# A TECTONIC RECONSTRUCTION OF ACCRETED TERRANES

# ALONG THE PALEO-PACIFIC MARGIN

# OF GONDWANA

by

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Abstract

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The southern oceanic margin of Gondwana was nearly 40,000 km long or 24,854.8 miles. The southern margin was the result of the Terra Australis orogen. Spanning 18,000 km or 11,184.7 miles and  is proposed as one of the largest and longest lived orogens in Earth history. The paleo-Pacific margin of Gondwana consisted of segments of the Australian-Antarctic craton, southern South America (modern Argentina and Chile), southern South Africa, Marie Byrdland, New Zealand and its adjacent continental shelf, the Ellsworth Mountains, and the Transantarctic Mountains. The process of terrane accretion has played a substantial part in the assembly of the continents as they look today. The paleo-Pacific margin of Gondwana was an active region of terrane accretion from the Neoproterozoic to the Late Mesozoic.

This research study examines the accretion of terranes across the paleo-Pacific Gondwana margin to provide a comprehensive reconstruction. A paleogeographic basemap was created using PALEOMAP Project maps and the geology data was provided by the School of Geoscience from the University of Witwatersrand of South Africa. Location and data analyzed for terranes were collected building a PDF library of journal articles across numerous geological publications.

Table of Contents

List of Illustrations

List of Tables

# Chapter 1

## Supercontinents

Austrian geologist Eduard Suess named the supercontinent Gondwana after observing the Upper Paleozoic and Mesozoic formations in the Gondwana region of central India are similar to formations of the same age on the other southern hemisphere Gondwanan continents. The notion that all of the continents of the southern hemisphere were once amalgamated to produce a supercontinent was proposed by German meteorologist Alfred Wegener in 1912 (Lotha, 2013). He put forth the idea of a single landmass, Pangaea, of which Gondwana was the southern section. South African geologist Alexander Du Toit expanded the idea of a Gondwanan supercontinent in his book titled "*Our Wandering Continent"*, written in 1937 in which he documented evidence that suggested possible supercontinent assembly. The book outlined the occurrence of tillite (glacial) deposits of Permo-Carboniferous age and similar flora and fauna that are not found in the northern hemisphere. One particular faunal example used by Du Toit was the seed fern *Glossopteris* that is found in Permian age formations all along the southern hemisphere including South Africa, India, South America, E. Australia, and Antarctica (Lotha, 2013).

Gondwana, also at one time referred to as Gondwanaland, is the name given to the southern supercontinent of Pangaea (the northern supercontinent is known as Laurasia) (Lotha, 2013). Gondwana was likely sutured between 570 to 510 Ma, which joined East Gondwana to West Gondwana. There were interior continental collisions that were occurring at the same time with a change from passive margin sedimentation to convergent margin activity along the paleo Pacific margin. Subduction that commenced along the paleo Pacific margin ranges in age from 540 to 550 Ma (Buchanan, 2004). Subduction is inferred by the oldest supra-subduction zone plutons along the Antarctic portion of the margin. Ophiolites were produced by the supra-subduction zone which conserved a phase of extension in greenstone sequences in eastern Australia and comes before the Ross-Delemerian orogenesis between 520 to 490 Ma, inboard of the plate margin, and simultaneously with final internal-suturing of Gondwana. Emplacement of supra-subduction zone igneous bodies continued throughout this period indicating that subduction was on-going (Buchanan, 2004). Between 320 to 230 Ma, major tectonic plate boundary restructuring was followed by regional orogenesis alongside the paleo Pacific margin. Convergent margin movement concluded about 305 Ma. in E. Gondwana and was supplanted by extension and strike-slip activity until about 260 Ma (Buchanan, 2004). Convergence resumed alongside the plate margin which is demonstrated by the emplacement of supra-subduction zone igneous bodies into the Paleozoic accretionary prism in eastern Australia. The Gondwanide orogeny occurred between 305 to 230 Ma and coincided with the phase of plate alteration (Buchanan, 2004).

Southern Gondwana has been reconstructed since the arrival of plate tectonic modeling but, its origin is of some question. Southern Gondwana consisted of portions of the Australian-Antarctic craton, southern South America (modern Argentina and Chile), southern South Africa, Marie Byrdland, New Zealand and its adjacent continental shelf, the Ellsworth Mountains, and the Transantarctic Mountains. The southern oceanic margin of Gondwana was nearly 40,000 km long or 24,854.8 miles. The southern margin was the result of the Terra Australis orogen. Spanning 18,000 km or 11,184.7 miles, it is proposed as one of the largest and longest lived orogens in Earth history (Cawood, 2005; Vaughan et al., 2005). The Terra Australis orogen currently consists of the eastern third of Australia, New Zealand, West Antarctica, the Transantarctic Mountains and large portions of southern South America (Vaughan and Pankhurst, 2008).

## Previous Work on Comprehensive Reconstructions of the Southern Gondwana Margin

Comprehensive reconstructions of the accretionary history of the paleo Pacific Gondwana margin have been difficult due to numerous extenuating factors that will be further discussed in this thesis. The goal of this thesis is to focus on the history and construction of the paleo-Pacific margin of southern Gondwana, i.e., southern South America, southern South Africa, East and West Antarctica, eastern Australia, Victoria Land and New Zealand. The intent is to provide an overview of the tectonic history of the accretionary terranes of southern Gondwana, providing a model as it applies to the Australide accretionary orogeny along the Pacific margin of southern Gondwana.

Recent publications have reviewed the construction of portions of the paleo-Pacific margin of Gondwana. South America has recently been reviewed by e. g.,, Ramos (1984, 1986, 1988); Astini et al. (1995); Rapela et al. (1998b); Pankhurst et al. (1998b); Vaughan and Storey (2000); Ferraccioli et al. (2006) Alonso et al. (2008). Pankhurst et al. (2006). Southern South Africa has been reviewed by e. g., Dalziel et al. (2000); Johnston (2000); Shone and Booth (2005); Pankhurst et al. (2006). Antarctica has been reviewed by e. g., Pankhurst et al. (1998b); Vaughan and Storey (2000); Tessensohn and Henjes-Kunst (2005); Ferraccioli et al. (2006). New Zealand has been reviewed by Coombs et al. (1976); Bishop et al. (1985); Bradshaw (1989); Cooper (1989); Muir et al. (1996); Waight et al. (1998); Muir et al. (1998); Mortimer et al. (1999); Campbell (2000). Australia has been reviewed by e. g., Li et al. (1997); Veevers et al. (1994); Li and Powell, (2001); McElhinney et al. (2003); Vaughan et al. (2005). More comprehensive works of Gondwana have been done by Veevers (2004); Cawood (2005); Casquet et al. (2008); Vaughan and Pankhurst (2008).

The aim of this thesis is to answer questions regarding the tectonic evolution of the southern Gondwana margin based on evidence provided by accreted terranes. This study hopes to add to the growing scientific knowledge of the southern margin of Gondwana in relation to a completed comprehensive tectonic model for the Pacific margin. The results of this study should provide valuable information geologists when devising plate models for future research of the southern Gondwana margin. Armed with greater, more detailed information of the region provided by the results of this study, future research articles will be able to provide a more complete tectonic history. Additionally, future research could be aimed towards looking at more specific terranes within the southern margin to finish an inclusive model of the southern margin.

# Chapter 2

## Methods of Data Acquisition and Figure Creation

Several authors (e. g., Meert, 2001; Cawood, 2005; Casquet et al., 2008; Vaughan and Pankhurst, 2008) over the last decade have published articles detailing a tectonic reconstruction of Gondwana and/or the paleo-Pacific margin of Gondwana. Whereas these articles provide an excellent framework for this thesis, the goal herein is to provide a comprehensive and detailed look at the entire southern margin of Gondwana. This chapter describes in detail the methods of data acquisition and the creation of the terrane map of southern Gondwana.

A PDF library of the literature was created with the goal to assemble published journal articles for each region of southern Gondwana including those papers that provide a comprehensive review of the entire southern margin. Using these published articles, several tectonic maps and figures were constructed that were beneficial to the completion of this paper. The objective was to produce a revised tectonic map identifying the accreted terranes of the southern Gondwana margin.

An Excel spreadsheet was created following the completion of the PDF library. Information from several articles was used to populate the fields of the spreadsheet. The idea of the Excel file was to gather a list of important tectonic features and information that would be relevant into the creation of a terrane map and the writing of this thesis. The list included 71 entries that are defined by several fields. The fields include the “Name”, “Abbreviation” of the name, type of “Tectonic Feature”, the “Country”, “Latitude” and “Longitude” given in decimal degrees, the range of age from “Old Age” to “Young Age” given in millions of years, and the “Period”. Information was gathered to transfer into a database on ArcGIS by ESRI in the creation of a terrane map.

To create the map an ArcGIS database for southern Gondwana that combines data obtained through a review of the literature and the Internet was used. The dataset titled “Gondwana Merge” is a collection of geologic information field provided by Nelson Mandela Metropolitan University. The important fields are described below. The first field of note is the self-titled “PERIOD.” The field describes the geologic Period for each location. The next two fields go hand in hand. They are the “MAX SUBERA” and the “MIN SUBERA.” Both fields provide a possible range of Subera’s over which each location was deposited. The next field is “EPOCH.” The field provides the name of the Epoch during deposition. The next two fields are the “MAX CHRON” and the “MIN CHRON.” The two fields provide the range of numerical age values. Ages are given in millions of years. The next field is titled “EON.” That describes the Eon during deposition. The next fields are called “MAX ERA” and “MIN ERA.” Both fields describe the range of Era’s over which each location was deposited. The next field is “ROCK CLASS.” The field describes whether the rocks are igneous, metamorphic, or sedimentary. The next field is titled “ROCK CATEGORY.” that describes whether the rocks are intrusive or extrusive. The next field is called “ROCK TYPE.” and is more specific on the rock’s mineralogy. The fields mentioned above were combined to form a comprehensive dataset which is projected in ArcGIS using the “Lambert Azimuthal Equal Area” projection. A separate column of data was created in the attribute table that combines “ROCK TYPE” and “PERIOD” to properly assign colors to the map based on geology and age.

Using ArcMap, a base map was created displaying accretionary terranes of southern Gondwana as the main focus. Figure 2-1 shows the locations and each region where the paper focused the research. The terranes are georeferenced and are defined by set of attributes including name, age, period and rock type if available. The data was gathered based on information and maps journal articles in the PDF library. The map features several geological and tectonic features/structures that are significant to terrane accretion events. Geologic features on the map include ophiolite locations, Late Precambrian Margin, Mesozoic Magmatic Arc, Ordovician Arc, Paleozoic Orogen, Paleozoic-Mesozoic Orogen, Rift Zones, Subduction Zones, Cratons, and Accreted Terranes. Table 2-1 provides a list of authors whose articles were used to gather the data relevant to the map. Over the course of the research and writing this paper, every effort has been made to avoid errors. Unfortunately and predictably, errors are part of the process and information cannot be considered 100% accurate. There are a couple issues worth noting as a possibility for errors. Several papers create their own tectonic maps. In doing so, boundaries for features like terranes and/or cratonic rock vary. In addition, multiple authors have different interpretations of terrane boundaries or the existence of terranes that remain questionable. The goal of this paper is to show the terrane boundaries as accurately as possible utilizing the information that is accepted through a peer reviewed, general consensus including mentioning controversial terranes. Another area of potential error is a lack of information available for certain features. For example, in putting together the database through Excel some of the information was incomplete. Some tectonic features were lacking distinct age ranges and periods. Also, this paper tries to accurately represent each author’s data and interpretation as close to accurate as possible. However, a misunderstanding of the author’s explanation and/or examination of the data is a possibility.

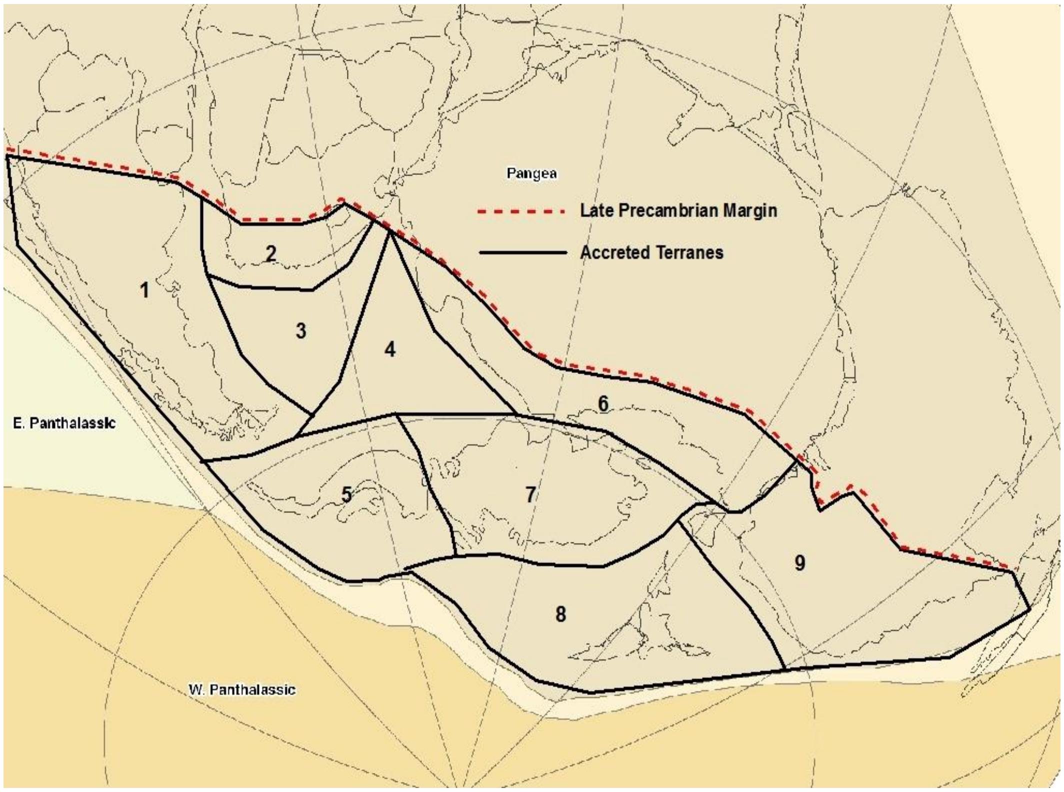


Figure 2-1 A reconstruction of the world 360 Ma showing the paleo-Pacific margin of southern Gondwana. Each number corresponds to a region of interest. The regions are as follows: 1)Patagonia; 2) Cape Fold Belt; 3) Falkland Plateau; 4) Ross Ice Shelf; 5) Palmer Peninsula; 6) Trans-Antarctic Mountains; 7) Marie Byrd Land); 8) New Zealand and 9) eastern Australia.

Table 2-1 Regions of Southern Gondwana with Probable Terranes

| Table 1. | | | |
|---|---|---|---|
| | **Region** | **Probable Terranes** | **References** |
| | 1. Patagonia | Chilena, Cuyania (Precordillera), Pampia, Patagonia | e. g., Ramos (1984, 1986, 1988); Astini et al. (1995); Rapela et al. (1998b); Alonso et al. (2008); Vaughan and Pankhurst (2008); Pankhurst et al. (2006) |
| | 2. Cape Fold Belt | Cratonic Rock | e. g., Dalziel et al. (2000); Johnston (2000); Shone and Booth (2005); Pankhurst et al. (2006) |
| | 3. Falkland Plateau | paleo-Pacific Ocean | e. g., Pankhurst et al. (1998b); Vaughan and Storey (2000); Ferraccioli et al. (2006) |
| | 4. Ross Ice Shelf | paleo-Pacific Ocean | e. g., Pankhurst et al. (1998b); Vaughan and Storey (2000); Ferraccioli et al. (2006) |
| | 5. West Antarctic Peninsula | Central Domain (possible smaller terranes), Western Domain | e. g. Pankhurst et al. (1998b); Vaughan and Storey (2000); Ferraccioli et al. (2006) |
| | 6. Trans-Antarctic Mountains | Ross Orogeny terranes | e. g., Tessensohn and Henjes-Kunst (2005) |
| | 7. Marie Byrd Land | Ross Province, Amundsen Province | e. g., Pankhurst et al. (1998b); Vaughan and Storey (2000); Ferraccioli et al. (2006) |
| | 8. New Zealand | Buller terrane, Takaka terrane, Brook Street terrane, Murihiku terrane, Dun-Mtn.- Maitai terrane, Caples terrane, Torlesse terrane | e. g., Coombs et al. (1976); Bishop et al. (1985); Bradshaw (1989); Cooper (1989); Muir et al. (1996); Waight et al. (1998); Muir et al. (1998); Mortimer et al. (1999); Campbell (2000); Vaughan and Pankhurst (2008) |
| | 9. Australia | Tasman Fold Belt terranes | e. g., Li et al., (1997); Veevers et al. (1994); Li and Powell, (2001); McElhinney et al. (2003); Vaughan et al. (2005); |

## Chapter 3

## Terranes

The idea of terranes came about in the 1970s from studies in the Western or Canadian orogenic margin of North America. Coney et al. (1980) presented the terrane model idea in his paper titled "*Cordilleran suspect terranes*" to describe tectonic and stratigraphic setting of the North American plate. The definition of terrane can vary depending on where one looks but it is generally referred to as fragments of continental crust that have been separated from continental landmasses as a result of rifting and which preserves the geologic history from its place of origin.

Ancient terrane accretion is thought to be similar to the processes that can be observed around present day orogens. The southwest Pacific contains active arc-continent collisions which are thought to be similar to the origin of various ancient terranes, as well as how they are emplaced along the continental margins (White, 2004). Sequences of continental, oceanic, and island arc material at subduction zones are carried along by subducting oceanic lithosphere and eventually obstruct the subduction zone as a consequence of their positive buoyancy (White, 2004). Thrust faults uplift the forearc and accretionary wedge as the collision begins, carrying or obducting them onto the continental margin. A new trench may form on the oceanward side of the impeded trench if subduction is slowed or stopped. The accretionary process will begin again (White, 2004).

Many ridges, rises and plateaus that comprise about 10% of the ocean basins at the present are thought to potentially be the origin of some terranes (Ben-Avraham et al., 1981). It is thought that many of these regions are extinct island arcs, microcontinents that were submerged and large igneous provinces. Positive buoyancy of these structures may constrain subduction trenches resulting in their accretion to the continental margin as *exotic* terranes An exotic terrane is a fragment that has broken off from a parent continent and accreted to another continent (White, 2004). Also, many exotic terranes are thought to have formed from processes related with the formation and breakup of supercontinents. There are two other geologic mechanisms of terrane accretion to continental margins, and continental growth: 1) the obduction of ophiolites and growth by magmatism andand sedimentation and 2) backarc, intraarc and forearc basins being generated and destroyed (White, 2004).

Accretionary tectonic processes like terrane accretion are identified in orogens by the occurrence of ophiolites (Metzler, 2007). Ophiolites are vertical cross-sections of oceanic crust that have been removed or obducted during a tectonic event. The characteristic layers from bottom to top are: 1) mantle-ultramafic cumulates; 2) moho-seismic rocks; 3) mafic cumulates; 4) massive gabbro; 5) mafic sheeted dike complex; 6) pillow basalt and; 7) pelagic/abyssal cherty sediments (Metzler, 2007). There are varying models on ophiolite obduction. Ophiolite obduction generally occurs during the thrusting of oceanic lithosphere onto a passive continental margin during continental collisions (Coleman, 1977; Pearce, 2003). Another method of ophiolite obduction happens when oceanic lithosphere breaks off the top of the descending slab and leaves a portion of the slab to be placed onto the continent. The lower portion of the oceanic lithosphere is then thrust under the continental slab (Coleman, 1977; Pearce, 2003). Emplacement can also occur when a slab of oceanic crust is added to an accretionary prism in an arc system. As the descending slab is subducted, the portions of the oceanic lithosphere break off and are emplaced onto the adjacent arc (Coleman, 1977; Pearce, 2003). As Metzler (2007) points out, the most common environments for ophiolite emplacement include accretionary prisms, continent-continent collisions, and arc-continent collisions.

Continental growth can also occur by the addition of magma and sedimentation, as well as collision and accretion of exotic terranes, as seen in southeastern Australia, the Lachlan orogen (White, 2004). The Paleozoic aged Lachlan orogen is an example of an accretionary orogen that has grown by more than 700km, mostly as a result of the aforementioned processes (Foster and Gray, 2000; Collins, 2002; Glen, 2005). The region lacks many features that are representative of major collisional orogens; however; there is a large volume of granitoid rock and volcanic sequence and extensive low grade quartz-rich turbidite which overlies thin continental crust and mafic lower crust of oceanic origin (Fergusson and Coney, 1992). The Lachlan orogen has preserved a geologic history of ocean-continent convergence lasting around 200 million years (Foster et al., 1999). Large extensional basins, up to 1000km wide and which contain a floor of basalt and gabbro, formed behind one or more island arcs that eventually accreted onto the margins of the continent (Glenn, 2005).

The Lachlan orogen contains backarc and intraarc extension cycles. The backarc and intraarc cycles produce thin, hot lithosphere. According to a model proposed by Collin (2002) a zone of intra-arc extension advances as an outcome of a subducting slab roll back. The Tuapo volcanic zone on the North Island of New Zealand is an equivalent tectonic setting (White, 2004). A backarc basin and remnant arc is created as the arc splits apart and drifts away from the trench, leading to crustal thinning and subsidence. Decompression melting generates basaltic crust as mafic magma underplates and intrudes the thinned crust. The subduction zone flattens and the upper plate of the orogen is compressed. The orogen region is possibly the onset of an oceanic plateau or an island arc at the subduction zone (White, 2004). Backarc basins are closed by contraction, possibly leading to the accretion of the arc and forearc to the margin of the continent. If a thick sediment sequence fills the basin, a short lived orogen forms.

Extension begins again and a new arc-backarc system forms along the continental margin once the plateau has been subducted (Kearey et al., 2009).

The evidence for an accreted terrane can come from several different avenues. The geologic history of a terrane will be different from the geologic history of the host continent. Varying geologic features indicate an exotic origin for the terrane. More specifically, geologists would look at the following factors to determine if the landmass is a terrane and to deduce its place of origin (Dawes and Dawes, 2001). These factors include but are not limited to fossil evidence, paleomagnetism, isotope ratio data, presence of ophiolites, and overlap formations and stitching plutons (Dawes and Dawes, 2001). Fossil evidence provides geologists a comparative analysis of the faunas present within the terrane and the host landmass. If the terrane contains faunas that are not present and/or do not match the continents fossil sequence, this is an indication the terrane is allochthonous (Dawes and Dawes, 2001).

Paleomagnetism occurs when magnetized minerals form within rocks. These minerals will preserve the direction of the earth’s magnetic field at the time of formation by aligning toward the earth’s magnetic poles. Paleomagnetic data will reveal the approximate latitude (the position relative to the equator and poles) at which the magnetized rock was formed. Paleomagnetic data will only work with beds where the original horizontal position was known. Measurements work best on lava flows or sedimentary beds that show their original horizontal bedding (Dawes and Dawes, 2001). Isotope data will provide evidence about terrane accretion in regions where rocks have been buried deep in the crust, metamorphosed, or melted and recrystallized as granite. Isotope geochemistry considers the natural variations in the relative abundance of isotopes of various elements. The variations of specific isotopes will divulge evidence about the ages and origins of rock, air or water bodies, or the processes of mixing between them (Dawes and Dawes, 2001). The importance of ophiolites was discussed above. Faults that separate an accreted terrane from its surroundings can only be seen if they are exposed on the surface (Dawes and Dawes, 2001). There are some regions where the contact of a terrane with the adjacent rock is not exposed due to the contact being buried beneath younger sedimentary deposits or intruded by igneous plutons (Dawes and Dawes, 2001). A sedimentary deposit that buries the contact of the terrane with the adjacent rock is called an overlap formation. A pluton that has intruded and obscured the contact of a terrane with adjacent rock is called a stitching pluton (Dawes and Dawes, 2001). Both features can be used to estimate how long ago a terrane was accreted to a region. Accretion must have taken place prior to the deposition of an overlap formation or the intrusion of a stitching pluton (Dawes and Dawes, 2001). Accretion must have occurred after the youngest rocks within the terrane had formed. Therefore, a terrane must have accreted after its youngest rocks had been formed and before its bordering faults were buried or intruded (Dawes and Dawes, 2001).

The following is a list of common terrane types described in the literature: 1) *Suspect terranes* are terranes where their origin is not clear; 2) *Native terranes* are terranes are terranes that are clearly related to the continental margin against which they presently reside; 3) *Exotic terranes*, as mentioned previously, are terranes that have broken off from a parent continent and accreted to another continent; 4) *Accreted terranes* are small crustal fragments, island arc, or seamount which are transported by a moving oceanic plate and added to a continental mass at the subduction zone; 5) *Displaced terranes* are the same as *Exotic terranes*; 6) *Allochthonous terranes* are the same as *Exotic terranes*; 7) *Composite terranes* are terranes made up of two or more distinct terranes that were joined together before their relocation; 8) *Superterranes* are an amalgamation of numerous smaller terranes; 9) *Disrupted terranes* are terranes that contain rocks from varying backgrounds and ages i. e., oceanic crust, shallow-water limestone, deep-water chert, and greywacke. *Disrupted terranes* are usually found to contain matrix of shale or serpentinite; 10) *Metamorphic terranes* are terranes that show signs of terrane-wide geologic changes, before or after collision, which have been dominant enough to hide original rock formations; 11) Contiguous terranes are the same as *Native terranes*; 12) *Tectonic terranes* stress the tectonic setting of the terrane; and 13) *Tectonostratigraphic terranes* stress the tectonic and stratigraphic setting of the terrane (Colpron et al., 2007).

Chapter 4

The Southern Gondwana Margin

South America

Southern South America is defined by several individual terranes (Ramos, 1988), which had their origins from pre-existing continental crust. The most studied terrane is the Precordillera terrane of central western Argentina, also referred to as Cuyania. The geology of the terrane is defined by Cambrian to Middle Ordovician limestone beds which are succeeded by an unconformity of Silurian-Devonian clastics that pass upward into Gondwana lacustrine deposits and Carboniferous to Triassic red beds (Vaughan and Pankhurst, 2008). Current research results suggest that the Precordillera terrane is exotic to the Gondwana margin and had its origins from Laurentia (Ramos, 2004). The terrane detached from Laurentia during the Cambrian and was transferred to western Gondwana during the Early to Middle Ordovician. The data indicate that the Precordillera terrane finally accreted to the protomargin of Gondwana during the Late Ordovician (Thomas and Astini, 2000; Ramos, 2004). Figure 4-1 shows the Argentine Precordillera rifting from the Ouachita embayment of Laurentia and its eventual accretion to the western margin of Gondwana. The sequence of lacustrine deposits and red beds are described by Alonso et al. (2008), as structural and sedimentological evidence for a passive margin (Vaughan and Pankhurst, 2008). Benedetto (1998) and Astini et al. (2004) demonstrate a distinct change in Cambrian brachiopod and trilobite faunas of Laurentian origin to a Middle-Late Ordovician fauna. *Salterella macculllochi*, a small conical shaped fossil of Early Cambrian age, provides the strongest paleobiogeographic evidence that link the Precordillera terrane to Laurentia. *Salterella macculllochi* has not been found anywhere in Gondwana. *Salterella* existed for a very brief interval according the fossil record, but spread suddenly around Laurentia. The occurrence of *Salterella* strongly correlates the

Precordillera terrane's derivation to Laurentia (Astini et al., 2004). The Famatinian magmatic arc, located in south-central Argentina, is indicative of the Precordillera terrane Laurentian origin. The amphibolite-facies metamorphism that affects the magmatic arc in the La Lampa province is attributed to the collision of the Precordillera terrane against the western margin of Gondwana (Chernicoff et al., 2010). Pankhurst et al. (1998) and Pankhurst et al. (2000) present evidence of I and S type granites developed on the marginal continental crust of Gondwana within the Famatinian arc.

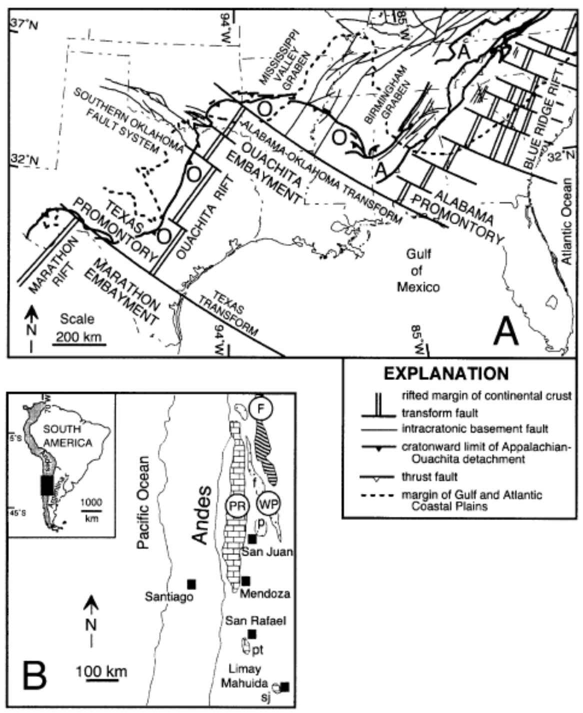


Figure 4-1 (A) An Outline Map presented by Thomas and Astini (2003) of the Late Precambrian-Cambrian rifted margin of Laurentia showing the Ouachita embayment and its relative location to Late Paleozoic thrust belts (A = Appalachian; O = Ouachita). (B) Map showing the present day location of the Precordillera (PR = Precordillera; F = Famatina; WP = western Sierras Pampeanas; f = Sierra de Valle Fértil; p = Sierra de Pie de Palo; pt = Ponon Trehue; sj = Cerro San Jorge).

The Precordillera terrane contains two K-bentonite ash band layers within the Lower Cambrian to Middle Ordovician age carbonate. A geochemical analysis (Huff et al., 1998) of the layers demonstrates the likelihood that the Famatinian arc volcanoes provided the ash for the K-bentonite bands within the carbonate suggesting that the Precordillera terrane was in close proximity to the Famatinian volcanoes during the deposition of the carbonate. The geochemical analysis of the K-bentonite bands supports the Middle Ordovician collision model (Fanning et al., 2004). Figure 4-2, below, shows K-bentonites within the San Juan Formation. Additional evidence supporting the Middle Ordovician collision model include paleomagnetic data detailed by Rapalini (2005); Middle Ordovician metamorphism in rocks east of the Precordillera (Casquet et al., 2001; Vujovich et al., 2004). Examples are summarized nicely by Vaughan and Pankhurst (2008).

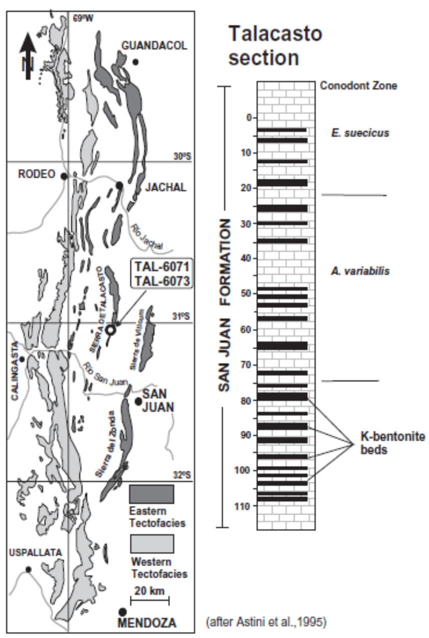


Figure 4-2 Map and cross section of Cambro-Ordovician limestone outcrops in the Argentine Precordillera. (Presented by Fanning et al., 2004).

Within the last decade, multiple papers have again proposed an alternative origin of the Precordillera terrane (Aceñolaza et al., 2002; Finney et al., 2003; Peralta et al., 2003), raising questions regarding its Laurentian origin. An alternative beginning has been proposed claiming the terrane is para-autochthonous to Gondwana. The source of the para-autochthonous origin of the Precordillera terrane dates back to Baldis et al. (1989), who argued that the Precordillera terrane had a para-Gondwanic origin and proposed a hypothesis that stated the Precordillera terrane was displaced by strike-slip transcurrent faults from the Pampean region to the south. The proposal recognized a displaced terrane in the Precordillera that's associated with the Pie de Palo dragged block, shown in Figure 4-3. (Ramos, 2004).

The crustal basement that underlies the Precordillera terrane is subject of debate as well. Precordilleran basement presents two possible scenarios for the Precordillera's origin. The basement is of Grenvillian age and contains xenoliths brought up in the Precordilleran Miocene volcanic rocks (Casquet et al., 2001; 2005; 2006). Zircon and isotopic data from the xenoliths indicate that the Precordillera basement is unique to South America and best explained geologically as having its origins from Laurentia. High grade metamorphic rocks in the Pie de Palo Range (McDonough et al., 1993; Varela et al., 1996; Pankhurst and Rapela, 1998; Casquet et al., 2001; 2005; 2006) to the east of the Precordillera also show similar results (Kay et al., 1996). Grenvillian age high-grade granite gneiss in Ponon Trehue (Heredia, 2002; Cingolani et al., 2005) and Grenvillian age tonalities in Las Matras (Sato et al., 2000) are all consistent with a Laurentian origin hypothesis.

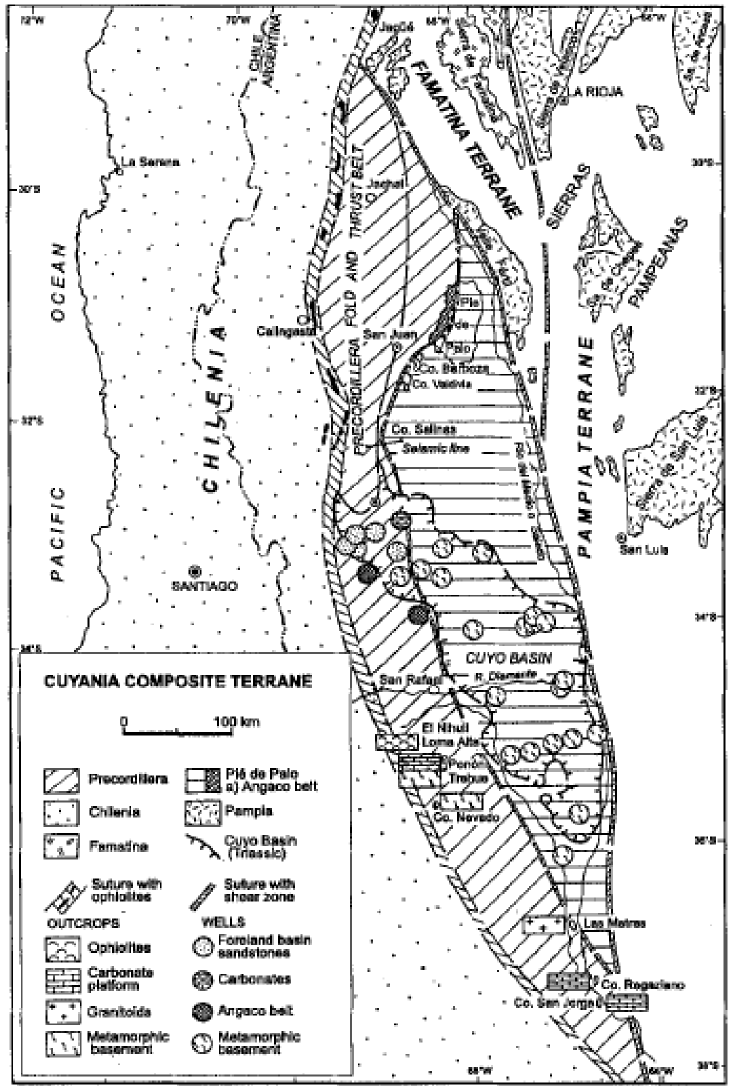


Figure 4-3 Outline of the Precordillera (Cuyania) terrane showing the Pie de Palo block (Ramos, 2004).

The Sierras Pampeanas provide an indication of additional terrane accretion. The Sierras Pampeanas is the location of the Pampia terrane (Ramos, 1988). Rapela et al. (1998) conducted a study of a 500 km segment of the Sierras Pampeanas (Figure 4-4) and documented two tectono-magmatic episodes that are pre-Silurian. Both events led to micro-continental collisions against the proto-Andean margin of Gondwana. According to data compiled by Pankhurst et al. (1998), Rapela et al. (1998) and Gradstein and Ogg (1996), collision of the Pampean terrane can be geologically documented during the Cambrian. Terrane accretion is supported by widespread anatexis and S-Type granites in the Eastern Sierras Pampeanas. There is Early-to-Mid Cambrian syn-metamorphic D2 deformation and S2 regional foliation (Gradstein and Ogg, 1996; Pankhurst et al., 1998; Rapela et al., 1998). Low grade metamorphism in the Eastern Cordillera and Famatina System to high grade metamorphism in the Eastern Sierras Pampeanas is visible. MORB-type ophiolite obduction is known in the Eastern Sierras Pampeanas (Gradstein and Ogg, 1996; Pankhurst et al., 1998; Rapela et al., 1998). Guereschi and Martino (2008) discuss two migmatization events in the Sierras Pampeanas. Data from the study of the two migmatization events concludes a Neoproterozoic-Early Paleozoic migmatization event that mirrors the subduction and collision of the Pampia terrane against the western Gondwana margin. The migmatization event is older than events proposed by Rapela. The timeline of geologic events put forth by Guereschi and Martino (2008) is unharmonious with the Paleoproterozoic Rio de la Plata craton to the east and the passive margin limestone in the Precordillera to the west Guereschi and Martino (2008) suggesting an exotic terrane accretion event. Nd isotope ratios put these rocks as Mesoproterozoic in age (Vaughan and Pankhurst, 2008). Because the Famatinian rocks and Rio de la Plata craton have a Mesoproterozoic signature, several authors including Rapela et al. (1998) concluded the evidence for metamorphism must have originated from a foundation to the east as a foreland sequence above an eastward dipping subduction zone. No Mesoproterozoic source is exposed so, it was suggested that the terrane had been previously rifted off from a similar position on the Gondwana margin during the Neoproterozoic (Rapela et al., 1998). The terrane was proposed to be similar to the Arequipa-Antofalla blocks of northern Chile and Peru (Vaughan and Pankhurst, 2008). Subduction of a spreading ridge in the Middle Cambrian was proposed by Simpson et al. (2003), Schwartz and Gromet (2004), Rapela et al. (2007) and Verdecchia et al. (2011), as an alternative to collision of a continental block.

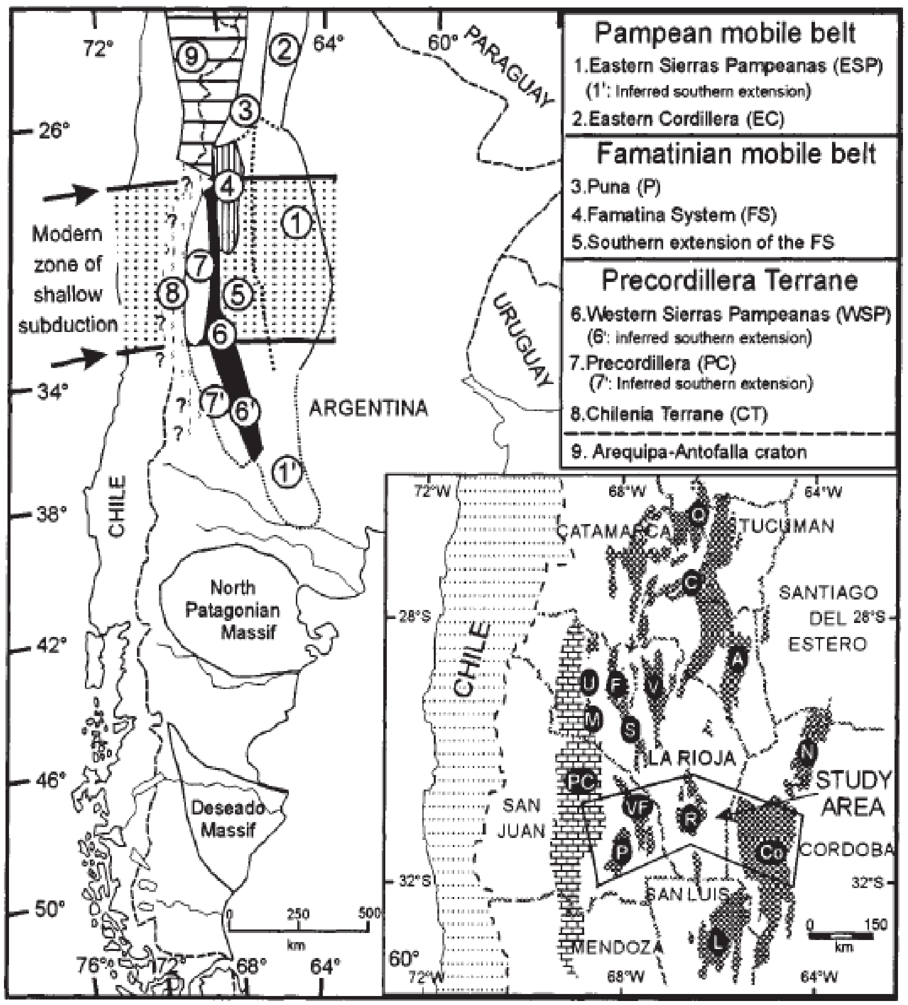


Figure 4-4 Early Paleozoic geological provinces in southern Andes (Q = Quilmes; C = Capillitas; A = Ancasti; F = Famatina; V = Velasco; U = Umango; M = Maz; S = Sañogasta; N = Norte de Córdoba; Co = Córdoba; L = San Luis; R = Llanos de la Rioja; VF = Valle Fertil; P = Pie de Palo; PC = Precordillera. Rapela et al., 1998).

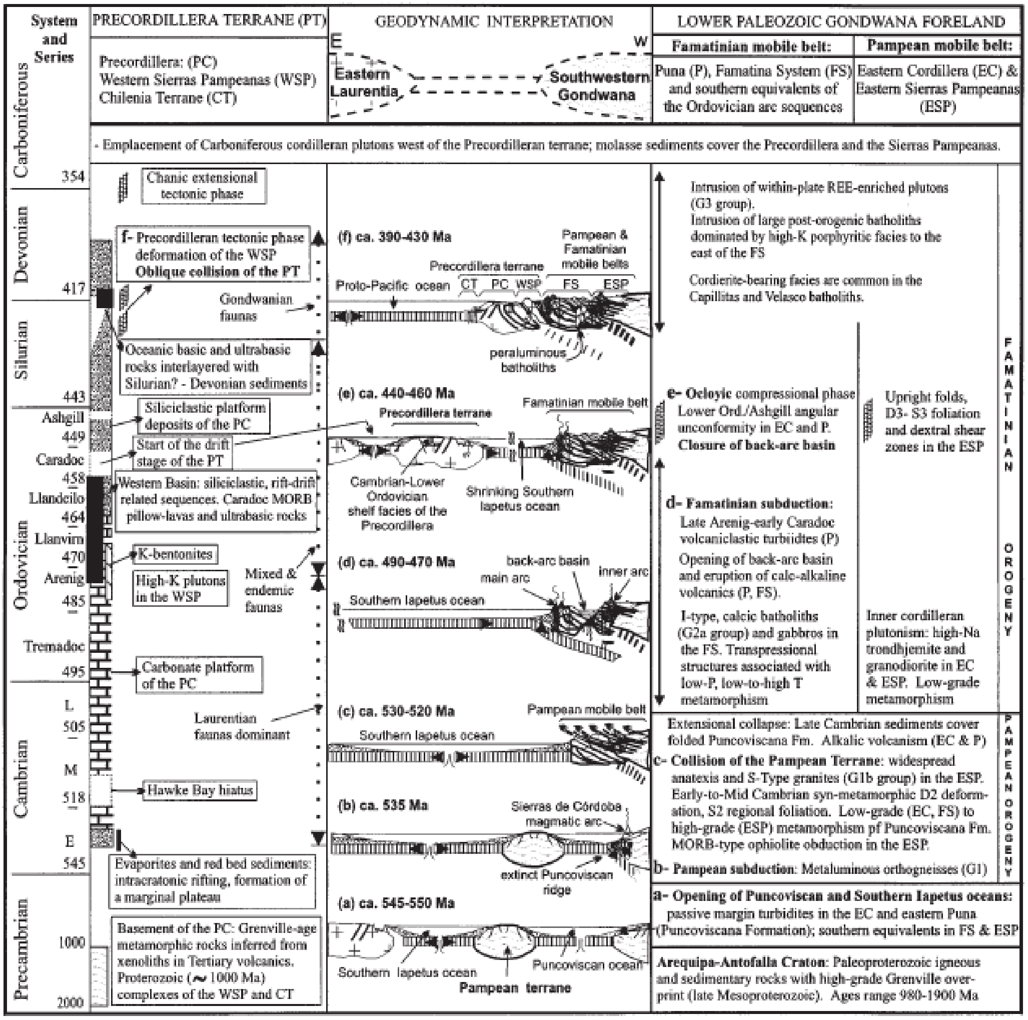


Figure 4-5 Sequence of Paleozoic orogenic events in proto-Andean margin of South America. Tectono-magmatic evolution from Pankhurst et al. (1998) and Rapela et al. (1998); time scale from Gradstein and Ogg (1996); Sedimentary sequences, fauna, and tectonic phases in the Precordillera from Astini et al. (1995), Benedetto, (1998), and Keller et al. (1998); plate tectonic phases or Precordillera terrane from Keller et al. (1998). (Presented by Rapela et al. 1998).

A sequence of Paleozoic orogenic events and tectonic settings in the proto-Andean margin of South America which is described in Figure 4-5 above, Escayola et al. (2007), proposed subduction towards the west which occurred in the Neoproterozoic with sediments being deposited in a back-arc basin from the Western Sierras Pampeanas and the arc itself. The subduction model suggests high-grade metamorphism of the Pampean belt followed Early Cambrian closure of the back-arc basin. Neoproterozoic subduction is demonstrated by U-Pb and Sm-Nd isotope data but ultimately does not support the existence of a back-arc basin (Escayola et al., 2007). Verdecchia et al. (2011) analyzed U-Pb SHRIMP detrital zircons from the Sierras Pampeanas and Sierras de Ambato and concluded the Rio de la Plata craton reached its current position in the Mid-to-Late Cambrian after the accretion Pampia terrane to the Gondwana margin.

The Sierras Australes, located in Argentina, contains a Late Permian fold and thrust belt that is Cambrian to Permian in age. The belt is continued in the Cape Fold Belt of South Africa and the Ellsworth Mountains of West Antarctica. The thrust belts are recognized as a single tectonic system that formed during the Late Paleozoic (Ramos, 1984, 1986). The Sierras Australes of Argentina is recognized as the accretion source for the Patagonian terrane. The origin of this terrane is subject to debate. The Patagonian terrane has been proposed to have both been an autochthonous part of Gondwana as well as an allochthonous origin. Ramos (1984; 1986) first proposed an allochthonous origin for the Patagonian terrane that collided with the South American segment of southern Gondwana along the Rio Colorado region during the Carboniferous. An allochthonous origin was suggested based on southwest-dipping subduction beneath the North Patagonian Massif. Charnicoff and Caminos (1996) document southward-verging folds, southward directed thrust of supracrustal rocks, and Devonian-Carboniferous penetrative deformation in the northeastern North Patagonian Massif. Von Gosen (2003) argued for Permian crustal shortening, and a northeastward-directed accretionary process. Ramos (2008) analyzes two magmatic belts in the region. Ramos (2008) looked at a western belt that was active from the Devonian to the Mid Carboniferous and a northern belt that is partially coeval that led to the collision of Patagonia against the southwestern margin of Gondwana in the Lower Permian concluding, that the basement of Patagonia is not exotic to Gondwana (Ramos, 2008). The idea of familiar Patagonian basement is demonstrated by the zircons observed in the magmatic and metamorphic rocks in the Somún Cura and the Deseado massifs. Brasiliano ages indicate the Patagonian block joined in the incorporation of Gondwana (Ramos, 2008). Ramos points out that occurrence of zircons supports a position taken by Rapalini (2005) that Patagonia is a para-autochthonous terrane. The two late Paleozoic magmatic and metamorphic belts imply that Patagonia was an independent plate since the Devonian. The western belt developed almost parallel to the continental margin but there is an arc trench-gap distance varying from 400-600 km from the present trench. The substantial trench-gap distance presents an issue that is subject to debate (Rapalini, 2005). Two models have been debated to account for this trench gap. The collisional model postulates a collision between southernmost Patagonia against the Deseado Massif. Dalla Salda et al. (1994) and Heredia et al. (2006) recognized a collisional deformation in the P-T paths of metamorphic rocks in the northern Patagonian Cordillera. Pankhurst et al. (2006) demonstrates a Mid-Carboniferous age for this metamorphism (discussed below). However, as Heredia et al. (2006), points out the Madre de Dios terrane is the only candidate for the slab that could have collided with the continental margin if the present extension of the magmatic belt is established. Hervé et al. (2003), points out that the carbonate rocks in this region are indicative of Late Carboniferous-Early Permian sedimentation and the youngest zircons of the metasedimentary sequence are Early

Permian. Early Permian sequences suggest that the Madre de Dios terrane cannot account for the Mid Carboniferous collision. Hervé et al. (2006), recaps U-Pb and Hf-Lu isotope data between southern Patagonia and the Antarctica peninsula that suggests these two land masses were in close proximity to each other. Ghidella et al. (2002, 2007) does a plate reconstruction of the Weddell Sea and puts forth this hypothesis as well. The Weddell Sea reconstruction model proposes that the southern Patagonian Cordillera was attached to the Antarctic Peninsula until the Early Jurassic and indicates a larger landmass could have impacted the margin during the Late Paleozoic (Ramos, 2008). The subduction model suggests that the Mid Carboniferous metamorphism points to shallowing of a subducted slab, followed by attenuation and Late Carboniferous-Early Permian granitic melts accompanied by injection of hot asthenosphere and melting of the crust during the steepening of the oceanic slab (Ramos, 2008). Both models, seen in Figure 4-6, could be combined depending on the location of the Antarctic Peninsula to explain the distance from the trench to the magmatic arc (Ghidella et al., 2007). Paleomagnetic pole data done in Patagonia has given different results. Paleomagnetic data presented by Rapalini (2005), demonstrate a conceivable para-autochthonous origin of Patagonia, involving rifting away from southwest Gondwana in the Late Proterozoic-Early Paleozoic the colliding again with southwest Gondwana in the Late Paleozoic. Tomezzoli and Vilas (1999), presented paleomagnetic data that demonstrated syntectonic sedimentation during the Early Permian in the Pillahuincó Group of Sierra de la Ventana, in the same sequence where López Gamundi et al. (1995) have established coeval deformation based on syngrowth strata (Ramos, 2008).

Pankhurst et al. (2006) revised the Patagonian collision model and claimed the majority of the rocks in the North Patagonian Massif are autochthonous to Gondwana. The paper shows the basement to the south of the Sierra de la Ventana includes Late

Neoproterozoic and Cambrian granites and volcanic rocks of similar age to those in the Pampean orogeny placing these rocks in a different tectonic setting (Pankhurst et al., 2006). The North Patagonian Massif has Ordovician granite magmatism and metamorphism equivalent to the Famatinian orogeny. Because there isn't any evidence of a Grenville-age belt, it was surmised that any collision must have occurred to the south of the massif with the deformed Cambro-Ordovician cover. Pankhurst et al. (2006), details subduction-related magmatism in the western magmatic belt during the mid-Carboniferous which led to the proposal of this area being the collision zone (Figure 4-7).

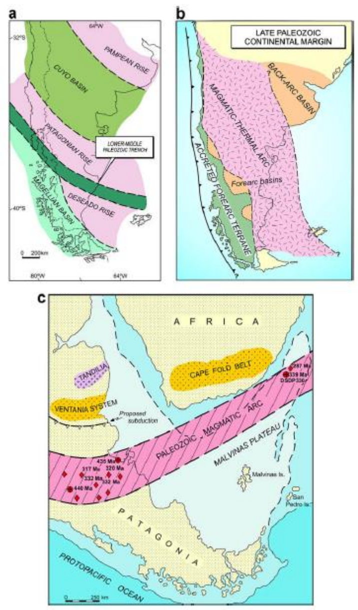


Figure 4-6 Models showing the allochthonous and autochthonous hypothesis. (a) Hypothesis showing Deseado Massif was allochthonous and early subduction zone split the Somún Cure and the Deseado Massif (Frutos and Tober (1975); Pankhurst et al. (2006)). (b) Autochthonous model where a wide magmatic arc crosses the entire Patagonia (Forsythe, (1982); Caminos and Llambías, (1984); Rapela et al. (1989); Dalla Salda et al. (1990)). (c) Allochthonous hypothesis (Ramos, 1984, 1986) (Presented by Ramos, (2008)).

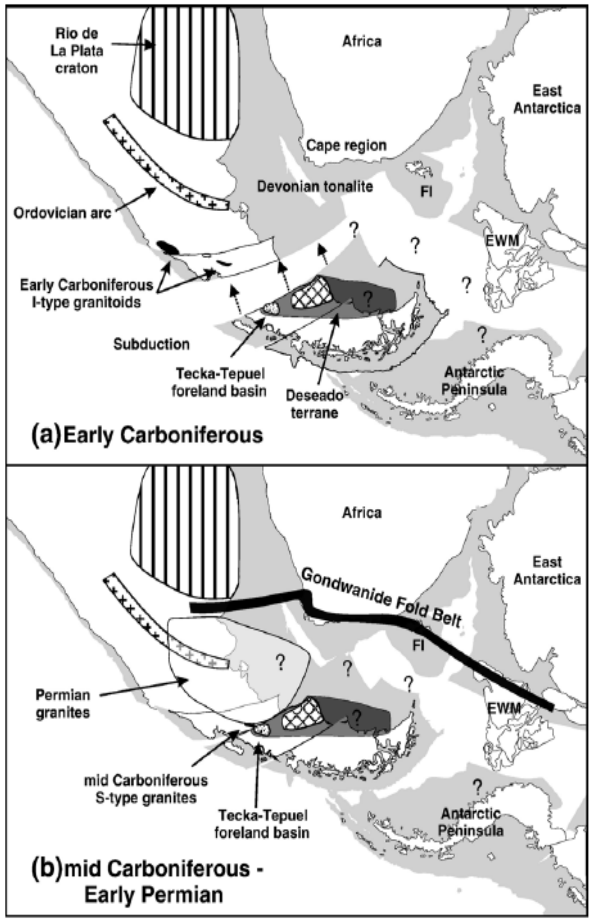


Figure 4-7 Late Paleozoic southwest Gondwana plate reconstructions. (a) Early Carboniferous subduction with the North Patagonian Massif forming part of the supercontinent since the Ordovician and is separated from a Deseado terrane to the south. Chile and the Antarctic Peninsula are shown in present day for easy identification. (b) Mid Carboniferous collision stage showing the degree of consequent deformation on the Gondwanide fold belts and Permian granitoid magmatism in the North Patagonian Massif. (FI = Falkland Islands, EWM = Ellsworth-Whitmore mountains crustal block. Pankhurst et al. (2006)).

The Madre de Dios terrane (Figure 4-8) was accreted to the western Gondwana margin in modern Patagonia during the Late Triassic to Early Jurassic Chonide orogeny (Hervé et al., 2003; Sepúlveda et al., 2008). Modern Patagonia is the potential location of a Late Carboniferous to Early Permian mid-ocean ridge (Sepúlveda et al., 2008). The Madre de Dios terrane is in an area of west Gondwana where there are accretionary complexes of Gondwana origin (e. g.,, Mortimer, 2004; Glen, 2005) along the margin that formed after Gondwana was assembled (Vaughan and Pankhurst, 2008).The Madre de Dios terrane contains Middle to Late Pennsylvanian and Early Permian Foraminifera that are found elsewhere in northern South America. Douglass and Nestell (1974) give a summary of fusulinid fauana in the Madre de Dios. The only record of Middle Pennsylvanian fusulinids in the terrane are *Eoschubertella* and *Fuslinella-Fusulina*. *Fusulinella* has been reported in Peru (e. g. Roberts, 1949) and Brazil (e. g. Petri, 1956). Eoschubertella has not been recognized with certainty elsewhere in South America (Douglass and Nestell, 1974). Each Late Pennsylvanian and Early Permian fusulinid genera recognized in southern Chile was recorded from the northern Andeas areas. There are differences at the species level but also considerable similarity ((Douglass and Nestell, 1974).

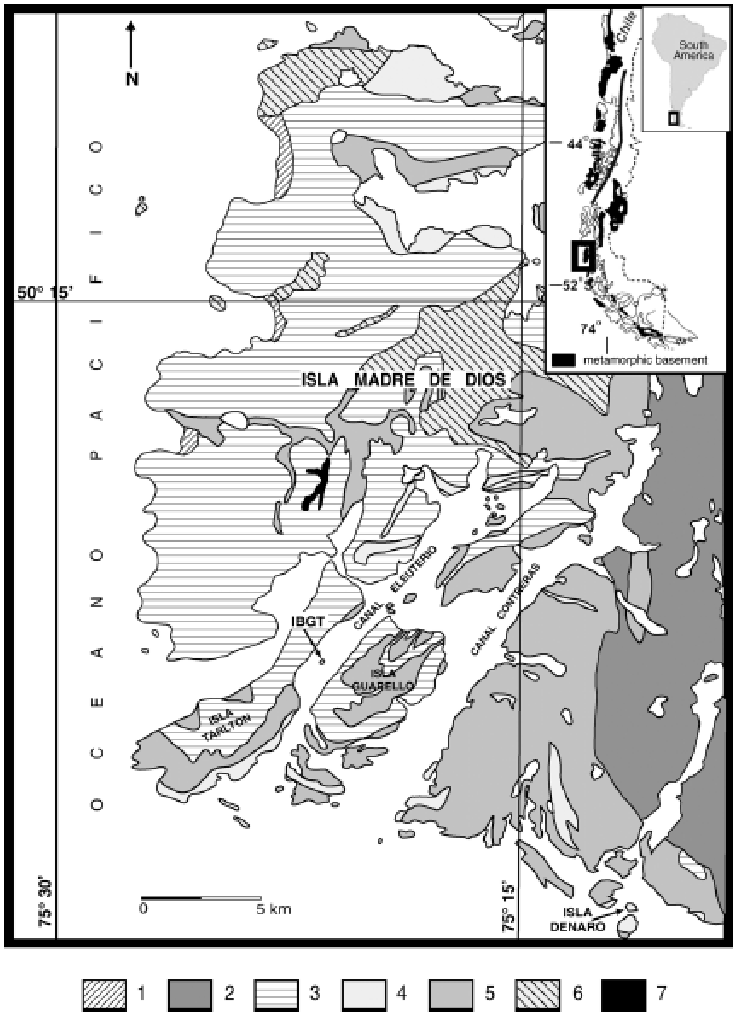


Figure 4-8 Geological map of the Madre De Dios terrane and surrounding area. (Map presented by Sepúlveda et al. (2008). Modified from Forsythe and Mpodozis (1983) and Lacassie (2003)). (1 = Quaternary deposits; 2 = South Patagonian Batholith; 3 = Tarlton Limestone; 4 = Denaro Complex; 5 = Duque de York Complex; 6 = Unmapped basement; 7 = Sill).

The Chilenia terrane (Figure 4-9), proposed by Ramos (1988) is a hypothetical terrane with little direct evidence to support its existence. The terrane is thought to have accreted to the Gondwana margin during the Devonian as a way to explain granite magmatism of Devonian age that occurs within the Pampean belt and to the south (Vaughan and Pankhurst, 2008). The Achala batholith in the southern Sierras de Cordoba consists of S-type granites consistent with post-orogenic features (Lira and Kirschbaum, 1990). Guena et al. (2007) use paleomagnetic data to support the idea of rapid cooling soon after crystallization. A Late Paleozoic arc and the Choiyoi Large Igneous Province have obscured possible direction evidence of the Chilenia terrane due to volcanics and intrusives in the region. Alvarez et al. (2011) use detrital zircon age populations from Late Paleozoic accretionary prims formed after its collision to the Gondwana margin. They postulate that the collisional event may have incorporated sediments derived from the erosion of the Chilenia basement (Alvarez et al., 2011). The paper focuses on zircons from three accretionary complexes (El Tránsito, Huasco, and Choapa) whose origins cannot easily be traced to well-known Gondwana sources and may have derived from the erosion of late Neoproterozoic to Early Cambrian magmatic/metamorphic foundations. The accretionary complexes potentially form a significant constituent of the Chilenia microcontinental basement (Alvarez et al., 2011).

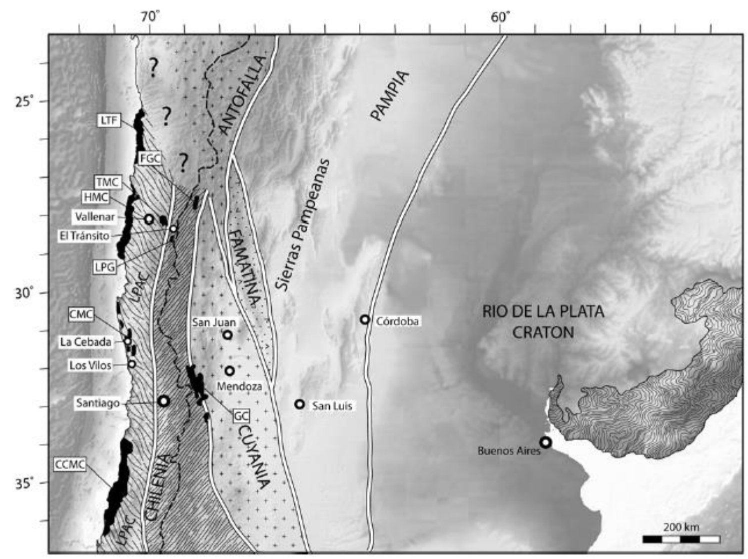


Figure 4-9 Map of tectonostratigraphic terranes of central South America including the boundary of the proposed Chilena terrane. Outcrops of the main metamorphic complexes are in black. (LPAC = Late Paleozoic Accretionary Complexes; LFT = Las Tortolas Formation; HMC = Huasco Metamoprhic Complex; LPG = La Pampa Gneisses; TMC = El Transito Metamorphic Complex; FGC = Filo Gris Complex; CMC = Choapa Metamorphic Complex; GC = Guarguaraz Complex; CCMC = Central Chile Metamorphic Complex. Figured presented by Alvarez et al. (2011) and modified from Ramos (2009)).

## South Africa

The current oceanic margin of southern South Africa contains sedimentary rocks from the Paleozoic Cape Supergroup, Natal Group and the Msikaba Formation (Johnston, 2000). The sedimentary sequences were deposited on a passive continental margin during multiple terrestrial and shallow marine-environments starting in the Early Ordovician and continuing until the Mid Carboniferous. There was a period of tectonism from about 278 Ma to 230 Ma which affected the Cape Supergroup rocks and caused in the Cape Fold Belt (Johnston, 2000). The formation of the Cape Fold Belt has multiple models. Johnston (2000) suggests the Cape Fold Belt developed inboard of the convergent margin of Gondwana and is ascribed to a left-step along a major intercontinental dextral shear zone. Dalziel et al. (2000) proposed that contact with a mantle plume was responsible for the flattening of the subduction zone and may have been responsible for continental break-up. A rifted continental margin is a potential setting for the Cape Supergroup, Natal Group and Msikaba deposits (Shone and Booth, 2005). The basement contains 2000-1000 Ma metamorphic volcano-sedimentary rocks of the Namaqua-Natal Belt (Dewey et al. 2006), and was deformed during the union of Gondwana (Jacobs et al., 2003). Pankhurst et al. (2006) supports a possible collisional model studying deposits gathered from the Sierra de la Ventana Fold Belt.

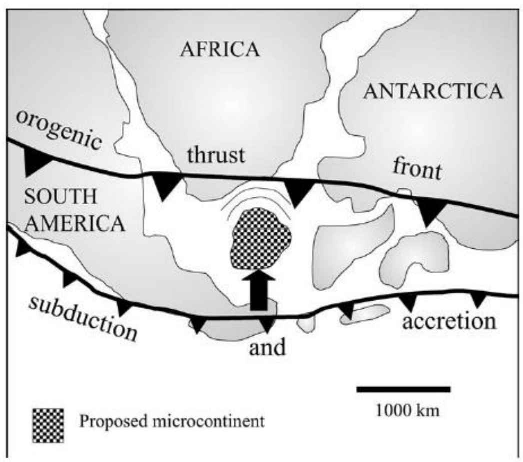


Figure 4-10 Plate tectonic model for the Cape Fold Belt. The proposed microplate, shown south of the South African coastline, could have collided with the African continent (Booth and Shone, (2002)). The hypothesized microplate could account for the greater degree of deformation in the eastern compared to the western portion of the fold belt. It also explains the lack of deformation in the Natal Group and the Msikaba Formation (Presented by Shone and Booth, (2005) and modified after de Wit and Ransome, (1992)).

Sedimentary rocks of the Cape Supergroup, Natal Group and Msikaba Formation contain an abundance of microfaunal fossil assemblages. The Graafwater and Sardinia Bay formations are separated by 700 km and occur at the western and easternmost ends of the fold belt. Both formations lie at the base of the Cape Supergroup. Successively overlying units of the Graafwater Formation contain an abundance of trace fossils including *Skolithos, Petalichnus and Arthrophycus* traces. No body fossils have been found in the Graafwater Formation (Shone and Booth, 2005). The Sardinia Bay Formation is notable for trace fossils (Shone, 1991). Genera identified within the formation include *Ophiomorpha*, *Thalassinoides*, *Diplocraterion*, *Skolithos*, (?)*Chondrites*, (?)*Planolites*, (?)*Fascifodina* and possibly *Cruziana*. Puzzling body fossils are possibly scraped fragments of stromatoporoid origin (Shone, 1983).

The Peninsula Formation appears in both the western and eastern parts of the Cape Fold Belt (Shone and Booth, 2005). Trace Fossils identified in the formation include *Diplichnites* (Rust, 1967; Anderson, 1975), *Cruziana* (Potgieter and Oelofsen, 1983) and various other arthropod traces attributed to *Rusophycus* and *Isopodichmus* (Broquet, 1990). *Planolites* and *Skolithos* trace fossils have also been identified (Broquet, 1992).

The Cedarberg Formation is a 150 m thick unit underlying the Pakhuis Formation. The Disa Member (mudrock, siltstone, and sandstone beds) contains a brachiopod assemblage which includes the genera (?)*Plectoglossa*, *Trematis*, *Orbiculoidea*, *Marklandella*, *Eostropheodonta*, and *Plectothyrella* (Cocks et al., 1970).

The Nardouw Formation sandstone is a 500 m thick succession. Trace fossils, especially *Skolithos*, are common (Rust, 1967). Fossil brachiopods occur at the top of the formation (Theron, 1970).

The Bokkeveld Group is possibly a 3000 m succession of mudrock and sandstone. The strata contain many trace fossils including the genera *Skolithos*, *Zoophycos* and (?)*Planolites* (Shone and Booth, 2005). Body fossils are also present. There is a wealth of crinoid, brachiopod, gastropod, bivalve and pteropod remnants. Reed (1925) and du Toit (1954) identified some cephalopod fragments, a few plant fragments, sponge and coral fragments.

The Witteberg Group is possibly a 2600 m succession of mudrock, siltstone and sandstone in the Eastern Cape. Trace fossils found in the Witteberg include *Skolithos* and *Zoophycos* (Shone and Booth, 2005). Body fossils found in the strata include the remains of lingulid brachiopods, palaeoniscid fish (Jubb, 1965) and lycopod and psilophyte plant fragments. Hill and Taylor (1992) describe fossilized fish and plant material in the eastern part of the fold belt.

The Msikaba Formation is exposed in the north of Port St. John's in the southeastern part of South Africa and overlies the Natal Group rocks along the coast at Wood Grange and Rock of Gibraltar near Hibberdene (Marshall and von Brunn, 2000). It is a 700 m thick succession of sandstones. The sheet sandstone facies contains the trace fossil *Scolicia* and the lenticular trough crossbedded sandstone facies contains *Planolites* traces (Hobday and Matthew, 1974). Lycopsid plant fragments were identified by Lock (1973).

## Antarctica

### *West Antarctica*

The Ellsworth-Whitmore Mountain block is the innermost block in West Antarctica. The mountains have a sedimentological similarity (Figure 4-11) to the Cape Fold Belt of South Africa (Curtis et al., 1999; Curtis, 2001; Vaughan and Pankhurst, 2008). The formation of the Ellsworth Mountains remains somewhat of a mystery as the strata do not present a clear presentation of early Paleozoic Ross-Delemarian orogenic deformation. The Transantarctic Mountains and eastern Australian margins show clear evidence of deformation making the formation of the Ellsworth Mountains enigmatic. The mountain strata show continuous sedimentation from the Early Cambrian to Permian (Dalziel, 2007). Randall and Mac Niocaill (2004) describe the Ellsworth block as preserving a passive margin volcano-sedimentary succession that ranges from the Cambrian to the Permo-Triassic and may have been derived from the Natal Embayment.

The remaining blocks of West Antarctica can be subdivided into at least three main terrane belts that appear to be continuous from the New Zealand region of East Gondwana to the Antarctic Peninsula (Pankhurst et al., 1998b; Vaughan and Storey, 2000; Vaughan and Pankhurst, 2008). The Ross province is the innermost and oldest terrane belt in West Antarctica and referred to as the Eastern Domain in the Antarctic Peninsula (Vaughan and Storey, 2000). The Eastern Domain is parauthochthonous terrane representing part of the margin of the Gondwanan continent (Vaughan and Storey, 2000). Is it likely an onshore exposed segment of East Antarctic continental crust beneath the Weddell Sea embayment, the Filchner Block. The Eastern Domain has similarities with the Western Province of New Zealand, the Ross Province of Marie Byrd Land, the Eastern Series of south-central Chile, the Pampa de Agnía and Tepuel rocks of north Patagonia, and the Cordillera Darwin rocks of Tierra del Fuego (Vaughan and Storey, 2000). The Antarctic Peninsula (Figure 4-12) contains granite, migmatite and paragneiss of Late Paleozoic to Mesozoic in age. Hf-isotope composition of inherited zircons studied by several authors (e. g.,, Flowerdew et al., 2006) show a Mesoproterozoic origin and suggests that this region is underlain by crust of this age. As Pankhurst et al. (1998) pointed out the oldest rocks of the Ross Province are from an Ordovician turbidite sequence in the Swanson Formation of Marie Byrd Land. The

Robertson Bay terrane of Victoria Land located in East Antarctica (Stump, 1995; Federico et al., 2006) contains turbidites of similar age; however, no equivalent turbidite sequences are known in West Antarctica (Vaughan and Pankhurst, 2008). A suite of granitoid deposited between 340 Ma – 320 Ma are apparent in Marie Byrd Land (Pankhurst et al., 1998) and are also observable at Target Hill in the northern part of the Antarctic Peninsula (Millar et al., 2002).

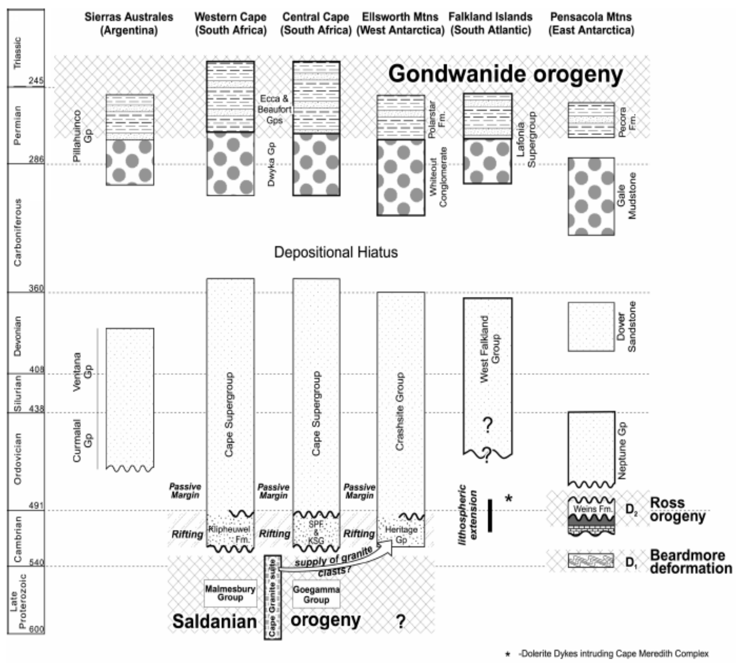


Figure 4-11 Tectonostratigraphic diagram for individual segments of the Gondwanian Fold Belt. The arrangement of the stratigraphic columns represent the continuous fold belt along the paleo-Pacific margin of Gondwana (Tectonostratigraphy of Pensacola Mountains is Storey et al. (1996); pre-Cape Supergroup is Barnett et al. (1997) and Armstrong et al. (1998); Presented by Curtis (2001). SPF = Schoemans Port Formation; KSG = Kansa Subgroup).

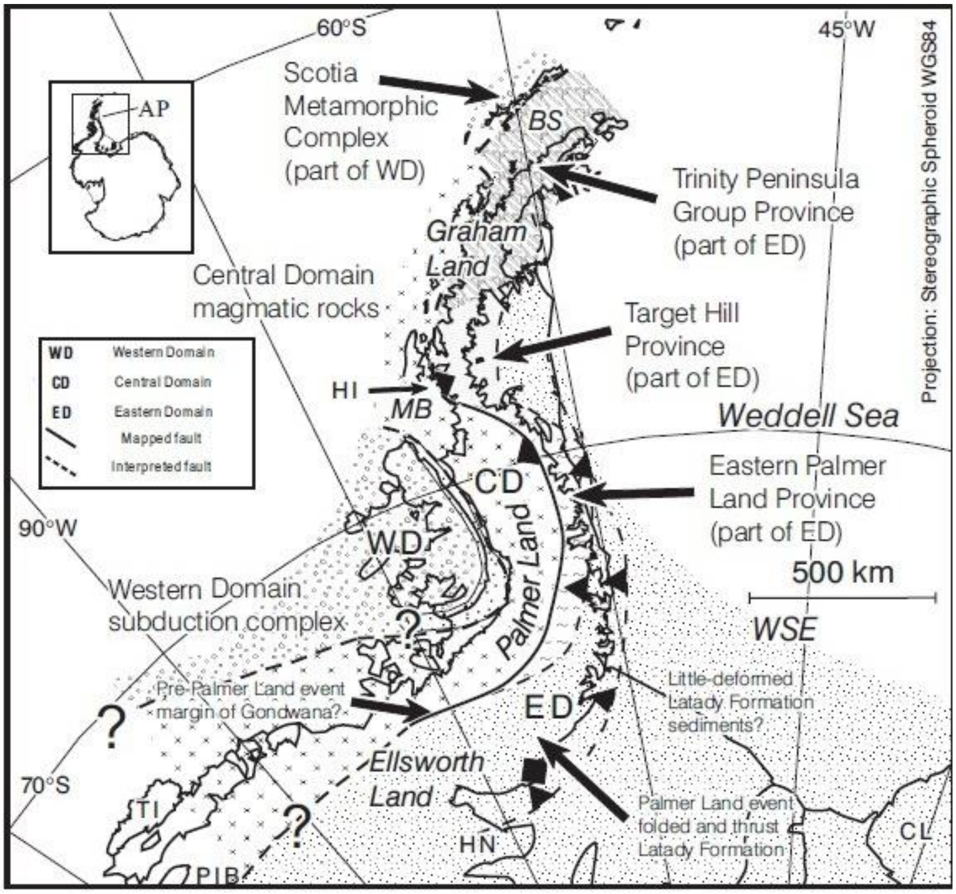


Figure 4-12 Map showing the provinces and domains of the Antarctic Peninsula (MB = Marguerite Bay; BS = Bransfield Strait; CD = Central Domain; CL = Coats Land; ED = Eastern Domain; HN = Haag Nunataks; PIB = Pine Island Bay Area; WD = Western Domain; WSE = Weddell Sea Embayment. Vaughan and Storey (2000).

Outside of the Ross Province are magmatic arcs terranes termed the Amundsen Province in Marie Byrd Land (Pankhurst et al., 1998b) and the Central Domain in the Antarctic Peninsula (Vaughan and Storey, 2000). Both terranes are largely magmatic and show many similarities in composition and timing of magmatic emplacement (Vaughan and Storey, 2000). Three distinct events of volcanism in the Late Triassic, Mid-Jurassic, and Late Jurassic to Early Cretaceous define the regions (Leat et al., 1995; Vaughan and Storey, 2000). Ferraccioli et al. (2006) used airborne geophysical data from the Antarctic Peninsula to suggest that the Central Domain is made up of smaller terranes. The data exposed subglacial imprints of crustal growth of the Antarctic Peninsula by Mesozoic arc magmatism and terrane accretion along the paleo-Pacific margin of Gondwana. The data also indicates that the Antarctic Peninsula batholith is an amalgamated magmatic arc terrane consisting of two divergent arcs. The eastern arc is mafic and the western arc is granitic (Ferraccioli et al., 2006). The amalgamation of these two arcs caused the Mid-Cretaceous Palmer Land orogenic event (Figure 4-13). Convergence and suturing may have been determined by two subduction zones or by a reduction in slab dip, leading to an inboard migration of the arc (Ferraccioli et al., 2006). The Central Domain strata demonstrate Late Triassic-early Jurassic and Mid-Cretaceous deformational events (Vaughan et al., 2002a; Vaughan et al., 2002b; Vaughan and Livermore, 2005). The Central Domain is suspect and may be an allochthonous, accreted microcontinental arc terrane (Vaughan and Storey, 2000). It shares similarities with the Median Tectonic Zone of New Zealand, the Amundesn Province of Marie Byrd Land, and the Coastal Cordillera of northern Chile (Vaughan and Storey, 2000).

The outermost terrane belt is the Western Domain (Vaughan and Storey, 2000). Similar accretionary complex terranes are developed in New Zealand, south-central Chile, and northern Patagonia although there is no equivalent in Marie Byrd Land. The domain represents either a subduction-accretion complex to the Central Domain or another separate crustal fragment (Vaughan and Storey, 2000). The Antarctic Peninsula is the location of a potentially allochthonous terrane-continent collision zone similar to models for New Zealand and South America (Vaughan and Storey, 2000). Kelly et al. (2001) describe an accretionary

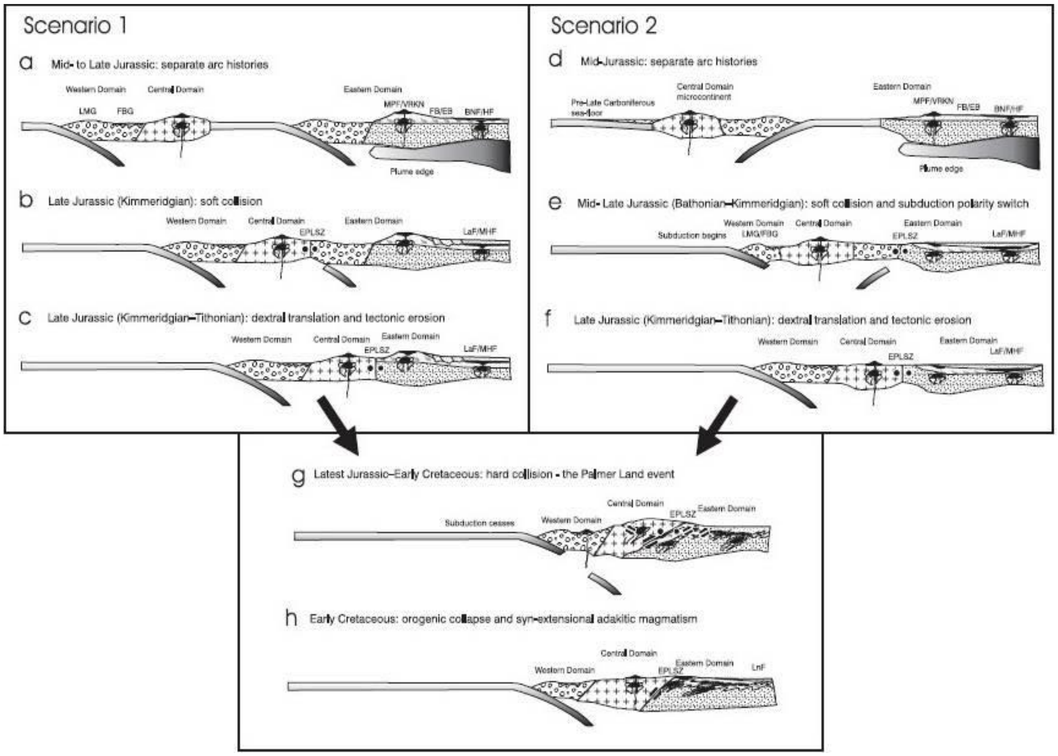


Figure 4-13 Tectonic reconstruction (Simandjuntak and Barber (1996)) the Palmer Land orogeny and subsequent events (BNF = Brennecke Nunataks Formation; EB = Erewhon Beds; EPLSZ = Eastern Palmer Land Shear Zone; FB = Fitzgerald Beds; FBG = Fossil Bluff Group; HF = Hjort Formation; LaF = Latady Formation; LMG = LeMay Group; LnF = Larsen Formation; MHF = Mount Hill Formation; MPF = Mount Poster Formation; VRKN = Volcanic rocks of Kamenev Nunataks. Presented by Vaughan and Storey (2000)).

complex related to an island arc in the Late Paleozoic hinting at possible terrane accretion around present day Alexander Island using Carboniferous and Permian marine microfauna from the Mount King strata. Vaughan and Storey (2000) provide a deformation history of the Western Domain looking at the LeMay Group. They describe accretion and deformation as being active from at the latest post-Early Jurassic to post-

Early Cretaceous. Ages are derived from the rocks in the accretionary complex. He also describes episodic deformation from the Late Carboniferous (Vaughan and Storey, 2000).

*Transantarctic Mountains and Victoria Land*

The northern segment of the Transantarctic Mountains, located in northern Victoria Land and extending to the Pensacola Mountains, is composed of Cambrian and Ordovician terranes that merged during the Ross Orogeny of the Early Paleozoic (Tessensohn and Henjes-Kunst, 2005). In Antarctica, rocks of the Ross Orogeny (Figure 4-14) border the East Antarctic craton and define the basement of the Cenozoic Transantarctic Mountains. Northern Victoria Land contains a wide exposure of Ross orogeny related rocks which led to tectonic models of the belt based on the notion of terrane accretion (Bradshaw, 1987, 1989; Kleinschmidt and Tessenholm, 1987; Stump, 1995; Tessensohn and Henjes-Kunst, 2005; Federico et al., 2006). Most models for Northern Victoria Land divide the region into three distinct terranes called the Wilson terrane, Bowers terrane, and Robertson Bay terrane. The Wilson terrane contains a low to high grade metasedimentary sequence that is intruded by calc-alkaline plutons with magmatic arc affinity (Federico et al., 2006). The Bowers terrane is comprised of low-grade metavolcanic and metasedimentary rocks that have been interpreted as an intra-oceanic arc. The Robertson Bay terrane is a low-grade flysch-like sequence. The proximity of each terrane in relation to one another is been ascribed to accretion during the Ross orogeny and not during the Devonian (e. g., Weaver et al. 1991; DiVenere et al. 1996) as previously thought (Federico et al., 2006). According to Federico et al. (2006) all three terranes were interpreted as an arc/back-arc/trench system, developed in a SW-dipping subduction zone setting. The subducting plate carried a continent that was originally positioned outward of the turbidite fan of the Robertson Bay terrane. A collision occurred between the aforementioned continent and the East Antarctic craton causing partial subduction of the intervening back-arc basin (Federico et al., 2006). The collision led to the end of the Ross-orogenic subduction. The collision forced the Robertson Bay terrane turbidite fan onto the continent and therefore, can partially account for the present day basement of the turbidite (Federico et al., 2006).

The Antarctic continent contains several Carboniferous and ?Permian marine faunas. The first Carboniferous and ?Permian marine fauna from Antarctica were found in Mount King in NE Alexander Island (Kelly et al., 2001). The fossils occur in the calcareous mudstone of the Mount King beds. The faunas are identified based on three locations within the Mount King beds and include bivalves (e. g., Anthraconeilo and Limipecten), brachiopods (e. g., Crurithyris) , bryozoans (e. g., Australofenestell cincta and Rectifenestella cf. R. loganesis), crinoids (e. g., Cyclocaudex and Pentaridica rothi), gastropods (e. g., Mourlonia and Ptychomphalina cf. P. kuttungensis), a possible monoplacophoran (Metoptoma?), nautiloids (e. g., Sueroceras) and a potential serpulid or microconchid. Possible Planolites and Zoophycos trace fossils are present as well (Kelly et al., 2001).

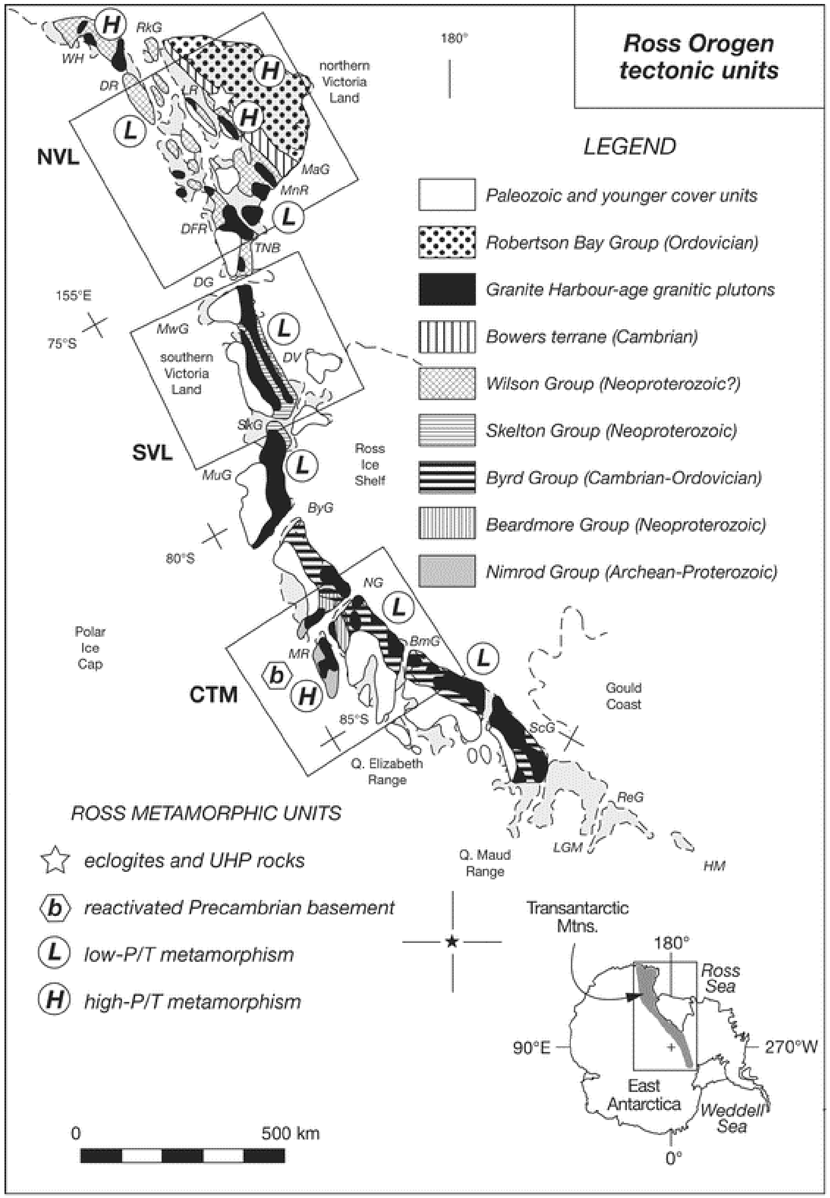


Figure 4-14 Tectonic map of Ross Orogen tectonic units (Collinson et al. (2006)).

The faunas of two Mount King localities are Carboniferous in age and are similar with the *Levipustula levis* Zone of Argentina and eastern Australia. Brachiopod evidence from another location indicates the presence of a linoproductid fauna of possible Carboniferous to Permian age. There are similarities with the Argentinian *Cancrinella* fauna (Kelly et al., 2001).

## New Zealand

New Zealand is on the eastern Gondwana margin and is made up of numerous terranes. New Zealand's patchwork of terranes is composed of basement rock that ranges in age from Early Cambrian to Early Cretaceous. The terranes can be subdivided into three provinces, the Western Province, the Median Province, and the Eastern Province (Coombs et al., 1976; Bishop et al., 1985; Bradshaw, 1989; Vaughan and Pankhurst, 2008). The following terranes make up New Zealand and are described from west to east.

The Buller terrane (Figure 4-15) is the westernmost recognized terrane in New Zealand. It is the closest terrane to the interior of Gondwana. The geology of the terrane consists of metamorphosed siliciclastic sandstone and mudstone. The sandstone originated on the continent and the fossils are dated to Ordovician in age (Cooper, 1989; Cooper and Tulloch, 1992; Roser and Korsch, 1988; Mortimer, 2004). The tectonic setting of the terrane is thought to be a passive or active continental margin. The Buller terrane is part of a Lower Paleozoic Gondwana Greywacke sedimentary suit which contains detrital zircons aged 500-600Ma, 1000-1200 Ma and 1500-1700Ma (Ireland, 1992; Ireland and Gibson, 1998; Mortimer, 2004).

The Takaka terrane (Figure 4-15) contains siliciclastic, carbonate and volcanic rocks dating between Cambrian to Early Devonian in age (Cooper, 1989; Cooper and

Tulloch, 1992; Bradshaw, 2000). The terrane contains Middle Cambrian trilobites which are New Zealand's oldest known fossils. Additional, Cambrian ultramafics and boninites are present (Münker, 2000), Ordovician limestone and Silurian orthoquartzite (Mortimer, 2004). The tectonic setting is interpreted as a primitive intra-oceanic island arc in the Cambrian, succeeded by continental passive margin sedimentation in the Ordovician to Devonian. The contact between the Takaka terrane and the Brook Street terrane marks the boundary of the Western and Eastern Province. The margin has been intruded by plutons of the Median Batholith and is no longer recognizable as a fault (Mortimer, 2004).

The Brooks Street terrane (Figure 4-15) geology is mainly Permian, and is a subduction-related volcanic pile and volcaniclastic apron that contains large amounts of pyroxene and basalt (Houghton and Landis, 1989; Landis et al., 1999). The tectonic setting is thought to be a primitive, intraoceanic island arc assemblage. The south coast contains Permian aged gabbro and trondhjemite plutons which are petrogenetically related to Brook Street terrane volcanic rocks and are thought to be an allochthonous part of the Median Batholith (Mortimer, 2004). Haston et al. (1989) use paleomagnetic data to suggest a possible low to intermediate paleolatitude in the Permian for the terrane (Mortimer, 2004). The Takitimu and Productus Creek Groups contain brachiopods and mollusks of Austrazean and East Australian affinity (Campbell, 2000). The Letham Ridge thrust notates the contact between the Brook Street and Murihiku terranes (Landis et al., 1999).

The Murihiku terrane (Figure 4-15) contains the Murihiku Supergroup, a Late Permian to Late Jurassic volcaniclastic marine sandstone (Mortimer, 2004). The Supergroup also contains conglomerate, mudstone and numerous tuff (Balance and Campbell, 1993; Campbell et al., 2001). There are isolated areas of shallow intrusive and/or volcanic rocks of Late Triassic to Early Jurassic known as the Park Volcanics

(Mortimer, 2004). Park Volcanics are exposed in the syncline's south limb in the southern South Island. Paleomagnetic data indicate high paleolatitudes (Grindley et al., 1981). The depositional environment is thought to be a long survived forearc or backarc basin. Fossil data shows Nothal, Tethyan, cosmopolitan (Campbell and Grant-Mackie, 2000) faunas as well as coal seams and plant fossils (Pole, 1999, Campbell et al., 2001). The Hillfoot fault (Bishop and Turnbull, 1996) is the contact between the Murihiku and Maitai terranes.

The Maitai terrane (Figure 4-15) contains an Early Permian ophiolite belt which is unconformably overlain by a Late Permian to Middle Triassic volcaniclastic sedimentary sequence, called the Maitai Group (Mortimer, 2004). Current models on the origin of the ophiolite belt suggest a near-arc setting. Coombs et al. (1976); Kimbrough et al. (1992); Malpas et al. (1994); Sivell and McCulloch (2000) have conducted studies and petrologic studies on the ophiolite belt. The Livingstone Fault is the contact between the Maitai and Caples terrane (Cawood, 1986).

The Caples terrane (Figure 4-15) is tectonically imbricated and weakly metamorphosed. It contains fossiliferous marine volcaniclastic Permian-Triassic greywacke and argillite (Mortimer, 2004). The greywacke and argillite have differing sandstone and petrofacies from the Rakaia and adjacent terranes (MacKinnon, 1983; Roser et al., 1993). There is lava, chert, and limestone present in the terrane (Mortimer, 2004). Deposition transpired as submarine fan deposits in lower trench-slope basins and on a trench floor adjacent to an island arc. The contact with the Rakaia terrane has been overprinted by the Otago Schist (Mortimer, 2004).

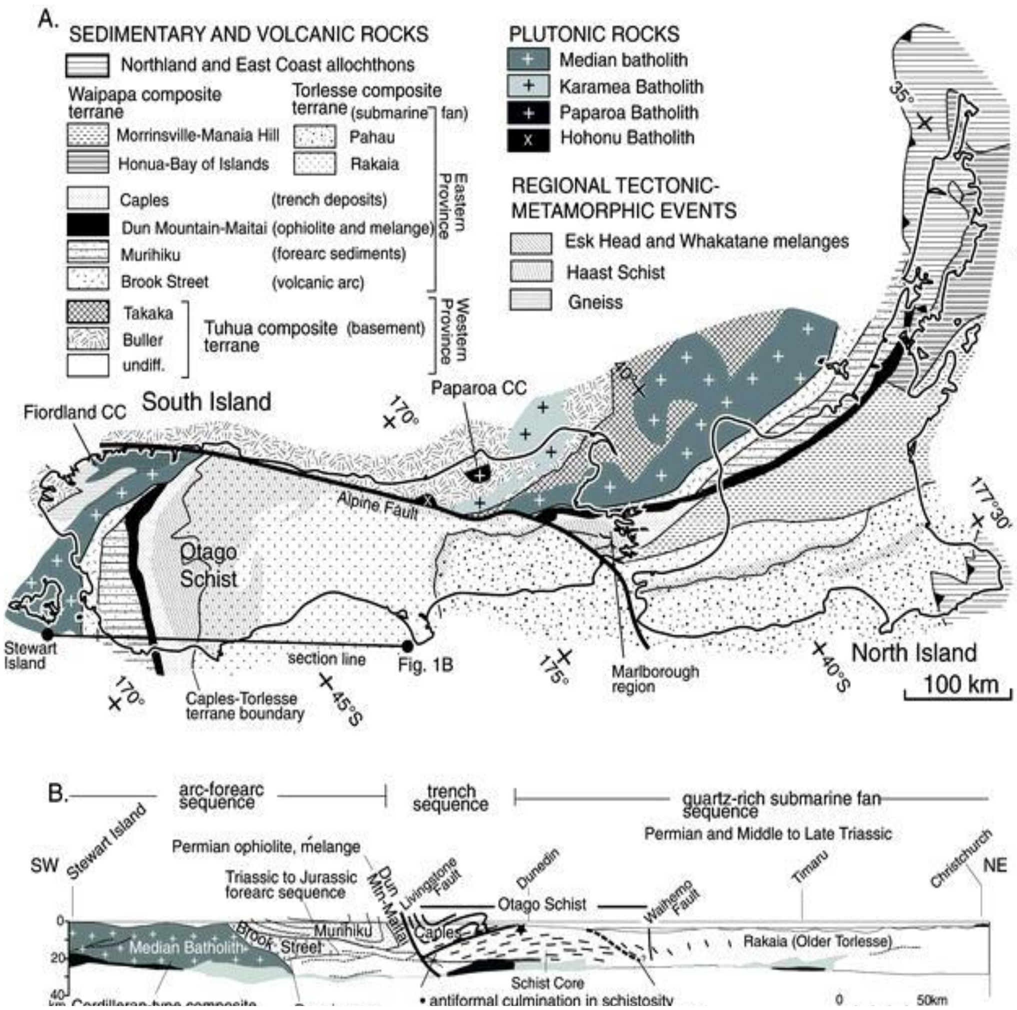


Figure 4-15 Basement geology and crustal cross section of the New Zealand (Presented in GSA memoirs 2010 (2005)).

The Rakaia terrane (Figure 4-15) consists mainly of a Permian to Late Triassic quartzofeldspathic sandstone-mudstone submarine turbidite. The substrate is oceanic crust of Carboniferous and Permian age. Clastic rocks were deposited over the ocean crust (Mortimer, 2004). The sandstone of the terrane is quartz-rich, plutoniclastic and of average rhyodacitic compistion. Sedimentological differences of the above mentioned sandstone beds, when compared to other terranes, prompted a petrographic and geochemical analysis done by MacKinnon (1983) and Roser and Korsch (1999). Results suggested an origin from an active continental volcanoplutonic arc. A Gondwana source for the Rakaia terrane, likely in northern Australia, is hypothesized by the ages of detrital zircons and mica (Ireland, 1992; Adams and Kelley, 1998; Pickard et al., 2000). The age analysis of the detrital zircons and mica has indicated similar ages to Buller terrane zircons with the Rakaia zircons predominantly dating Permian to Triassic. Campbell and Grant-Mackie (2000) show Austrazean paleobiogeographic flora and fauna represented in the Rakaia terrane. Tectonic features indicate an east-facing accretionary wedge. It remains uncertain whether the past depositional environmental was an active or passive margin (Beggs, 1993). The Esk Head Mélange is recognized as the boundary between the Rakaia terrane and Pahau terrane (Mortimer, 2004).

The Bay of Islands terrane (Figure 4-15) is a proposed terrane in the Hunua facies portion of the Waipapa terrane (Spörli, 1978; Kear and Mortimer, 2003). Permian, Triassic and Early Jurassic basalt, chert, and limestone are inserted into Triassic to Late Jurassic trench and trench-slope sandstone and mudstone (Black, 1994). Sandstone petrofacies bear a resemblance to those in the Rakaia terrane with certain mafic volcaniclastics extant as well (Mortimer, 2004).

The Morrinsville-Manaia Hill terrane (Figure 4-15) consists of two distinct facies. The first is the younger Morrinsville facies which is compositionally similar to the coeval

Torlesse terrane in the eastern North Island. The second is the older Hunua facies which includes a mélange zone containing slices of Jurassic ocean floor (Vaughan et al., 2005).

The Pahau terrane (Figure 4-15), also known as the Younger Torlesse Terrane, is similar in lithology and structure to the Rakaia terrane. The terrane contains Late Jurassic and Early Cretaceous sandstone-mudstone and Triassic to Early Cretaceous limestone and chert (Mortimer, 2004). Cawood et al. (1999) and Kamp (2000) show some detrital zircons as young as 100 Ma. Tuffs are common in the Pahau terrane. Much of the clastic detritus on the terrane is likely from recycled Rakaia rocks (Mackinnon, 1983). There is a volcanic component to the detritus as well (Pickard et al., 2000). The Pahau terrane characterizes trench deposits that have a similar age and origin to the metamorphism in and unearthed of the Rakaia terrane (Mortimer, 2004).

There are other terranes (Figure 4-15) recognized in New Zealand that go by different names. Mortimer (2004) uses the name Tuhua terrane to signify the Takaka and Buller terranes as one large terrane.  The Median Tectonic Zone (Figure 4-16) is a region of terrane remains and igneous complexes of ambiguous status and association. It has also been proposed as a superterrane that lies between the Brook Street and Takaka terranes (Mortimer, 2004). There are Northland and East Coast allochthons in the North Island that have not been given terrane identification. The allochthons contain Early Cretaceous to Paloegene ophiolitic, volcanic and sedimentary rocks that were thrust over the North Island from the northeast at the end of the Oligocene (Balance, 1993; Malpas et al., 1994).

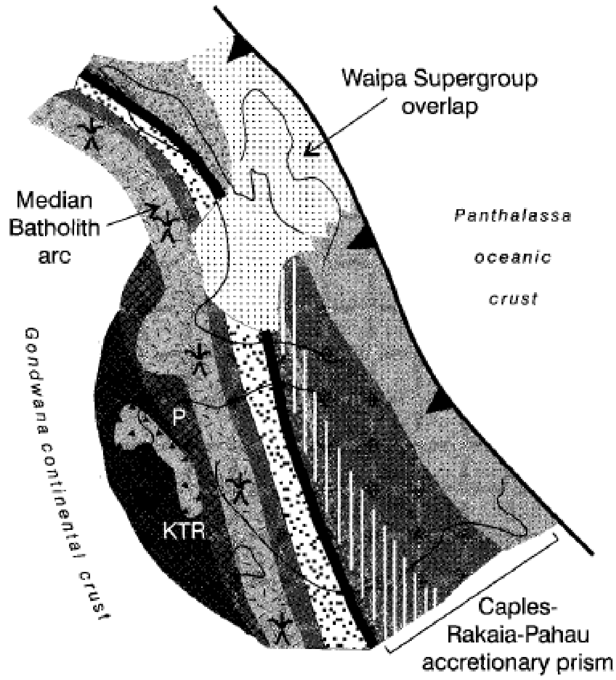


Figure 4-16 Potential paleogeographic interpretation of the terranes and batholiths at 120 Ma that emphasizes a single unified accretionary wedge and batholithic arc system (Mortimer (2004)).

Campbell and Warren (1965) and Speden (1976) created the original fossil database for the Torlesse (Pahau and Rakaia), Caples and Waipapa (Morrinsville-Manaia Hill and Honua-Bay of Islands) terranes. The catalog was based on macrofauna fossil assemblages (Adams et al., 1998). Progress was made in making a microfauna database for this region including contributions from Aita and Spörli (1992) for radiolarians; Leven and Grant-Mackie (1997) and Leven and Campbell (1998) for fusuline foraminifers; calcispheres from Campbell and Handler (1996); conodonts from Ford (1995); and dinoflagellates from Wilson et al. (1988). Each terrane contains fossiliferous lithologies comprising of limestone, phosphoritic and calcareous nodules, hemipelagite and chert. Sampling of these sequences led to a more precise measurement in the stratigraphic age of each unit within the terranes (Adams et al., 1998).

The microfauna data supports a typical accretionary margin subduction complex for the above mentioned terranes. Specifically, a predominantly clastic wedge with a fauna that is characterized by features of a Gondwanaland margin sequence that can trace its origins back to the Austrazean Province and a minor but apparent biogenic sedimentary piece that relates to off-scraped oceanic plate sequences (Adams et al., 1998). Guyots, isolated underwater volcanic mountains, are present and were introduced tectonically by sea floor spreading (Adams, 1998). Oceanic sequences typically, but not always, reflect warmer oceanic and climatic conditions of Tethyan origin and older ages than the clastic sequences (Hornibrook and Shu, 1965; Hada and Landis, 1995). This guideline is consistent with the subduction-accretion model with depletion of old sea floor central to a riverine or riparian continental mass with high relief (Adams et al, 1998). Campbell et al. (1993) study of fossil assemblages within the clastic sequences in the Torlesse terranes indicates a series of major sedimentation episodes during the Late Permian, Middle to Late Triassic and Late Jurassic to Early Cretaceous. Fossil biotas of clastic sequences within the Torlesse, Caples and Waipapa terranes are compatible with original depositional sites central to the eastern Australian margin of Gondwanaland (Adams et al., 1998).

There are several suggested smaller terranes but the geologic variances amongst them and currently recognized terranes are questionable. The Drumduan terrane (Johnston et al, 1987) and the Largs terrane (Williams and Smith, 1979) are similar to the volcanics of the Median Tectonic Zone; the Te Akatarewa terrane (Cawood et al., 2002) and Kakahu terranes (Bishop et al., 1985) can be regarded as Permian and Carboniferous portions of the Rakaia terrane; the Chrystalls Beach Complex (Coombs et al., 2000) is potentially part of the Caples terrane and Willsher Group (Campbell et al., 2003) is thought of as part of the Maitai terrane.

Wandres and Bradshaw (2005) call attention to the fact that the majority of New Zealand's continental crust is submerged by the sea. Rb-Sr metamorphic and U-Pb detrital zircon ages from island locales of the submerged continental crust reveal that the Campbell Plateau region of Zealandia has a clear parallel with the Western Province/Ross Province and the Median Province/Amundsen Province (Adams, 2008).

## Australia

The Tasman Fold Belt represents the boundary between cratonic Australia to the west and the collage of terranes to the east. During the Late Neoproterozoic and Early Cambrian, the Tasman Fold Belt collected quartzose turbidites in a passive margin setting. The Tasman Fold Belt changed in the late Early Cambrian to convergence related to the Cambro-Ordovician Ross-Delamarian orogeny (McElhinney et al., 2003). Terrane accretion in eastern Australia started in the Early Paleozoic along the Tasman line. The boundary of the Australian craton extended eastward after each successive terrane accretion event (Powell et al., 1990). Various exotic terranes could have accreted between 520 Ma and 490 Ma after the western portion of the Tasman Fold Belt became a marginal sea behind an oceanic island arc. Fragments are preserved in the Molong-Monaro terrane and in the Lolworth-Ravenswood terrane (McElhinney et al., 2003).

The Tasman Fold Belt (Figure 4-17) can be divided into five major superterranes: the Adelaide and Kanmantoo Fold Belt (Glenelg and Western Tasmania terranes), Lachlan Fold Belt (Stawell, Howqua, Melbourne-Mathinna, Wagga-Omeo, Girilambone, Molong-Monaro, and Narooma terranes) Thomson Fold Belt (Lolworth-Ravenswood terrane), Hodgkinson-Broken River Fold Belt and New England Fold Belt. The origin of the Kanmantoo superterrane was link to the Australian craton in the Early Cambrian and accreted to Australia by the Late Cambrian (Powell et al., 1990; Li et al., 1997). The Lachlan superterrane contains Ordovician quartzose flysch. The Ordovician quartzose turbidite sequence is an overlap accumulation covering incomplete Cambrian outcrops, which belong to several probable terranes (McElhinney et al., 2003). The Ordovician sequence indicates that the Stawell, Howqua and Melbourne-Mathinna terranes must have lain near the Australian continental margin since the Cambrian. The Ordovician overlap sequence contains terrestrial to shallow-marine conglomerate, sandstone and limestone that define an east-facing shoreline extending from Tasmania to far western New South Wales (Webby, 1978; Powell, 1984). Post-Cambrian terranes are unlikely (Chapell et al., 1988). The Wagga-Omeo, Girilambone and Molong-Monaro terranes origin may coincide with the western Lachlan Fold Belt. During the Middle Devonian there was widespread Tabberabberan deformation that affected all the terranes (McElhinney et al., 2003).

The eastern half of the Lachlan and Thomson Fold Belts experienced deformation in the Early Silurian to Middle Devonian. During this period of deformation the western portion of this region had continuous passive margin sedimentation (Powell et al., 1990). Smaller terranes were shuffled around along the eastern paleo-Pacific margin of the Tasman Fold Belt including the above mentioned Molong-Monaro terrane. The Molong-Monaro terrane might have been displaced along the form passive Gondwana margin in the Mid Silurian to Mid-Devonian (Powell, 1983). The Molong-Monaro terrane is not exotic to Gondwana because it contains the same Ordovician turbidites found in the Lachlan and Thomson Fold Belts (McElhinney et al., 2003). The extent of the displacement of the Molong-Monaro terrane is unknown. All the reliable

Early Paleozoic paleomagnetic poles from the Lachlan Fold Belt, with the exception of the Snowy River Volcanics pole, are in the Lambie Facies overlap assemblage (Li et al., 1991). The Lambie Facies overlap assemblage is Late Silurian to Early Devonian in western Victoria and western New South Wales and becomes gradually younger eastwards (Powell et al., 1990). The Molong-Monaro terrane was accreted to the Gondwanan margin of Australia by the Mid-Devonian. Accretion was during the Tabberabberan orogeny as it deformed the eastern and western parts of the Tasman Fold Belt (Powell et al., 1990). Powell et al. (1990) suggests these terranes can be considered to once have been a single terrane since the Early Devonian. The union of eastern Tasmania with western Tasmania occurred during the Mid-Devonian Tabberabberan deformation (Powell and Baillie, 1992).

The Lolworth-Ravenswood terrane is to the north of the Thomson Fold Belt. The terrane features an ENE structural trend of the Seventy Mile Range Group (McElhinney et al., 2003). ENE trending is a remarkable feature as the remaining Tasman Orogen trends NNW (Henderson, 1986). Henderson (1986) interpreted this terrane, while noting the ENE structural trends across the entire terrane, as part of a continental margin calc-alkaline succession that extended from North Queensland all the way to Tasmania. The Lolsworth-Ravenswood terrane contains three batholiths (Reedy Springs, Lolworth, and Ravenswood) that contain granitoids of Silurian-Devonian age that are similar to those found in the Georgetown Inlier (Bain and Draper, 1997). There are inherited zircons (approximately 2000 and 1500-1550 Ma) in the Reedy Spring and Lolworth batholiths that are similar in age to zircons in Georgetown Inlier however, there are younger zircons (Approximately 1100 Ma) found in the Lolworth Batholith as well. Isotopic data also suggest the source material from the Ravenswood Batholith also dates to approximately 1100 Ma. (Hutton et al., 1996). Hutton et al. (1996) suggest that the Lolworth-

Ravenswood terrane is underlain by Proterozoic continental crust that varies in age depending on location, but with links to the main craton.

The Georgetown Inlier contains metavolcanic and interlayered metasediment similar to the Seventy Mile Range Group of the Lolworth-Ravenswood terrane. Henderson (1986) uses this data to interpret that the terrane must have been adjacent to the main craton by the Mid-Silurian. It is possible however, that the Lolsworth-Ravenswood terrane was adjacent to the Georgetown Inlier during the Late Silurian to Early Devonian as sediment deposition was taking place in the Camel Creek terrane. The Camel Creek terrane contains bounding structures that would indicate the Lolsworth-Ravenswood terrane underwent a 90º rotation during this time (Henderson, 1987). The age range for the date of rotation would have occurred between 435 - 380 Ma. The older age limit is restricted by the age of the plutons in the Lolsworth-Ravenswood terrane and the Georgetown Inlier and the younger age is limited by the oldest sediments in the Late Devonian-Early Carboniferous Drummond Basin (McElhinney et al., 2003).

By the Late Devonian, an Andean-style magmatic arc had formed along the paleo-Pacific margin of the Thomson and Lachlan Fold Belts presenting the first sign of an assembly of the New England and Lachlan-Thomson Fold Belts (Li and Powell, 2001). The Yarrol terrane is thought to have accreted to the Thomson Fold Belt by the Mid-Devonian (Murray, 1986; Scheibner and Basden 1997) The Tamworth terrane may have been accreted to the Lachlan Fold Belt as early as the Late Devonian. The Late Devonian timeline is based on evidence gathered by Flood and Aitchison (1992) and Aitchison and Flood (1995) of quartzite clasts of hypothetical Lachlan origin. Whereas there is some sedimentary strata linking amalgamation of the New England and Lachlan Fold Belts in the Early Carboniferous, the abundance of data that exists currently point to an unmistakable unification in the Late Carboniferous (McElhinney et al., 2003). The

Westphalian Rocky Creek Conglomerate contains numerous clasts of Lachlan Fold Belt plutonic rocks which solidifies the connection. Strike-slip movements between the Lachlan Fold Belt and the Tamworth terrane (Harrington and Korsch, 1985 a, b; Veevers et al., 1994) could have continued into the Mid-Triassic. (McElhinney et al., 2003).

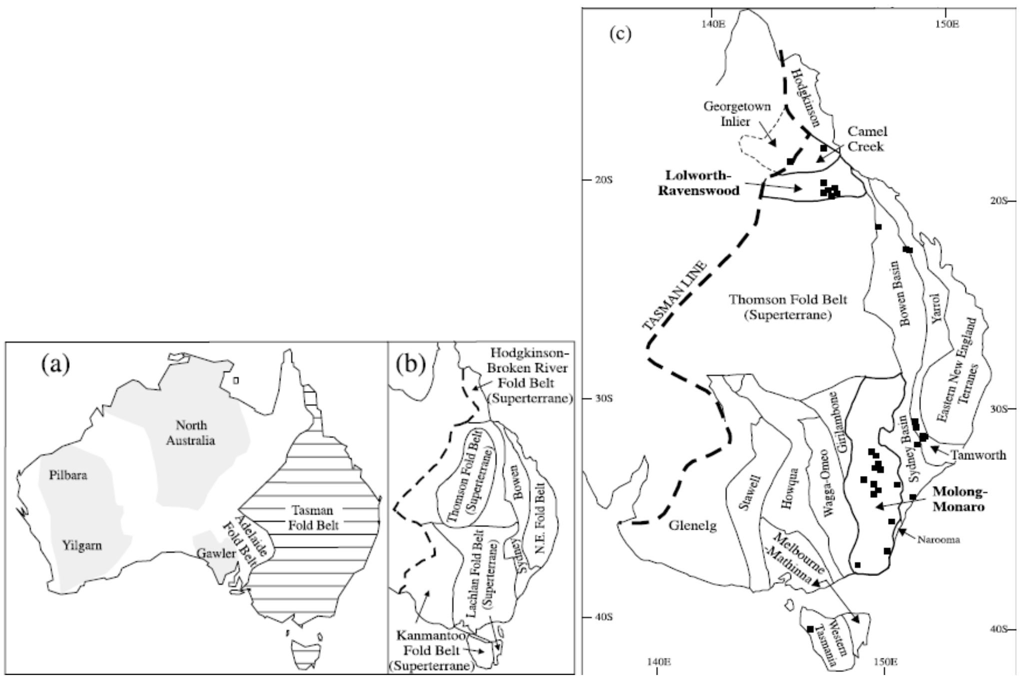


Figure 4-17 Geology and terrane map of Australia. (a) Precambrian cratons and the Tasman Fold Belt. (b) Tasman Fold Belt superterranes. (c) Subdivided terranes of the superterranes. Paleomagnetic data was taken at the sites marked with black squares (McElhinney et al. (2003)).

Australia contains a distribution of Australiasian archeocyathids which reflect Australia's position adjacent to and conjoined with Antarctica during the Cambrian. Carbonate platforms extended from central Australia through South Australia and along the Ross Fold Belt of Antarctica in the Mid-Late Cambrian (Brock et al., 2000).

Archaeocyathids that flourished during this interval are now found in the Amadeus and Georgina basins, Arrowi, Stansbury and Officer basins, Gnalta Shelf, Ross Fold Belt and outliers in Antarctica (Brock et al., 2000). 26 species of archaeocyathids are common to both Australia and Antarctica (Debrenne et al., 1990) lending support to the idea of a amalgamated Australia-Antarctica province. Nine of the ten valid named species recorded from allochthonous limestone clasts in the Dwyka Subgroup of South Africa are known also from Antarctica and/or Australia (Debrenne 1975; Debrenne and Kruse, 1989).

Early Cambrian Australian lingulid brachiopods share three genera in common with West Africa and West Antarctica (Brock et al., 2000). Middle Cambrian faunal assemblages share strong affinities with ten genera in New Zealand, three from East Antarctica, and two from West Antarctica. South America shows the least commonality among affinities with only a single genus (Brock et al., 2000). Late Cambrian faunas share seven genera in common with West Antarctica (Brock et al., 2000).

The Late Cambrian calcite brachiopod species *Roanella* in Australia (Brock and Talent, 1999) is similar to a form described as *Billingsella* cf. *B*. *borukaevi* (Henderson et al., 1992) from the Early-Late Cambrian in the Ellsworth Mountains (Brock et al., 2000).

A Middle Cambrian molluscan fauna from allochthonous limestone blocks of the Tasman Formation (MacKinnon, 1985) bears resemblance with a fauna from the Murrawong Creek Formation in the New England Fold Belt. The Late Cambrian species *Ribeiria* has been documented in West Antarctica (Webers et al., 1992).

# Chapter 5

## Discussion and Conclusions Based on the Evidence

### South America

Work on the meta-sedimentary rocks of the southern Pacific Andean margin has shown that there is no correlation to the Paleozoic Gondwana basement. Instead, the rocks are interpreted as Mesozoic accreted material (Hervé, 1988; Fang et al., 1998; Hervé and Fanning, 2003; Vaughan et al., 2005). The Argentine Precordillera is regarded as a large exotic terrane derived from Laurentia and accreted during the Early Paleozoic (e. g., Ramos et al., 1986; Moore, 1994; Astini et al., 1995; Thomas andand Astini, 2003). The Precordillera terrane (also known as the Cuyania terrane) has been the focus of numerous tectonic models including those that have doubted its Laurentian origin. Some models conclude an autochthonous origin with Gondwana (Aceñolaza et al., 2002). Geochronology and detrital zircon evidence have been salient to both sides of the debate (Casquet et al., 2001; Thomas et al., 2004; Finney et al., 2005). Western South America retains a fragmentary record of high grade Proterozoic metamorphic rocks with the same origin as those found in the Grenville belt of North America (Thomas et al., 2004). Proterozoic Grenvillian fragments in the Columbian Andes and Grenville-age massif anorthosite in western Argentina (comparable to rocks of the Grenville province) may represent a common orogeny that links Laurentia and Western Gondwana with Rodinia (Cordani et al., 2005; Casquet et al., 2005; Vaughan et al., 2005).

The Famatina Complex potentially represents an autochthonous arc-continent collision on the Gondwana margin in the Late Proterozoic to Ordovician. The evidence from the Famatina Complex could be related to the accretion of the Precordillera terrane (Miller and Söllner, 2005). Late Proterozoic to Cambrian sediments of the Puncoviscana basin shows a peripheral foreland basin succession to the Pampean orogeny

(Zimmerman, 2005). Lucassen and Franz (2005) show an alternative history for the early Paleozoic of the Central Andes. The authors propose a non-terrane, tectonic circumstance similar to the present day.

Rapela et al. (2005) identify an Early Jurassic magmatic arc and demonstrate that magmatism in the Triassic-Jurassic break expose a rotational tectonic system which would be a major restriction on the plate structure of Patagonia and the association of South America and the Antarctic Peninsula in pre-break-up Gondwana reconstructions.

The Denaro Complex, a portion of the Madre de Dios terrane, is composed of metamorphosed pillow basalt, metahyaloclastite, banded metalliferous and radiolarian metachert, metapelite and redeposited calcerous metasandstone (Sepúveda et al., 2008). Basaltic rocks show primary textures, minerals and structures and are foliated in the vicinities of thrust faults. The rocks are interpreted to have developed during the accretion of the terrane to the Gondwana margin (Sepúveda et al., 2008). Composition of augite and chromite cystals indicates a relationship to volcanic rocks erupted along a construction plate margin, likely in a spreading axis-centered oceanic plateau or ridge (Sepúveda et al., 2008). Metamorphic assemblages of pumpellyite-actinolite facies reveal metamorphism in a frontal accretionary wedge at elevated pressure and low temperature conditions. The assemblages may be related to the Late Triassic-Early Jurassic Chonide event. The Chonide event has been recognized in other regions of the Patagonian Andes (Sepúveda et al., 2008).

## South Africa

In South Africa, the geology of the terranes within the Neoproterozoic Zambezi belt suggest that the Congo and Kalahari cratons have been a cohesive block since approximately 1.1 Ga. According to Wilson et al. (1997), the Khomas/Adamastor Ocean did not transect the Congo/Kalahari block. It is instead proposed that any tectonic connection with the Neoproterozoic-Cambrian collisional plate boundary in the Mozambique belt was likely in the shear zones like the Mwembeshi dislocation. The Mozambique belt was active ~ 550 Ma.

The record of sediment accumulated in the Karoo Basin is directly tied assembly and breakup of Pangea. The tectonic setting of the basin was defined by compression and accretion along the southern margin of Gondwana simultaneously with extension growing into the supercontinent from the Tethyan margin (Duncan et al., 1997). The combination of tectonic stressed from the convergent and divergent margins of Gondwana formed several various types of basins across Africa. Karoo sedimentation continued until the breakup of Gondwana in the Middle Jurassic (Duncan et al., 1997). The Karoo Basin was replaced by the emplacement of a large igneous province. The youngest preserved Karoo deposits in the upper portion of the Karoo deposit varies from Triassic to Middle Jurassic due to erosion post-Gondwana breakup (Cantuneanu et al., 2005). The sediment within the Karoo basins accumulated due to tectonism and climate allogenic processes. Tectonism was dominantly flexural in the south in relation to processes of subduction, accretion and mountain building along the paleo-Pacific margin of Gondwana. In the north tectonism was extensional in relation to spreading processes along the Tethyan margin of Gondwana (Cantuneanu et al., 2005).

A portion of the Pan-Gondwanian fold-thrust belt that formed along the convergent margin of Gondwana is preserved in South Africa as the Cape Fold Thrust Belt, as mentioned previously in this paper. The orogen led to the formation of the Karoo retroarc foreland system which includes the main Karoo Basin as well as other smaller Karoo basins (Cantuneanu et al., 2005).

The beginning of subduction and tectonic loading is dated as Namurian coinciding with the progression of a subduction related volcanic arc adjacent to the

Panthalassan continental rim of Gondwana and marked the transition from a divergent margin to a convergent margin (Smellie, 1981; Mpodozis and Kay, 1992). The beginning of subduction, compression and tectonic loading along the southern margin of Gondwana dates to the Early Carboniferous and is consistent with the preliminary deformation discovered for the South America segment of the Pan-Gondwanian Mobile Belt. The end of the oldest major tectonic event identified in the Cape Fold Belt is dated as Late Carboniferous indicating an area of active tectonic loading during the sedimentation of the oldest deposits of the Karoo Supergroup (Hälbich et al., 1983). Cantuneanu et al. (2005) concludes that the beginning of the Cape Orogeny and the related Karoo foreland system precedes the oldest preserved Karoo sedimentary rocks and is properly placed in the Namurian.

Sedimentation patterns in the Karoo basins were additionally influenced by climate changes. Climate shifted from cold conditions during the Late Carboniferous-Permian to warmer and eventually hot conditions with fluctuating precipitation during the rest of the Karoo time (Cantuneanu et al., 2005).

## Antarctica

The Ellsworth Mountains in West Antarctica characterize a portion of a displaced terrane that was once situated along the paleo-Pacific margin of Gondwana (Curtis et al., 1999). Prior to the break-up of Gondwana, it was positioned adjacent to South Africa and the Weddell Sea coast of East Antarctica. The southern Ellsworth Mountains contain Middle Cambrian sedimentary rocks that contain thick volcanic and subvolcanic rocks forming five igneous centers (Curtis et al., 1999). Most of the igneous samples are mafic and are geochemically varied. They range from mid-ocean ridge basalts (MORB) to shoshonitic and lamprophyric. Silicic rocks comprise of melt segments from Late

Proterozoic crust which is interpreted to be from the basement of the Ellsworth Mountains. Curtis et al. (1999) infer that these igneous rocks formed in a continental rift environment where the MORB-like basalts erupted near the rift axis and melts from the lithosphere emplaced on the rift shoulder. The Middle Cambrian rift event is geologically consistent with the rift-related sedimentary rocks in South Africa (Curtis et al., 1999).

The eastern Palmer Land Shear zone, located in the southern Antarctic Peninsula, is a major ductile fault zone that separates multiple geological domains ((1) Parautochthonous Eastern Domain; (2) Allochthonous, microcontinental, magmatic arc suspect Central Domain; (3) A possible allochthonous, subduction-accretion complex, or separate crustal fragment suspect Central Domain) (Vaughan and Storey, 2000). The shear zone shows signs of Late Jurassic terrane accretion and collision of two, possibly three separate terranes. The accretion events created the Palmer Land Orogeny. Eastern Palmer Land contains the best developed portion of the orogeny (Vaughan and Storey, 2000). Eastern Palmer Land contains shallow-marine sedimentary rocks of the Latady Formation and a metamorphic and igneous basement complex of possibly Lower Paleozoic to pre-Early Jurassic which are thrust and folded forming a foreland, fold and thrust belt that runs parallel to the axis of the Antarctic Peninsula (Vaughan and Storey, 2000).

Present day outcrops in Northern Victoria Land, which border the East Antarctic Craton, record the Ross Orogenic event. They define the basement of the Cenozoic Transantarctic Mountains which extend from Northern Victoria Land to the Pensacola Mountains (Federico et al., 2006). Northern Victoria Land contains the widest exposure of Ross orogenic rocks which include some without equivalents anywhere else in the Transantarctic Mountains. There have been several tectonic models proposed for the Ross orogeny over the years that have focused on reassembling the terranes present in

Northern Victoria Land (Federico et al., 2006). Models have included: (1) Westward dipping subduction under the craton (Kleinschmidt and Tessensohn 1987; Flöttmann et al., 1998); (2) Eastward dipping subduction followed by westward dipping subduction (Findlay et al., 1991); (3) Eastward dipping subduction and continent-arc collision (Wodzicki and Robert 1986; Meffre et al., 2000); (4) Westward dipping subduction followed by strike-slip faulting (Weaver et al., 1984). According to Federico et al. (2006), the most recent data collected from this region do not fit together with the proposed models. Federico et al. (2006) proposed a Ross orogeny model that assumed a southwest dipping subduction under the east Antarctic Craton starting from the early Cambrian places the Wilson and Bowers terranes on the upper plate. The lower plate carried a continent which consisted of crust that is presently buried underneath the Robertson Bay terrane turbidites and possibly in western Marie Byrd Land, Campbell Plateau and the Western Province of New Zealand (Federico et al., 2006), suggesting that these continental fragments accreted to east Antarctica during the Cambrian Ross Orogeny and not during the Late Devonian (Weaver et al., 1991; DiVenere et al., 1996). Federico et al., (2006) postulates that during the Mid-Cambrian, an extensional stage in the overriding plate resulted in the opening of a back-arc basin, possibly due to the roll-back of the subducting plate. The Sledgers Group of the Bowers terrane represents an arc/back-arc association accounting for the variety of volcanites and mixed continental/oceanic sediment (Federico et al., 2006). After the back-arc opening, the continuing convergence led to the collision of the lower plate continent with the margin of Northern Victoria Land. After the collision, the basin entered a stage of compression with the development of a fault bounded uplifted zones and related basins (Federico et al., 2006). When the lower plate continent entered the subduction zone, subduction slowed down and eventually spread to the back-arc. The projected subduction depth exceeded

90 km (Federico et al., 2006). The Ross orogeny reached its final setting during the Early Ordovician. The Robertson Bay terrane contains only subduction-related sediments which may have been decoupled from their original oceanic basement as a product of tectonic substitution. They reside on the continental crust of the lower plate continent (Federico et al., 2006).

## New Zealand

Before sea floor spreading during the Late Cretaceous, New Zealand was a single landmass with Australia and Antarctica. Very little continental crust in Zealandia is exposed above sea level in the North and South islands (Mortimer, 2004). No Precambrian cratonic core is exposed inland New Zealand. The basement is Cambrian to Early Cretaceous and contains geological features consisting of nine significant volcano-sedimentary terranes, three composite regional batholiths, and three regional metamorphic-tectonic belts that overlay the terranes and batholiths (Mortimer, 2004).

According to Mortimer (2004), the Phanerozoic tectonic history of New Zealand is interpreted as a progressive Pacific-ward growth of the Gondwana/Pangea supercontinent by terrane accretion and batholith intrusion at an oblique convergent margin. The Buller and Takaka terranes contain Gondwana outcrops indicating that they accreted to Gondwana by the Devonian. Oblique dextral transpression may have existed at the New Zealand Gondwana margin that lasted from the Carboniferous to the Cretaceous (Mortimer, 2004), explaining the phases of magmatism in the Median Batholith, the southward relocation of the Rakaia terrane from Queensland and the expansion of the Caples-Haast Schist-Rakaia-Bay of Islands-Pahau-Waipa accretionary wedge (Mortimer, 2004).

Oceanic-continental subduction occurred between 85 - 120 Ma (Bradshaw, 1989). The subduction interval may have been strongly influenced by a spreading ridge against the Gondwana margin, stalling of a spreading ridge offshore (Luyendyke, 1995) and/or collision of the Hikurangi Plateau with the Mesozoic accretionary wedge (Mortimer, 2004). The geologic record inland shows that the main pulse of normal I-type calcalkaline magmatism in the eastern Median Batholith stopped about 125-130 Ma but was directly succeeded by the capacious intrusion of sodic, adakitic, high Sr calcalkaline magmas until 105 - 100 Ma (Mortimer, 2004). The adakitic plutons are found primarily in the western third of the batholith and have been explained as a MTZ-Western Province collision (Muir et al., 1995) shallowing slab and widening arc trench gap (Mortimer et al., 1999) and/or continentward underthrusting and deepening of the batholith root (Tulloch and Kimbrough, 2003). The crustal thickening is associated with granulite formation in Fiordland. The resulting rapid exhumation of mid-deep crustal gneiss is recognized in the 110 - 90 Ma growth of the Fiordland and Paparoa core complexes (Mortimer, 2004). Subduction complexes in the eastern North Island contain 100 Ma detrital zircons which indicate that subduction continued beneath the New Zealand Eastern Province until at this this point (Kamp, 2000).

The 100 Ma Tapuaenuku alkaline rocks and fault-controlled basins of cover strata indicate a rapid transition to an extensional regime (Laird and Bradshaw, 2004). Continental extension concluded in Tasman Sea and Southern Ocean seafloor spreading from ca. 85 Ma. (Mortimer, 2004). The New Zealand basement was exposed to reestablished deformation in the Neogene with the initiation of the contemporary Australia-Pacific plate margin (Mortimer, 2004).

## Australia

Paleopole data taken from the main craton in Australia and in multiple terranes in the Tasman Fold Belt have followed the same apparent polar wandering path since 400 Ma for the Lachlan and Thomson superterranes. It was not until 250 Ma or sooner that the New England superterrane followed the same path (McElhinney et al., 2003). The paleopole and geologic data suggest that the Lolworth-Ravenswood and Molong-Monaro terranes were amalgamated to the Australian craton by the Late Devonian. The Tamworth terrane likely accreted with the Lachlan superterrane by the Late Cretaceous. There is a possibility for movement in that accretion area until the Middle Triassic (McElhinney et al., 2003).

The Tasmanides occupy the eastern third of the Australian continent and is part of the larger Gondwana orogeny. The geology of the Tasmanides records the break-up of Rodinia and the opening of the proto-Pacific Ocean in the Neoproterozoic until the start of convergent margin tectonism around the Middle Cambrian (Glenn, 2006). Convergence occurred along the eastern Gondwana margin until the Late Triassic. From the Middle Cambrian to the Late Triassic, the Tasmanides underwent extensive stages of sedimentation, arc and back-arc action alongside the eastern Gondwana margin fixed with the development of regions of subduction associated to generally west-dipping subduction (Glenn, 2006), followed by short lived, potentially orthogonal and highly oblique accretion of turbidite terranes of cratonic origin, island arcs and subduction clusters which developed along the east Gondwana margin. Closing of the back arc basins occurred subsequent to the events cited above (Glenn, 2006).

The Tasmanides can be divided into five orogenic belts in order from west to east: 1) the Delemarian orogen; 2) the Lachlan orogen; 3) the Bowen-Gunnedah-Sydey Basin System and the southern New England orogen; 4) the Thomson orogen; and 5) the

North Queensland orogen (Glenn, 2006). The Delemarian orogeny was deformed by the Middle-Late Cambrian Delemarian orogeny, the Lachlan orogen was deformed at the end of the Ordovician by the Benembran orogeny, as well as in the Middle Devonian by the Tabberabberan orogeny and in the Early Carboniferous by the Kanimblan orogeny. The rocks get younger from west to east as they moved away from the Australian craton (Glenn, 2006). Glenn (2005) does list two qualifications regarding the accretion model for the Tasmanides. 1) The North Queensland orogen show no signs of accretion where Neoproterozoic and Permian rocks are stacked on top of each other and remain close to the cratonic margin. Rollback of the proto-Pacific plate only occurred in the southern Tasmanides. 2) Rocks of the Delemarian and Lachlan supercycles occur in the New England orogen (Glenn, 2006). The rocks mentioned in the accretion model demonstrate the rollback of the southern part of the proto-Pacific Plate. They represent pieces of older cycles that were rifted off during rollback and forming partial basement to the new cycle (Glenn, 2006).

Figures 5-1 – 4: Reconstruction of the Southern Gondwana Margin showing cratonic and Brasiliano–Panafrican–Early Palaeozoic elements. Cratons shown: ANT, Antarctic; AUS, Australian (inc. Yilgarn, Pilbara, Gawler and Musgrave); K-G, Kalahari–Grunehogna; LA, Luis Alves; M, Mawson; P, Paraná; RA, Río Apa; Accreted terranes shown: South America (Ramos, 2004; 2008), A–A, Arequipa-Antofalla massif; CH, Chilenia; CY, Cuyania; PA; Pampia; NP, North Patagonian Massif; PT, Patagonian; South Africa (Johnston, 2000; Dalziel et al., 2000; Shone and Booth, 2005); Antarctica (Pankhurst et al., 1998 a,b; Vaughan and Storey, 2000; Collinson et al., 2006), EM, Ellsworth–Whitmore Mountains; TAM, Transantarctic Mountains; ED, Eastern Domain; CD, Central Domain; WD, Western Domain; RP, Ross Province; WP, Western Province; New

Zealand (Mortimer, 2004), EP, Eastern Province, WP, Western Province; Australia (McElhinney et al., 2003), AFB, Adelaide Fold Belt, LFB, Lachlan Fold Belt, TFB, Thompson Fold Belt, NEFB, New England Fold Belt.

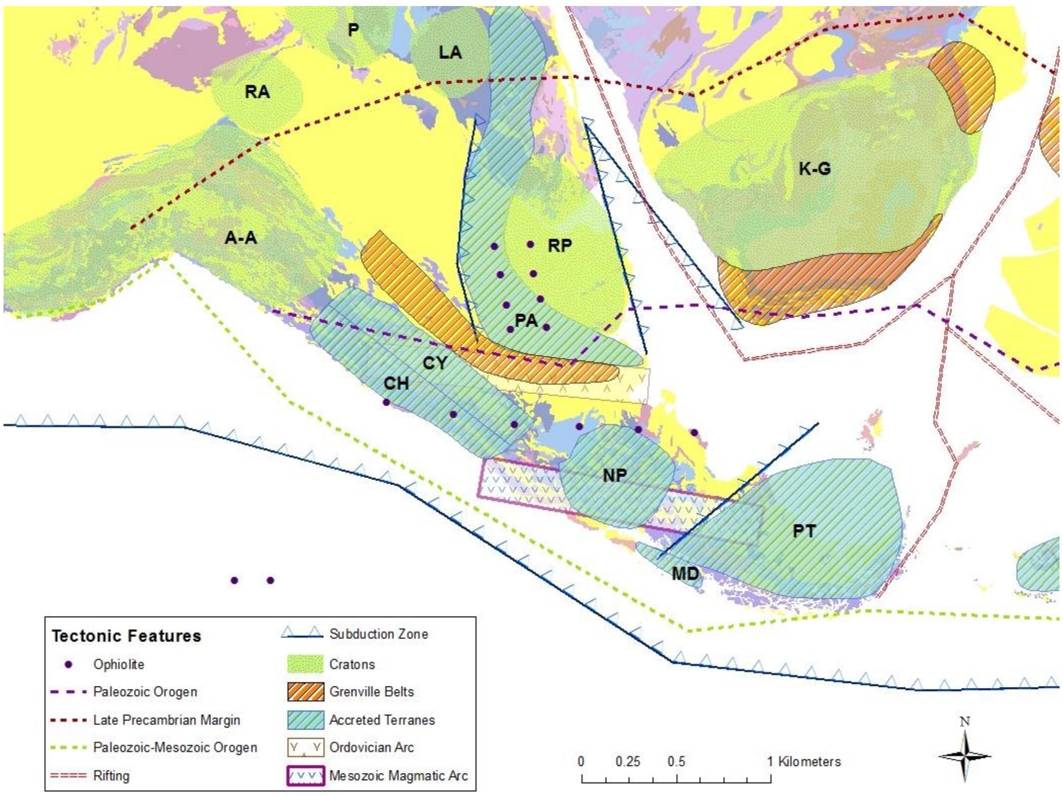


Figure 5-1 Plate reconstruction of Southern South America and Southern South Africa.

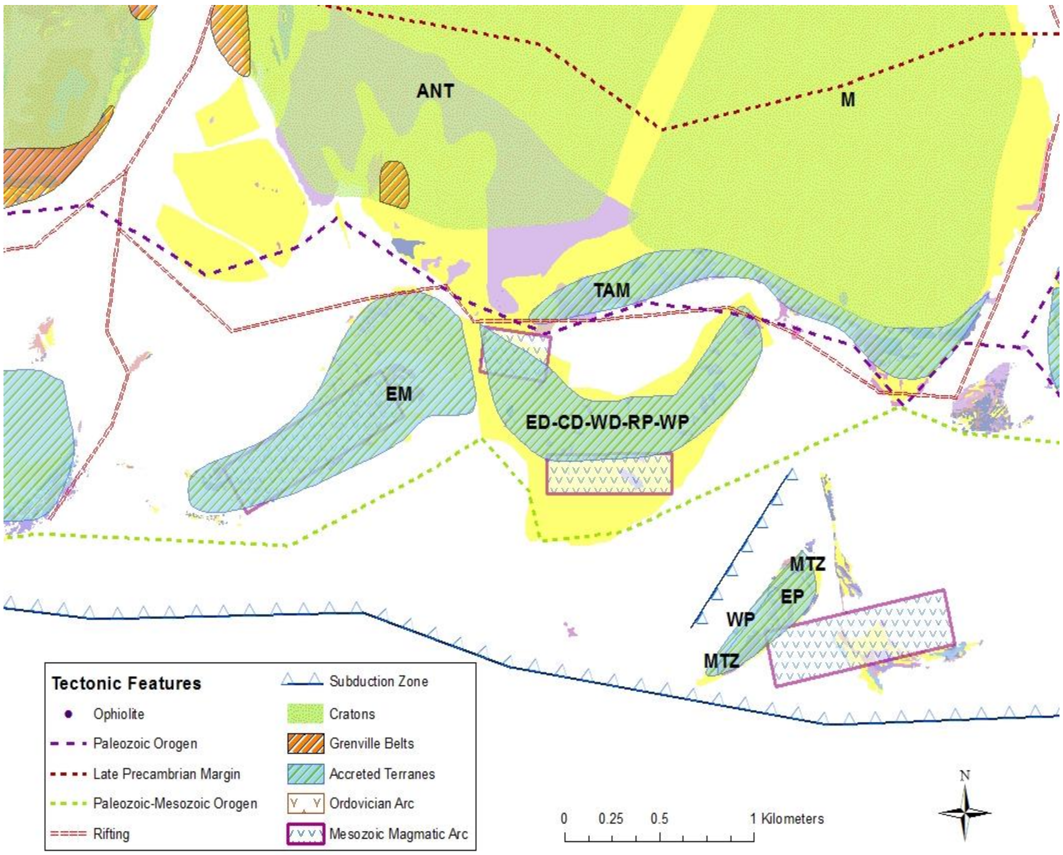


Figure 5-2 Plate reconstruction of Antarctica and New Zealand.

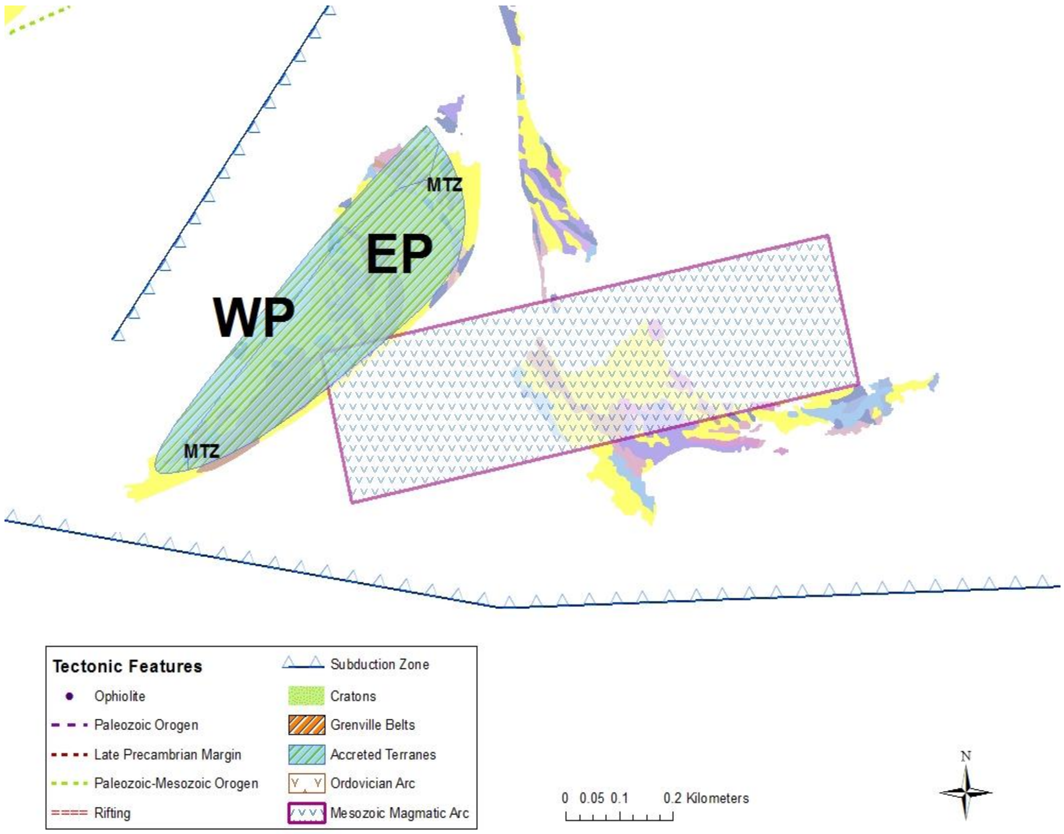


Figure 5-3 Plate reconstruction of New Zealand.

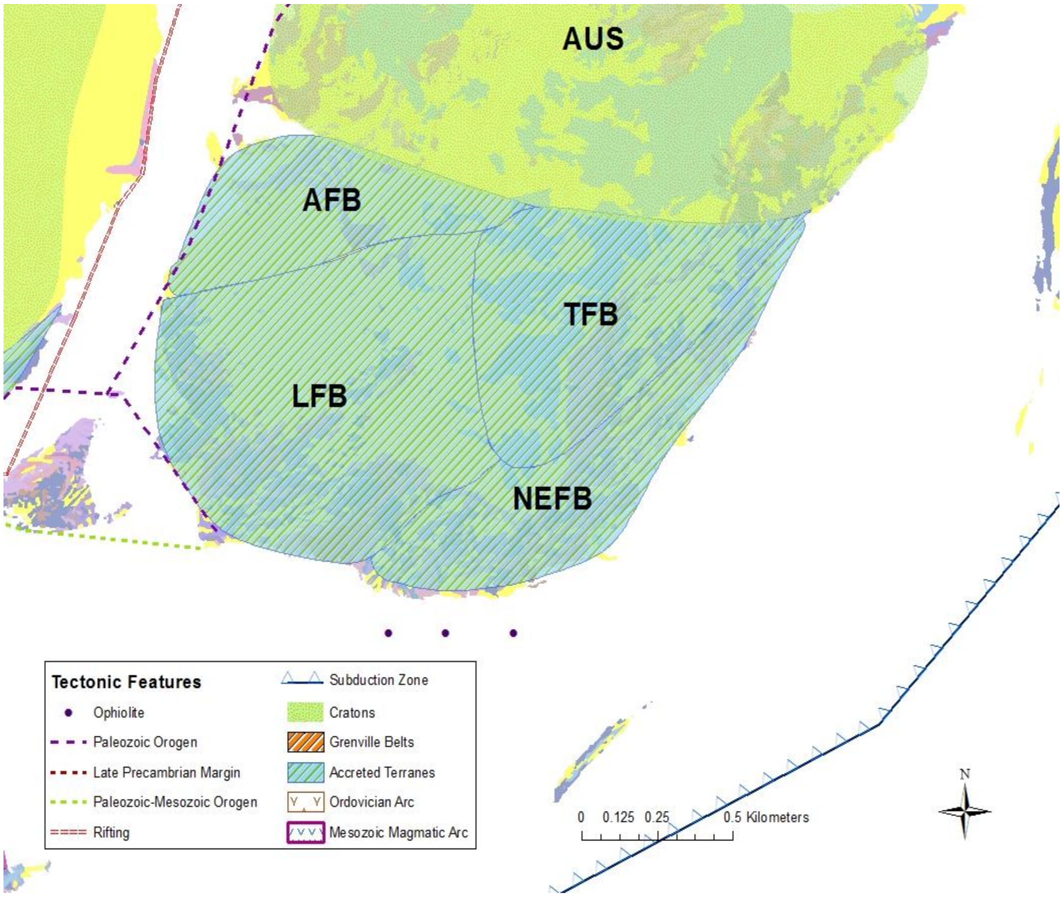


Figure 5-4 Plate reconstruction of Australia.

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