987).

The Graham Land Volcanic Group is formed by the volumetrically dominant silicic Mapple Formation and correlatives, and the andesitic Mt Tucker Formation (Riley & Leat 1999, Riley et al. 2010; Fig. 2-3a). In its type area in the Oscar II Coast region, the Mapple Formation forms one of the largest outcrops of silicic volcanic rocks in the Antarctic Peninsula. In the Cape Disappointment hinterland, it reaches a maximum thickness of 1 km and has an estimated total volume there of c. 2000 km^3. Subaerial crystal tuffs and ignimbrites with a wide degree of welding (non-welded to rarely rheomorphic) dominate (c. 80–85%) the Mapple Formation, together with rhyolite lavas (c. 10%), co-ignimbrite breccias, bouldery debris flow deposits and minor sandstones and mudstones with locally abundant plant fragments, woody stems and tree trunks up to 50 cm in diameter (del Valle et al. 1997). Eruptive centres have not been identified and the rocks are metamorphosed up to greenschist grade. Other correlatives of the Mapple Formation in Graham Land are the Kenney Glacier Formation at Hope Bay composed of rhyolite/dacite ignimbrites, lavas and breccias (Birkenmajer 1993), outcrops on Joinville Island formed by >300 m of strongly welded ignimbrites, rhyolite lavas and tuffs (Riley et al. 2010), and intensely metasomatised silicic ignimbrites and lavas and up to 400 m of tuffs on Jason and Churchill peninsulas (Smellie 1991b, Riley et al. 2010). By contrast, the Mt Tucker Formation, exposed only in the Tower Peak and Camp Hill areas of north-eastern Graham Land, consists of sequences of 'agglomerates' (probably volcanic breccias) and less common andesite lavas, which reach a maximum thickness of c. 750 m (Riley et al. 2010). The andesites are distinctively garnetiferous (Hamer & Moyes 1982) and they overlie and/or are interbedded with terrestrial sedimentary rocks of the Botany Bay Group (also garnet-bearing). They are also compositionally distinctive and include high-magnesian andesites that might have been generated after subduction of young hot oceanic lithosphere following ridge–trench collision (Alabaster & Storey 1990).

2.5.3. Jurassic–Early Cretaceous basin sedimentation linked to Weddell Sea formation [Latady Basin]

The Latady Basin extends along most of eastern Palmer Land and offshore beneath the western Weddell Sea (Hunter & Cantril 2006; Fig. 2-1). It was originally interpreted as a back-arc basin related to a Jurassic arc (Laudon & Ford 1997) but is now regarded as an intracontinental basin formed by extension and formation of the Weddell Sea during the early break-up of Gondwana (Hathway 2000, Willan 2003, Willan & Hunter 2005, Hunter & Cantrill 2006). The sedimentary basin fill is represented by the mainly shallow-marine Latady Group and a likely correlative in northern Palmer Land called the Mount Hill Formation that is more strongly deformed and metamorphosed (Laudon et al. 1983, Meneilly et al. 1987, Hunter & Cantrill 2006). The Cape Framnes beds, exposed at the eastern extremity of Jason Peninsula, might be the most northerly correlatives of the Latady Basin (Riley et al. 1997). They are also regarded as basal strata of the Larsen Basin (Hathway 2000), although there is a significant age difference with the latter (Fig. 2-4b). The Latady Group is several kilometres thick, dominated by fossiliferous black siltstone and shale with less common pale-coloured fine sandstone and conglomerate, and rare coal. It is locally interbedded with, and probably overlies, volcanic facies of the Palmer Land Volcanic Group. Deformation is pervasive and involves well-developed cleavage and open to isoclinal folds up to a kilometre across that mainly verge to the south or south-east (Kellogg & Rowley 1989). The deformation has been ascribed to the compressional 'Palmer Land event' (terrane accretion?), to gravitational spreading during uplift and shouldering aside by a coeval magmatic arc, and to thrusting related to retro-arc subduction of Weddell Sea crust below Palmer Land (Kellogg & Rowley 1989, Grunow 1993, Vaughan & Storey 2000, Ferraccioli et al. 2006). The 'Palmer Land event' is dated at c. 107–103 Ma (Vaughan et al. 2002a, b).

The Latady Group has been divided into five formations, from base up: Anderson, Witte, Hauberg Mountains, Cape Zumberge and Nordsim formations (Hunter & Cantrill

2006). They are described in Table 2-3. The group youngs from north-west to south-east, apart from the youngest (Nordsim) formation, which crops out in the north-west. Deposition extended between Early/Middle Jurassic and earliest Cretaceous. The Jurassic Mount Hill Formation is a sparsely fossiliferous sequence of strongly deformed and metamorphosed metasedimentary rocks, mainly siltstones and mudstones of uncertain thickness (Meneilly et al. 1987). Deformed metavolcanic rocks associated with the Mount Hill Formation are silicic and include massive greenstones, and are now included here in the Palmer Land Volcanic Group (Brennecke and Hjort formations; Storey et al.

Table 2-3. Summary of the principal characteristics of formations in the Latady Group in southern Palmer Land (after Hunter & Cantrill 2006).

Formation name	Thickness (m)	Age	Lithofacies	Interpretation
Nordsim	> 120	Latest Jurassic/earliest Cretaceous	Fine sandstone and mudstone in repeating cycles, commonly with palaeosol tops; thicker medium to coarse sandstones; very rare conglomerate	Deltaic setting, with channelized waterways tens of metres wide flanked by muddy deposition and crevasse splays in overbank areas
Cape Zumberge	[uncertain]	Latest Tithonian	Black mudstone and very fine sandstone; lesser orange fine sandstone	Very quiet, low-energy conditions with restricted circulation (anoxic), probably outer shelf
Hauberg Mountains	Probably several km	Late Middle Jurassic to Kimmeridgian/Tithonian	Long Ridge Member (base; several hundred metres): dominated by cream/orange fine-medium sandstone, becoming finer and siltier up; rare fine conglomerates; Mount Hirman Member (>900 m): dark fine sandstone and siltstone and thick grey medium sandstone; rare mud pebble conglomerate; Novocin Member (top; >650 m): similar to the other members but greater proportion of dark fine sediments (mainly black siltstone and orange fine sandstone); rare coarse and fine conglomerates and very thin coals	Long Ridge Member: lower foreshore deposits becoming shoreface or shallow marine upward, and possibly subaerial at times; Mount Hirman Member: Quiet, low-energy marine environment with fluctuating energy conditions, possibly further offshore than in the Long Ridge Member; Novocin Member: Low-energy deposition and slow sedimentation rates on the middle to outer shelf below storm wave base
Witte	> 700 m	Late Middle Jurassic (Callovian); c. 165 Ma	Dominated by blue-black siltstone and claystone with lesser yellow/grey medium to fine sandstone; concretions; widespread small channels; includes the Mt Rex Witte tuff	Very quiet water low-energy offshore setting deposited in restricted to semi-restricted basins; sedimentation punctuated by turbulent currents; restricted circulation with periodic or continuously anoxic conditions
Anderson	> 1500 m	Late Early Jurassic? to early Middle Jurassic	Orange fine sandstone & minor mudstone in lower part becoming somewhat more common upward; thin conglomerate beds throughout; minor volcanic interbeds basally	Foreshore-shoreface setting with relatively high-energy shoaling wave conditions; volcanic beds include airfall tuff and block-and-ash deposits

1987, Wever & Storey 1992). The Kimmeridgian–early Tithonian Cape Framnes beds consist of highly fossiliferous volcaniclastic sandstone and lesser siltstone seen only as piles of scree (Riley et al. 1997).

The Latady Basin is believed to have evolved in a broad rift zone created during the early extensional stages of Gondwana disintegration (Willan & Hunter 2005; see also Jokat & Herter 2016; Fig. 2-11). The earlier formations have a mature quartzose petrology characteristic of recycled-orogen provenances and identified as most likely the Ellsworth–Whitmore Mountains, and volcanilithic grains derived from the Mount Poster Formation, which was active at that time. The youngest formation (Nordsim Formation; Hunter & Cantrill 2006) also contains a substantial proportion of felsic grains and volcanic quartz but the source for the fresh volcanic detritus, believed to be from an active continental margin arc, is uncertain and may have been the relatively distal Thurston Island block a few hundred km to the south-west. The shift from extension-related to arc-related sources is thought to reflect changes in the large-scale tectonics, which evolved from dominantly plume-related uplift and silicic volcanism to resumption of a continental margin arc-dominated setting. The phase of marine deposition represented by the Cape Zumberge Formation was under anoxic, sediment-starved deep-water conditions in a basin that stretched along the entire eastern margin of Palmer Land, and the sequence is capped inland by deltaic sediments deposited in small isolated terrestrial lacustrine basins situated in the north-west and west (Nordsim Formation).

2.6. Cretaceous–Palaeogene "Andean" Arc–Trench System and Early Glacial Stages

2.6.1. Jurassic–Cenozoic accretion [LeMay Group and Scotia Metamorphic Complex (Terrane A)]

The LeMay Group is a thick variably deformed and metamorphosed sedimentary sequence that crops out exclusively on Alexander Island. It has a known depositional age of Early Jurassic–Early Cretaceous and youngs in a westerly direction (Pimpirev & Sinnyovsky 1989, Holdsworth & Nell 1992). Carboniferous (Namurian) and possibly Permian (Gzhekian–Artinskian) fossils also occur at the eastern outcrop limit (Mt King beds; Kelly et al. 2001) and those strata are presently included in the LeMay Group although the geological affinities are uncertain (see later). The sedimentary rocks in the LeMay Group are arkosic, with an arc-like provenance, and they represent a variety of sedimentary gravity flow deposits, mainly turbidites, deposited in submarine systems at an active continental margin (Burn 1984, Tranter 1991). They form two main sedimentary lithofacies associations: conglomerate–sandstone–mudstone characteristic of eastern LeMay Range outcrops and probably deposited in an inner fan channel, on levées and in interchannel areas; and thinly-bedded sandstone–mudstone in western LeMay Range outcrops, which probably represent deposition in suprafan or lower fan settings (Tranter 1991). Sedimentary strata known as the Light Sandstone Member situated on the eastern edge of the LeMay Range at Atoll Nunataks and Mount Umbriel are distinguished by unusual light brown-grey relatively coarse sandstone and lesser conglomerate interbedded with more typical dark (and finer) sandstone and mudstone, the light coloration probably being diagenetic (Doubleday et al. 1993). Two other lithofacies associations are recognised: radiolarian chert–basalt and basalt–tuff. The latter association is geographically

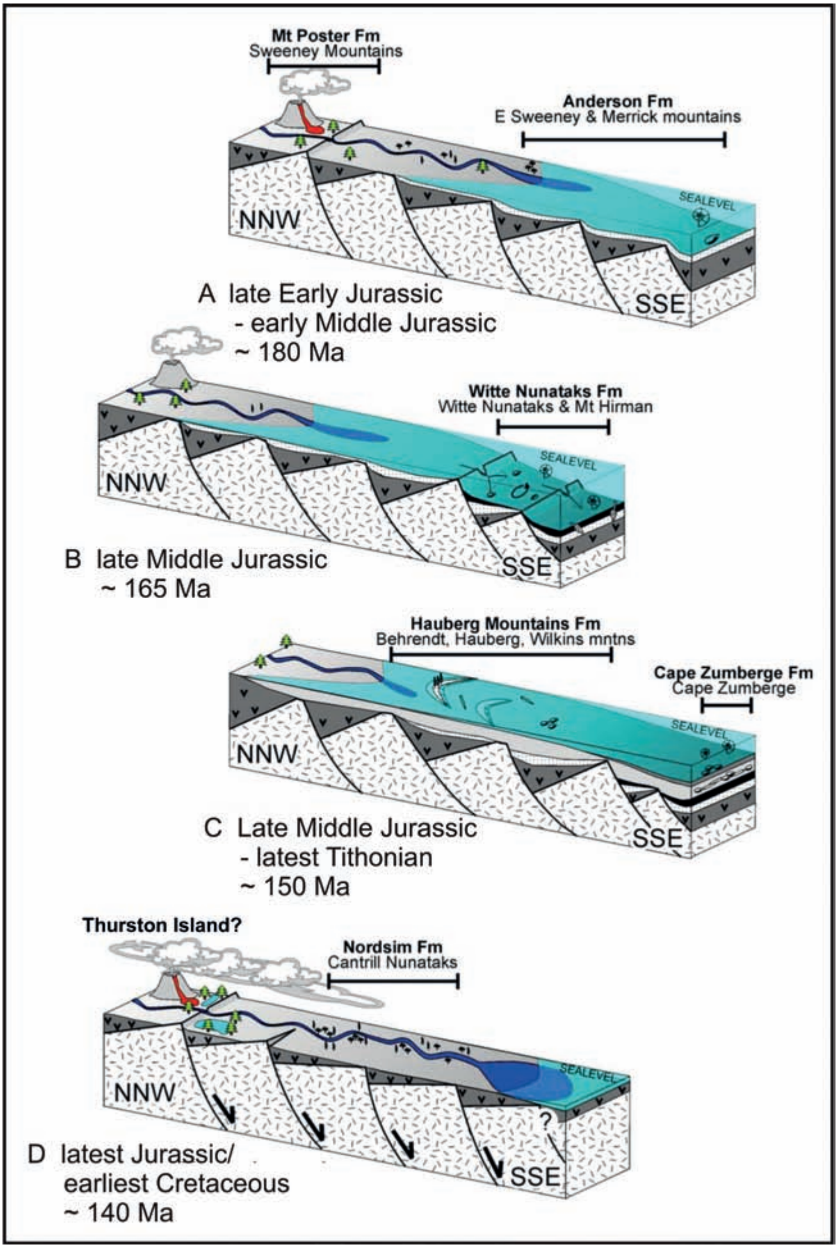


Fig. 2-11. Schematic reconstructions showing the sequential development of the Latady Basin during the Jurassic (from Willan & Hunter 2005). Note that the ages have been changed and the Nordsim Formation has been repositioned following the revision by Hunter & Cantrill (2006).

widespread (>55 km^2) and is distinguished as the Lully Foothills Formation (Burn 1984). It contains pillow lava units up to 30 m thick, hyaloclastite breccias and a variety of (shallow) waterlain tuffs, and it yielded Early Jurassic (Sinemurian) ammonites (Thomson & Tranter 1986). A major tectonic boundary comprising a >150 m-wide mélange zone separates the Lully Foothills Formation from other LeMay Group outcrops. In the eastern LeMay Range, early structures are deformed by easterly (arcward-) verging folds and thrusts whilst, to the west, upright folds are deformed by a late phase of westward-directed thrusting. The metamorphic grade broadly increases from west to east. It is predominantly prehnite-pumpellyite and pumpellyite-actinolite facies but locally reaches greenschist grade in the north-east and is transitional to blueschist facies in metabasalts at two localities (Burn 1984). Sedimentation and deformation were diachronous (younging oceanwards to the north-west) and overlapped with deposition in the fore-arc basin Fossil Bluff Group (Holdsworth & Nell 1992). Taken together with features such as syn-sedimentary bedding disruption and sediment mobilization, a general lack of penetrative fabrics, and the suggestion of high P-T metamorphism, the LeMay Group has been assigned an accretionary complex origin (e.g. Suárez 1976, Smellie 1981, Burn 1984, Doubleday & Tranter 1992, Tranter 1991, Holdsworth & Nell 1992; Fig. 2-12). The Light Sandstone Member has a generally lower degree of metamorphism and deformation and these rocks might represent the top of a frontally accreted slice or the upper fill of a trench-slope basin (Holdsworth & Nell 1992, Doubleday et al. 1993). The Lully Foothills Formation has a geophysically modelled crustal root extending c. 4 km and it is probably an accreted oceanic seamount, whilst strata of the chert-basalt lithofacies association are probably thrust-bound slivers of oceanic crustal material (Tranter 1991, Doubleday et al. 1994).

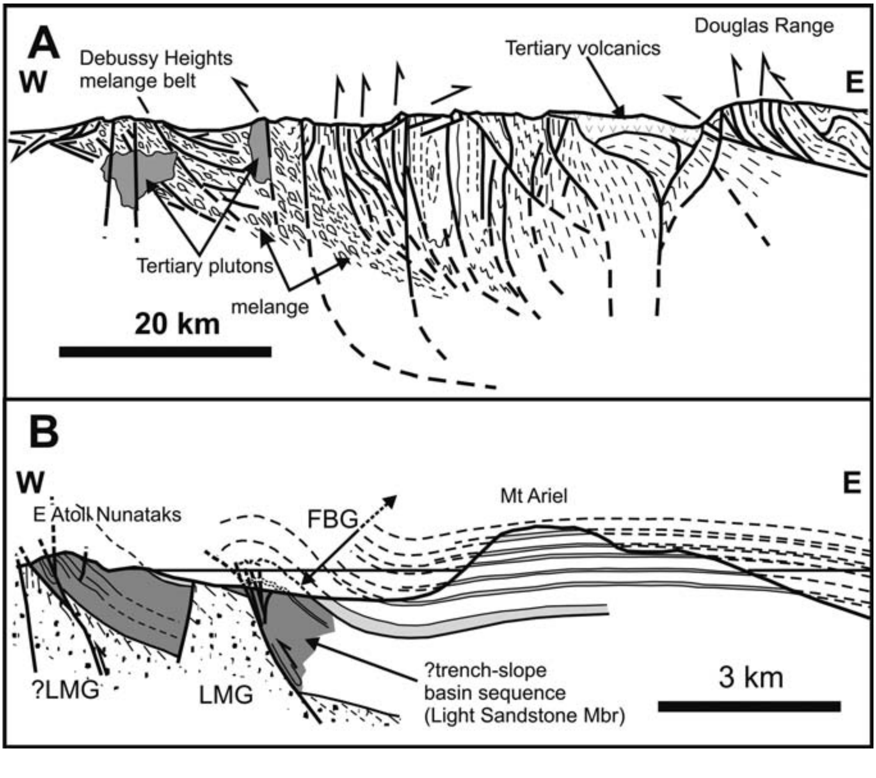


Fig. 2-12. A. Schematic true-scale cross section through the LeMay Group in northern Alexander Island illustrating the principal structural characteristics of the accretionary complex. B. Cross section depicting relationships between the accretionary complex (LeMay Group (LMG)) and upper slope and trench slope basin sediments of the Fossil Bluff Group (FBG; both figures modified after Holdsworth & Nell 1992).

Terrane A of the Scotia Metamorphic Complex crops out in northern Elephant Island, Clarence Island and Smith Island (South Shetland Islands). It is composed of phyllite and greenschist, with blueschists restricted to Smith Island and central Elephant Island (Smellie & Clarkson 1975, Rivano & Cortés 1976, Tanner et al. 1982, Dalziel 1984, Trouw et al. 1991, Grunow et al. 1992). Mafic to intermediate volcaniclastic conglomerate, sandstone and siltstone, volcanic breccia, chert and pelite are also present on the north coast of Elephant Island (Trouw et al. 1991, Grunow et al. 1992). The rocks form a lower temperature metamorphic assemblage than in Terrane B, but reflect similarly high pressures. Three metamorphic zones are identified on Elephant Island, comprising (from north to south) a low-grade chlorite zone in which relict sedimentary and volcanic textures are still evident, a garnet zone and a biotite zone; the latter two zones are characterised by schists. Boundaries between zones are gradational and they cut obliquely across the supposed tectonic boundary with Terrane B metamorphic rocks (cf. Tanner et al. 1982, Dalziel 1984, Trouw et al. 1991). Terrane A is predominantly formed of chlorite- and (lesser) garnet-grade rocks whilst the biotite zone falls entirely within Terrane B together with a large part of the garnet zone. Clarence Island falls within the chlorite zone whilst the abundant blueschists on Smith Island probably belong to the garnet zone. P-T estimates based on metabasites suggest pressures of c. 7 kb and temperatures ≥ 350 °C (Grunow et al. 1992). Most of the protoliths are typical of an ocean floor environment (e.g. including pelagic to hemipelagic sediments, manganiferous chert, limestone and probable oceanic pillow basalt).

At least three deformational phases are recognised in Terrane A (Smellie & Clarkson 1975, Dalziel 1984, Grunow et al. 1992, Trouw et al. 1991). Trouw et al. (1991) suggested that large-scale D2 structures verge predominantly to the south or south-east, which is difficult to reconcile with subduction from north to south and might be related to sinistral strike-slip movements along the Shackleton Fracture Zone. However, according to Grunow et al. (1992), whilst main-phase folds on Smith Island also verge predominantly to the south-east, the main-phase folds on Elephant Island are predominantly upright and deformation intensity increases from north to south. D2 structural trends cross-cut the metamorphic isograds at a high angle, consistent with overprinting by the metamorphic zones. Grunow et al. (1992) suggested that a simple prograde metamorphism is absent and that greenschist facies assemblages overprinted the blueschists, with retrograde metamorphism of the amphibolite facies schists. The age of the metamorphism on Smith Island is 54 (± 14)–47 Ma. It is 122 (± 51)–69 Ma on Elephant Island but with a concentration of ages around 109–101 Ma, and there is a systematic (though small) increase in age from south to north (Tanner et al. 1982, Grunow et al. 1992 and references therein).

The observed northerly decrease in metamorphic grade, deformational intensity and age across Elephant Island, from Terrane B to Terrane A, is continued into the Seal Islands. Known as the Seal Islands Formation, the rocks are mainly matrix-supported conglomerate (pebbly mudstone) and siltstone, with lesser volcanic greywacke and mudstone (Trouw et al. 1991, Grunow et al. 1992). An irregular cleavage is present and parts of the sequence resemble a mélange-type broken formation, consisting of disrupted greywacke clasts in mudstone (Dalziel 1984). Miocene – Recent foraminifera occur in siltstone and some clasts.

2.6.2. Jurassic–Cretaceous fore-arc basin sedimentation [Fossil Bluff Group and correlatives]

The Fossil Bluff Group is superbly exposed in eastern Alexander Island. It represents the sedimentary infill of a fore-arc basin situated between the magmatic arc in Palmer

Land and the LeMay Group accretionary complex to the west. It ranges in age from Kimmeridgian to late Albian. Other likely correlative sequences are exposed at Carse Point on the east side of George VI Sound, Adelaide Island, Biscoe Islands, Low and Livingston islands (South Shetland Islands) and in the South Orkney Islands (eastern Coronation Island and Mathews Island). The Carse Point and Low Island sequences are Upper Jurassic (Tithonian?) and early Late Jurassic–early Cretaceous (late Oxfordian–Berriasian?), respectively. On Livingston Island, the Anchorage Formation is Kimmeridgian–early Tithonian and the President Beaches Formation is Berriasian. Apart from rare and undiagnostic radiolaria and trace fossils, the Biscoe Islands outcrops are unfossiliferous and their age is uncertain but probably Late Jurassic–Early Cretaceous. The two Adelaide Island formations are Kimmeridgian–Tithonian (c. 150 Ma) and Aptian–Albian (c. 114 Ma) in age whilst the South Orkney outcrops are Early Cretaceous.

The Fossil Bluff Group is the thickest and best exposed fore-arc sequence in the region, extending as a 30 km-wide strip that extends 250 km in eastern Alexander Island (Macdonald et al. 1993). It exceeds 6.8 km in thickness and the oldest strata are exposed in the centre-west and north (Middle–Late Jurassic), becoming younger southward (to mid Cretaceous (Albian)) accompanied by a facies change from deeper to shallower marine and finally up into terrestrial conditions (Fig. 2-13; Butterworth & Macdonald 1991, Moncrieff & Kelly 1993, Nicholls & Cantrill 2002). Basal contacts are faulted (LeMay Range Fault or, more accurately, fault zone) or are unconformable against the LeMay Group accretionary complex in the west, but are obscured beneath George VI Sound in the east (Fig. 2-12; Doubleday et al. 1993). The Fossil Bluff Group is divided into eight formations, from base up (Fig. 2-14): Selene Nunatak, Atoll Nunataks, Ablation Point, Himalia Ridge, Spartan Glacier, Pluto Glacier, Neptune Glacier and Triton Point, which are described in Table 2-4. Most of the group is a deep- to shallow-marine deposit laid down during basin shallowing in response to local tectonism in the fore-arc region (Butterworth 1991). Deposition on the upper trench slope began in the Early to Middle Jurassic, with the fore-arc basin (sensu stricto) forming later, from Late Jurassic time (Doubleday & Storey 1998). Several major deformational events affected the Fossil Bluff Group, comprising (1) Middle–Late Jurassic transtensional strike-slip faulting (LeMay Range Fault) that led to emergence of the underlying LeMay Group accretionary complex and deposition of the Selene Nunatak Formation; (2) Late Jurassic–Early Cretaceous basin inversion under dextral transpressional conditions, probably responsible for the development of spectacular large slump units in several of the formations (Macdonald et al. 1993); (3) mid Cretaceous compression ('Palmer Land event') that caused regional uplift, final emergence and the transition to terrestrial conditions (Triton Point Formation); and (4) Late Cretaceous or Tertiary post-inversion extension that created the linear graben of George VI Sound, again in a dextral transpressional setting and possibly related to the evolution of the West Antarctic Rift System (Nell & Storey 1991, Doubleday & Storey 1998, Nicholls & Cantrill 2002, Eagles et al. 2009).

Moving north, the only extensive outcrops of likely fore-arc basin marine sedimentary rocks are on Adelaide Island (Riley et al. 2012, also Griffiths & Oglethorpe 1998). The Buchia Buttress Formation is composed of >830 m of volcanic breccia, silicic tuff, coarse volcaniclastic sandstone and cobble/boulder conglomerate deposited rapidly in a shallow marine apron fan environment. The Milestone Bluff Formation is at least 1.5 km thick and dominated by turbiditic sandstone deposited in a transitional to shallow marine environment, with subordinate cobble and boulder conglomerate, ignimbrite and tuff. The volcanic beds are mostly subaerial and the conglomerates were probably linked to

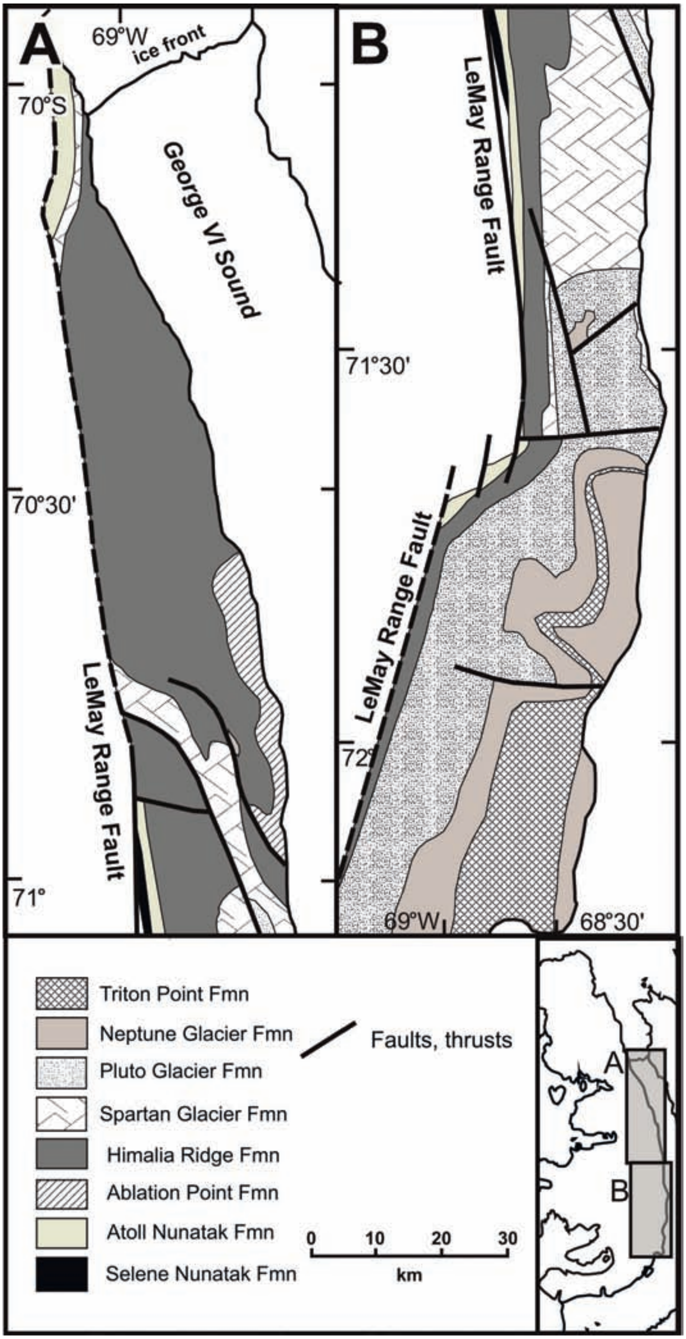


Fig. 2-13. Geological map showing the distribution of sedimentary formations in the Fossil Bluff Group. The formations get younger up-section from north (Selene/Atoll fmns) to south (Triton Point Fmn); inset shows the location of the maps in eastern Alexander Island (modified after Doubleday & Storey 1998 and Nichols & Cantrill 2002).

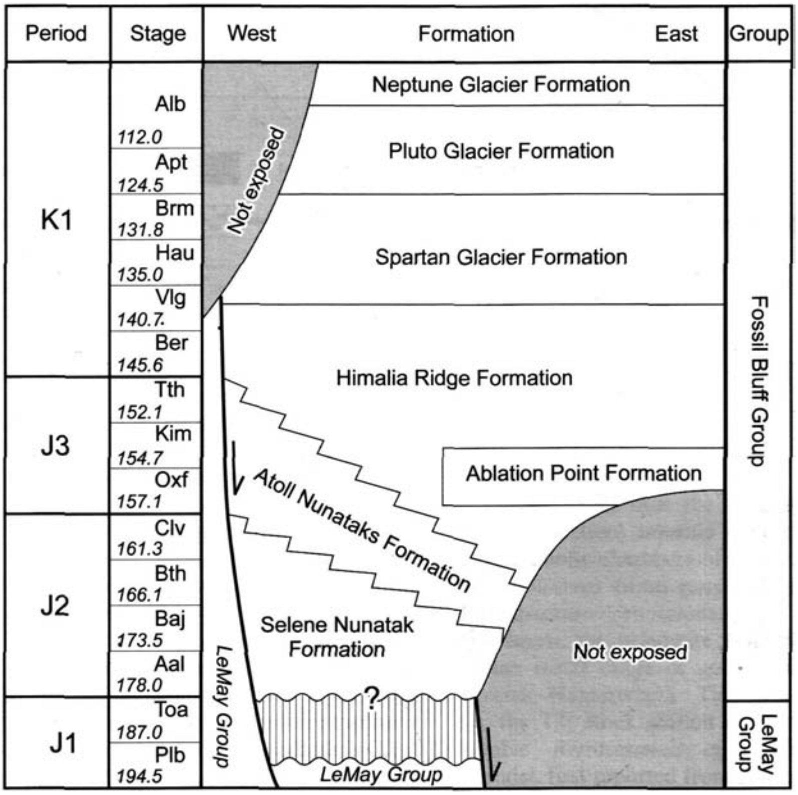


Fig. 2-14. Lithostratigraphy of the Fossil Bluff Group, eastern Alexander Island (from Macdonald et al. 1999).

alluvial fans. Both formations are locally fossiliferous. The Carse Point sequence might also have been deposited in the Fossil Bluff fore-arc basin on its eastern margin flanking Palmer Land. It is a thin-bedded sequence >60 m thick dominated by fossiliferous mudstones and shales with subordinate volcaniclastic sandstones and conglomerate lenses (Culshaw 1975).

Other possible fore-arc basin correlatives include un-named sequences composed of thin- and thick-bedded sandstone–siltstone–mudstone, and fine to very coarse sandstones and granule conglomerates sparsely present in the Biscoe Islands and deposited predominantly by sediment gravity flows (Smellie et al. 1985), and on Low Island (South Shetland Islands; Smellie 1979) and possibly the Anchorage Formation (Byers Group) on Livingston Island (Hathway & Lomas 1998). The Low Island sequence consists of well-bedded volcanic-sourced mudstone, tuff and lapillistone in laterally continuous even beds (probably mainly turbidites; Smellie 1979, Thomson 1982a). The lapillistones include unwelded ignimbrite and the volcaniclastic deposits were sourced in coeval silicic (dacitic) volcanism, with sedimentation directly from ash fall and from density currents, and the poorly fossiliferous mudstones representing normal hemipelagic background sedimentation between eruptions. The Anchorage Formation is currently regarded as part of the Byers Group. It is exposed only on westernmost Livingston Island and comprises two members in faulted contact, with a total thickness of c. 120 m (Hathway

Table 2-4. Summary of the principal characteristics of formations in the Fossil Bluff Group, eastern Alexander Island (after Butterworth et al. 1988, Doubleday et al. 1993, Moncrieff & Kelly 1993, Nicholls & Cantrill 2002).

Formation name	Thickness (m)	Age	Lithofacies	Interpretation
Triton Point	c. 900–1200	Albian	Very coarse & pebbly sandstone with at least 13 fossil forest horizons with trees up to 5 m tall in life position; sparse interbedded tuff & lapillistone	Fluvial, including braidplain channel-fill and overbank deposits; rare shallow marine deposits (tidal, upper shoreface); splits Neptune Glacier Formation into two parts
Neptune Glacier	2200	Albian	Dominated by coarse to gravelly sandstone and conglomerate; rare mudstone; two members that sandwich the Triton Point Formation: Deimos Ridge (lower) and Mars Glacier (upper)	Nearshore/paralic; Mars Glacier Member records a major transgressive event on the Triton Point Formation
Pluto Glacier	1400	Aptian-Albian	Mainly mudstone and thin interbedded fine to medium sandstones; major slump zone locally; coarse sandstone and conglomerate in Rhea Corner Member; rare green crystal tuffs in upper part	Tidal shallow shelf; Rhea Corner Member beds are high-density sediment gravity flow deposits in a channelized turbidite system
Spartan Glacier	1000	Valanginian-Barremian	Thick monotonous discontinuous sequences 2-20 m thick of mudstone and siltstone with subordinate thin fine sandstone beds; syn-sedimentary melanges up to 120 m thick with rare folds and faults	Outer shelf to slope
Himalia Ridge	2200	lower Tithonian-Berriasian	Wide range of facies with marked lateral variation; channelled conglomerate complexes 80-170 m thick, mudstone, interbedded sandstone/siltstone, thick-bedded sandstone, pebbly mudstone; slump sheets, folds & syn-sedimentary faults common; coeval sills & pillow lava with bimodal OIB-like compositions	Deep-marine submarine fan channel complex
Ablation Point	> 440	Kimmeridgian	Alternating irregular melange zones (small sandstone blocks in sheared mudstone matrix) c. 200 m thick & up to 2 km long alternating with sheets of coherent strata up to 500 m long & 50 m thick; wide range of slump-folded and rafted blocks of sedimentary & volcanic rocks arranged chaotically within one huge slump zone; the blocks are mainly interbedded sandstones & mudstones, ignimbrites, lavas, tuffs, & rare breccias & conglomerates	Slope collapse slump units; series of very thick, very large folded sheets defining a syn-sedimentary duplex; the melange zones acted as local decollement horizons

Atoll Nunataks	1000–1200 (probably 1050)	Bathonian?-Tithonian	Thin-bedded mudstone and silty mudstone; single thin limestone bed; diminished LeMay Group provenance	Hemipelagic deposition on the upper trench slope; marine transgression and trench-slope deposition, representing a phase of subsidence probably in response to tectonism of the underlying accretionary prism; overlain by the Himalia Ridge Formation
Selene Nunatak	> 115	Bathonian?	Pebble-cobble conglomerates and sandstone mainly derived from the LeMay Group accretionary complex	Marine; marks the emergence and erosion of the inner fore-arc area; LeMay Group source exposed to the **east**

& Lomas 1998). Both members are composed of dark grey mudstones, thin pale tuffs and an upward increasing proportion of thin volcaniclastic sandstones but the New Plymouth Member (older?) is pervasively bioturbated whilst Punta Ocoa Member sediments are unbioturbated and laminated. Most of the mudstones are hemipelagic suspension deposits, whilst the tuffs and sandstones are probable turbidites introduced during storms. The sedimentation took place in relatively deep quiet water below wave base under oxygenation conditions that changed upward from anaerobic to dysaerobic-aerobic (Pirrie & Crame 1995). Although considered to be a correlative of the Nordenskjöld Formation deposited in a broad shelf sea with either a discontinuous or no emergent Antarctic Peninsula arc, the proposed setting is hard to reconcile with the evidence for an extensively subaerially exposed Peninsula, comprising the terrestrial Palmer and Graham Land Volcanic groups and terrestrial sedimentation in the Botany Bay Group, and anoxic conditions prevailing in an open-ocean fore-arc position (cf. Farquharson 1983, Pirrie & Crame 1995). An alternative possibility is that the Anchorage Formation might have been deposited in a silled fore-arc basin situated between a shallow trenchward barrier(s) formed by uplifted accretionary complex rocks and the emergent Peninsula landmass nearby, restricting circulation and promoting a stratified water column and anoxia. The unconformably overlying President Beaches Formation is >600 m thick, composed mainly of dark grey laminated mudstones, minor lenticular packages of pale sandstone (turbidites) and clay-altered tuffs, and shows abundant evidence of syn-sedimentary deformation. It was deposited as a marine slope apron below storm wave-base under dysaerobic conditions. The sea floor was relatively steep and unstable, resulting in slumping and other soft-sediment deformation.

The Spence Harbour Conglomerate (South Orkney Islands) is also tentatively included. It comprises an association of debris flow and sheet-flood deposits, rare landslide blocks and braided stream sandstones composed of angular fragments derived from the underlying metamorphic basement, deposited mainly in a mid-fan position in a series of coalesced alluvial fans related to fault scarps (Elliot & Wells 1982, Wells 1984). Although terrestrial, the conglomerates prograde over 200 m of sparsely fossiliferous shallow-marine sandstones on Mathews Island and they overlie the 2 m-thick, fossiliferous Gibbon Bay Shale (a pebbly mudstone) on Coronation Island, also marine (Thomson 1974), suggesting that they represent the trenchward, mainly subaerial margin of a fore-arc basin (cf. Selene Nunatak and Triton Point formations, Fossil Bluff Group; e.g. Doubleday et al. 1993).