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South Georgia Microcontinent: Displaced Fragment of the Southernmost Andes

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Abstract

The mountainous, glaciated island of South Georgia is the crest of one of the most isolated fragments of continental crust on Earth. It is located approximately 1700 kilometres east of the southern termination of the Andean Cordillera of South America. The island is primarily composed of Lower Cretaceous turbidites, the infill of a marginal basin floored by stretched continental and ophiolitic crust. Remnants of a volcanic arc are preserved on offshore islets to the southwest. The Pacific hinterland of the southernmost Andes is missing in Tierra del Fuego, terminating at a submarine escarpment forming the continental margin immediately east of Cape Horn. The arc and back-arc basin infill rocks of South Georgia correspond exactly to part of the missing Cordilleran hinterland.

The mechanism of transport of the South Georgia microcontinent eastward relative to South America remains obscure, but likely involved some form of 'escape tectonics' during mid- to Late Cretaceous counterclockwise rotation of the arc that led to closure and inversion of the marginal basin.

1. Introduction: where the Andes end

In "Facts and Theories on the Andes", the *26th William Smith Lecture* before the Geological Society of London, Augusto Gansser, legendary Swiss geologist and 'Father of Himalayan geology', recounted sitting on a promontory in eastern Trinidad looking at the Northern Range of the island plunge into the Atlantic Ocean. Mindful of the many thousands of kilometres of the Andean Cordillera that lay behind him, he mused that he was in "an ideal spot for geological meditation" (Gansser, 1973). In 1989 he travelled to the southernmost Andes on a field excursion led by one of the authors (IWDD) in connection with the 28th International Geological Congress. Rounding, in an open boat, the rugged fur seal-infested promontory at Roca Sapo on the eastern extremity of Isla de los Estados where the main range of the Cordillera plunges into the Atlantic and large standing-waves mark the northward flow of the Antarctic Circumpolar Current out of Drake Passage (Figs 1 and 2), he remarked that this was "not an ideal spot for geological meditation".

_____________________-Abbreviations Footnote about here

Figure 1 about here
1. Location map for the South Georgia microcontinent, southernmost Andes and Scotia Arc (GeoMapApp). BwB–Burdwood Bank; FAFZ–Falkland-Agulhas Fracture Zone; IE–Isla de los Estados; MEB–Maurice Ewing Bank; NGR–Northeast Georgia Rise; SOI–South Orkney Islands; SSI–South Sandwich Islands; SShI–South Shetland Islands. X–Shag Rocks, o–Clerke Rocks

Charles Darwin, sighted Isla de los Estados, "the rugged, inhospitable Staten-land", as a mere outline "amidst the clouds" from the deck of HMS *Beagle* in December 1832 on his first visit to Tierra del Fuego. As he sailed southwestwards towards Cape Horn he inferred that "This part of Tierra del Fuego may be considered as the extremity of [a] submerged chain of mountains….." (Darwin, 1845). The higher and even more rugged island of South Georgia lies ~1700 km due east of Isla de los Estados along that chain, now known as the North Scotia Ridge and part of the Scotia Arc, which physiographically and tectonically links South America and Antarctica (Figs. 1,3). The first sighting of that island took place in the late 17th Century, but it was not until 1775 that Royal Navy Captain James Cook made the first landing on the uninhabited island, naming it and charting the NE coast.

Figure 2 about here
2. Physiography of the southern extremity of the Andes (GeoMapApp). CD–Cordillera Darwin; CH–Cape Horn; IE–Isla de los Estados; IH–Isla Hoste. IN–Isla Navarino; PM–Peninsula Mitre; SM–Strait of Magellan

Examination of the physiographic termination of the Andean Cordillera in Tierra del Fuego reveals two domains: the main range north of the Beagle Channel continues eastward from Cordillera Darwin through Peninsula Mitre into Isla de los Estados; in contrast the Pacific hinterland south of the channel terminates abruptly in a northeast-southwest trending segment of the continental margin immediately east of Cape Horn, here termed the Cape Horn escarpment (Fig. 2). The escarpment is the rifted continental margin formed at approximately 30 Ma when the now extinct northeast-southwest spreading ridge of the west Scotia Sea was initiated (Barker and Burrell, 1977; Eagles et al., 2005). Clearly an easterly continuation of the Pacific hinterland of the Andes that formerly lay to the south of Isla de los Estados was rifted from this margin. Where is that former continuation?

Figure 3 about here

3. Physiography of the Scotia Arc region with path of the Antarctic Circumpolar Current (GeoMapApp; modified from Dalziel et al. 2013a).

ANT–Antarctic plate; **SAM**–South American plate; **SCO**–Scotia plate.

Physical features: ASSA–Ancestral South Sandwich arc (Dalziel et al., 2013b; Pearce et al., 2014). BP–Barker Plateau; BwB–Burdwood Bank; BB–Bruce Bank; BP–Barker Plateau; CHE–Cape Horn escarpment; DB–Discovery Bank; DvB–Davis Bank; FAFZ–Falkland-Agulhas Fracture Zone; HB–Herdman Bank; JB–Jane Bank; NGR–Northeast Georgia Rise; PB–Pirie Bank; SFZ–Shackleton Fracture Zone; SRB–Shag Rocks Bank; SSI–South Sandwich Islands volcanic arc; SG–Island of South Georgia; SOI–South Orkney Islands microcontinent; TR–Terror Rise. X–Shag Rocks o–Clerke Rocks.

Red lines denote plate boundaries, with transform segments indicated by red arrows, active spreading centers by double red lines, teeth indicating the upper plate of active subduction zones. Black triangles represent active volcanoes.

Ocean currents and fronts (from Naveira Garabato et al., 2002): ACC–Antarctic Circumpolar Current; PF–Polar Front; SACCF–Southern Antarctic Circumpolar Current Front; SAF–Subantarctic Front; SB–southern boundary of the ACC. The Polar Front is the core of the flow of Circumpolar Deep Water of the ACC. White arrows show the three main pathways of the ACC.

Gateways for ocean currents: ESSP–East Scotia Sea Passage: SGP–South Georgia Passage; SRP–Shag Rocks Passage

Bold line NV is seismic refraction profile of Ludwig et al. (1968) shown in Figure 4b. Red numerals denote the approximate original positions of the South Georgia microcontinent according to Dalziel et al. (1975) and Tanner (1982a), #1, and Eagles (2010), #2.

The first Europeans to study the Andean Cordillera, familiar with the western Alps, Appenines, Dinarides and Hellenides, were puzzled by the paucity both of turbiditic 'flysch' and of the 'Steinmann trinity', the mafic and ultramafic ophiolitic rocks and radiolarian chert that characterize these Tethyan 'geosynclinal' orogens. This led Steinmann himself to suggest that the Cordillera in Peru was merely an embryonic 'geosyncline', or the fringe of one (Sengor and Natal'in, 2004), while French geologist Jean Aubouin proposed that the Andes were a different class of mountain belt, originating in 'liminal' basins on the edge of an ocean basin rather than the 'geosynclinal' basins that were believed to give rise to the Alpine ranges between the colliding European and African continents (Aubouin, 1972).

It is only in the southernmost extremity of the Andes in Patagonia and Tierra del Fuego that the thick turbidite sequences, ophiolitic complexes and deep marine cherts they had expected are found in abundance. These remote regions of the main Cordillera are only accessible from the sea. The first geologist to work in this area was Charles Darwin, who visited on board HMS *Beagle* in 1832 and 1834. Among many other geologic observations in the area, he recognized that the flysch was Mesozoic; he also made the first geological map of the region; sadly, this was never published (Dott and Dalziel, 2016). From Darwin's visit until the mid-20th Century, scientific visits were few and far between.

Following the Second World War, Chile initiated a program of petroleum exploration in the southern Province of Magallanes led by geologists of the Empresa Nacional del Petróleo (ENAP). Mapping in the Cordillera, supported mainly by small fishing boats, led to the recognition of an orogen-parallel belt of mafic 'Rocas Verdes' (green rocks). Swiss-born ENAP geologist H.R. ('Rudy') Katz interpreted these as representing the 'eugeosynclinal' component of a 'mio-eugeosynclinal couple' with the non-volcanic Magallanes-Austral basin foredeep to the east, located on the continental side of a Pacific margin volcanic arc now mainly represented by the underlying Patagonian granitic batholith (Katz, 1964, 1972). He further suggested that the Rocas Verdes belt represents the uplifted floor of a basin comparable to those beneath the marginal seas behind the active volcanic arcs of the western Pacific.

In the late 1960's one of us (IWDD) initiated a study of the structure and tectonic history of the southernmost Andes in collaboration with the ENAP geologists and sedimentologist Robert H. Dott Jr., his then-colleague in the University of Wisconsin–Madison. At that time the seafloor-spreading/plate tectonics paradigm was being developed and 'ophiolites' were being recognized as a pseudo-stratigraphic sequence of igneous rocks representing fragments of the floor of ocean basins or marginal seas (Conference Participants, 1972). In collaboration with then post-doctoral fellows Maarten de Wit and Charles Stern at Lamont-Doherty Geological (now Earth) Observatory of Columbia University, an expedition to the Patagonian Andes was planned to test the hypothesis that the mapped occurrences of mafic Rocas Verdes might indeed represent fragments of the floor of a former small ocean basin marginal to the Pacific.

Two fjords between 52° and 53° South, Fjordo Lolos and Fjordo Encuentro were selected for study because they cut across part of the north-south trending Rocas Verdes outcrop belt. Layered gabbros were found at sea level, vertical north-south striking sheeted dykes at higher elevations transitioning upwards into pillow lavas. Only the ultramafic component of a complete ophiolite complex is missing in outcrop; presumably, if preserved, it lies below sea level (Dalziel et al., 1974a). The localities studied are part of the largest ophiolitic body in the southern Andes, the Sarmiento Complex (Fig. 4a).

Figure 4 about here
4a. Simplified geologic map of the southernmost Andes (GeoMappApp) showing Antarctic, Scotia and South American plates

BL–Bahia Lapataia; BT–Bahia Tekenika; CD–Cordillera Darwin; IE–Isla de los Estados; IH–Isla Hoste; IN–Isla Navarino; MFFS–Magallanes-Fagnano fault system; NSRT–North Scotia Ridge transform; P–Punta Arenas; PH–Peninsula Hardy; SC–Sarmiento Complex; TC–Tortuga Complex. Red arrow points to location of Fjordo Lolos and Fjordo Encuentro within the Sarmiento Complex (see text)

b. Seismic refraction profile from Falkland Platform (Plateau) across the Falkland Chasm (Trough) and Burdwood Bank to the Scotia Sea (Ludwig et al., 1968). For location see Line NV on Figure 3.

Adding to the earlier mapping by Katz and his ENAP colleagues Sergio Cėspedes and Raul Cortés, the Columbia University project was able to trace the mafic component of the Rocas Verdes basin from north of Cordillera Sarmiento to Isla Navarino in the south (approximately 48-55° S; Fig. 4a). This included two major mafic massifs: Cordillera Sarmiento and also Cerro Tortuga on Isla Navarino, south of the Beagle Channel (Stern and de Wit, 2003). Initial opening of the Rocas Verdes marginal basin was accompanied by continental rifting associated with the earlier eruption of the widespread Late Jurassic silicic 'Tobífera' (tuffaceous) volcanic province, known as the Tobifera Formation in Chile and the Lemaire Formation in Argentina (Bruhn et al., 1978). The regional interpretation was carried forward through mapping by the ENAP geologists and Dott and his students (Katz and Watters, 1966; Scott, 1966). The Katz and Watters collaboration had arisen opportunistically when Bill Watters, a geologist from New Zealand, joined a 1958-1959 British Royal Society expedition to southern Chile and met Katz unexpectedly on Isla Navarino. The sum of these various contributions demonstrated that the basin was infilled by Lower Cretaceous volcaniclastic turbidites derived mainly from the Pacific margin volcanic arc. Closure and tectonic inversion of the basin in the first pulse of Andean compression pre-dated intrusion of Late Cretaceous calc-alkaline plutons and was accompanied by the mid-Cretaceous downwarp of the Magallanes-Austral foredeep to the east (Natland et al., 1974). Thus the geologic structure of the southernmost Andes is dominated by the uplifted remnants of a Late Jurassic to Early Cretaceous "fossil marginal basin" with ophiolitic complexes complete except for the ultramafic component.

In the far south, round the Patagonian orocline (Carey, 1955, 1958) in Tierra del Fuego, the main Cordillera continues eastward to strike out to sea in easternmost Isla de los Estados; this island is dominated by the outcrop of highly deformed Upper Jurassic silicic volcanics of the Tobífera/Lemaire Formation (Dalziel et al., 1974b; Dalziel and Palmer, 1979). Seismic refraction data indicate that these rocks continue east along the southern edge of the Burdwood Bank, the westernmost segment of the submerged North Scotia Ridge, and revealed as high-velocity basement (Ludwig et al., 1968; Fig. 4b). South of Beagle Channel, it is the ophiolitic floor and Lower Cretaceous sedimentary infill of the Rocas Verdes marginal basin plus the Patagonian batholith (with remnants of its former volcanic superstructure), forming the Pacific hinterland of the southernmost Andes, that are abruptly truncated at the northeast-southwest trending Cape Horn escarpment (Figs.1-4). We refer to the continental margin re-entrant between the Cape Horn escarpment and the Isla de los Estados-Burdwood Bank continental margins as the Staten embayment.

The purpose of this contribution is to review what we believe to be incontrovertible geological evidence for the Andean origin of the isolated South Georgia microcontinent, now situated more than 1,500 km east of its original position. Despite a consensus for this correlation developed by a century of geological fieldwork, there is a current tendency to dismiss such evidence if it does not conveniently fit with regional plate modelling of the large geophysical databases that are now available. As one recent example of that approach (but by no means the only one), we would draw attention to Eagles and Eisermann (2020). These authors, amongst other controversial proposals, reject – with slight disdain – the Andean correlation of South Georgia because their regional model derived from an interpretation of geomagnetic data "rules out venerable correlation-based interpretations for a Pacific margin location and subsequent long-distance translation of the South Georgia microcontinent" (*op. cit*., abstract). We find this approach unhelpful and untenable. Whilst the exact tectonic mechanism of separation of the South Georgia microcontinent from the South American continent and its relative eastward translation remains tantalizingly obscure, in our view this in no way detracts from the robust evidence of South Georgia's Andean origin.

2. South Georgia: emergent crest of a microcontinent

2.1 Historical background

Delineation of the Scotia Arc as a submarine feature linking emergent islands and surrounding the Scotia Sea (Figs 1, 3) was one of the principal achievements of the 1902-04 Scottish National Antarctic Expedition (Bruce, 1905). The name *Scotia* commemorates that expedition's ship, but speculation as to the Arc's likely existence began earlier. Those historical accounts were reviewed by Stone (2015a) with the first such suggestion from the 1819-21 Russian expedition, after which Bellingshausen (1831) wrote of a submerged mountain chain linking the South Sandwich Islands with South Georgia and continuing westward to South America through the Falkland Islands (1945 English translation, p. 110). The concept of a once-continuous mountain range linking the Andes and the mountains of the Antarctic Peninsula was geologically speculative until Arçtowski (1901) collected metamorphic, volcanic and intrusive igneous rocks from the west coast of the Peninsula during the 1897-99 Belgian Antarctic Expedition. These were thought comparable to similar lithologies known from the southern Andes and so supported the pre-expedition arguments favouring an Andean–Antarctic connection.

Despite the Scottish oceanographical work, there was reluctance to carry the Scotia Arc fully eastward from South America to South Georgia and the South Sandwich Islands. Even Suess (1909, p. 488-497), whilst supporting the idea of Andes-Antarctic Peninsula geological continuity in principle, and noting the existence of the Scotia Arc as a submarine feature, preferred to take that connection directly from the Burdwood Bank to Elephant Island and the South Shetland archipelago off the northern Antarctic Peninsula (Fig. 3). This neglect of South Georgia, the largest of the islands on the northern limb of the Arc (now the North Scotia Ridge) was perhaps understandable given contemporary uncertainties over the island's geology, but by the early 20th century those uncertainties were becoming resolved.

The history of geological exploration of South Georgia has been reviewed by Stone (2015a) who also provides an extensive bibliography. The earliest contributions of note arose, *en passant*, from the 1901-03 Swedish South Polar Expedition (Andersson, 1907) and the 1911-1913 German South Polar Expedition (Heim, 1912). They described a folded succession of greywacke and mudstone, at least part of which was of Mesozoic age, that extended over most of the island, but with a large body of intrusive igneous rocks at its south-east extremity. The German expedition had benefitted from the logistical support provided by South Georgia's newly established, shore-based whaling industry, and it was an initiative by the Salvesen whaling company that led next to the attempt by Ferguson (1915) to establish a lithostratigraphy for the sedimentary succession. Ferguson's scheme was dismissed as untenable by Wordie (1921), Holtedahl (1929) and Douglas (1930) but, together with the series of petrographical descriptions by Tyrrell, culminating in his work on Douglas's specimens (Tyrrell, 1930), it did establish the overall nature of the South Georgia succession, allowing tentative regional comparisons.

Those comparisons were made against the background of the continental drift debate provoked by the proposals of Wegener, published from 1915 onwards and maturing into the later editions of his seminal book, in English translation *The Origin of Continents and Oceans* (Wegener, 1929). The initial, widespread rejection of continental drift is well documented (e.g., Oreskes, 1999) but against the majority opinion, the Scotia Arc continued to be cited by drift enthusiasts as evidence in its favour; the development of that regional discussion has been reviewed by Stone (2015b). Wegener (1929, p. 94) had described the evolution of the Arc as a consequence of the separation of Africa, South America and the Antarctic Peninsula: "the écheloned chains were torn off *seriatim* from the drifting blocks and then left behind". Influential support for the idea came from Holmes (1929, p. 595) who also described the distribution of the islands of the Scotia Arc as arising from the westward drift of South America and the Antarctic: "…a lag effect due to a strung–out belt of sial…left between the two continental blocks as they advanced" (Fig. 5a).

Figure 5 about here
5a. Evolution of the Scotia arc according to Holmes (1929). Bold arrow added to emphasize supposed incursion of Pacific Ocean floor into Scotia Sea.

5b. Evolution of the Scotia arc according to Hawkes (1962).

Interpretations such as those by Wegener and Holmes implied an originally close, 'pre-drift', spatial association between islands such as South Georgia and the southern extremity of South America. Wilckens (1933) noted the broad similarity between the rocks of South Georgia and those of Tierra del Fuego, and Kranck (1932), after working in Tierra del Fuego and establishing the Yahgan Formation on Navarino Island, correlated the lithostratigraphy of that formation with the succession described by Ferguson (1915) from South Georgia. Although the details of Ferguson's stratigraphy had been challenged, Kranck's overall correlation of sedimentary lithologies, with similar Mesozoic ages, was sound. We will expand on it in this review.

For the anti-drift majority, the islands of the Scotia Arc had to be relics either of a sunken trans-Atlantic continent (e.g., Gregory, 1929) or of a land bridge that had linked Tierra del Fuego and the Antarctic Peninsula (Schuchert, 1932). An origin for the arc by means of continental drift was maintained by Du Toit (1937) in his now classic book *Our Wandering Continents*, but from 1939–1945 any further resolution of the problem was deferred by the Second World War. Ironically it was the War which reignited British interest in the region and led to the establishment of small, covert military outposts: *Operation Tabarin*. At war's end, these outposts were expanded as civilian research bases and geological investigations formed part of the scientific programme of the Falkland Islands Dependencies Survey (FIDS), later to become the British Antarctic Survey (BAS).

Two FIDS geologists expanded their survey work in the South Orkney and South Shetland archipelagos into broader assessments of the Scotia Arc. For Matthews (1959), the continuity of the North Scotia Ridge (NSR) posed a problem if the continental fragments represented by South Georgia and Shag Rocks were to be interpreted as relics abandoned during continental drift. What caused the linking submarine topography, why are there no intervening rifts? Subsequent work provided the answer by recognising a transpressional tectonic regime at the South America–Scotia plate boundary wherein accretionary thrust construction accompanied lateral fault movement, thus completing the positive bathymetry of the NSR (e.g., Cunningham et al., 1998) as will be further discussed below.

Hawkes (1962) revisited the idea of the Scotia Arc islands as pieces of an original, rectilinear link between Patagonia and the Antarctic Peninsula that had been swept eastwards by an advance of Pacific crust into the Scotia Sea (Fig. 5b). His model gained a wide audience when included by Holmes (1965) as a figured example in the second edition of the influential textbook *Principles of Physical Geology*. By that time, the plate tectonics concept was emerging from WW2 and Cold War oceanographical surveys (a positive result of submarine warfare) and was rapidly gaining support and credibility, subsuming ideas of the Scotia Arc's origins by means of continental drift (Stone, 2015b). Rudwick (2014, p. 258) has noted that the partisans of the 'new' theory, whilst of necessity acknowledging Wegener, rarely extended that courtesy to the likes of Du Toit and Holmes, let alone the less well-known contributors to the Scotia Arc story.

Detailed knowledge of South Georgia's stratigraphy and structure had been much advanced during the 1950s by the three, privately financed, South Georgia Survey Expeditions described by expedition geologist Alec Trendall half a century later (Trendall, 2011). Two contemporary publications by Trendall (1953, 1959), laid the foundation for the modern interpretation. He identified two sedimentary lithofacies of turbidite character ("Cumberland Bay type" and "Sandebugten type", subsequently defined by Stone, 1980, as formations), confirmed a Cretaceous age for most of the succession, established a coherent structural model, and delineated the igneous complex that forms the extreme south-east of the island. Trendall did not speculate on regional correlations with South America but his work reinforced the comparisons made earlier by Kranck (1932), as did a pioneering sedimentological analysis by Frakes (1966). In the opposite direction, work on the Yahgan Formation of Navarino Island described by Katz and Waters (1966) made the geological similarities to the South Georgia succession indisputable.

2.2 Outline of geology

The island of South Georgia (centred at about 54° 20´ S, 36° 40´ W–Figs.1, 3) is the highest, terrestrial part of the largest continental block within the North Scotia Ridge. It measures about 170 km NW–SE with a maximum width of about 30 km NE-SW, and the central mountainous spine rises to a highest point, Mount Paget, of 2934 m. The SW coast, facing the prevailing westerly winds, is steep and extensively ice-covered, whereas large glaciers flow from the mountains into the deeply indented fjords of the NE coast (Fig. 6); all the glaciers are currently in retreat, those on the NE coast dramatically so (Pelto, 2017). Annenkov Island, about 10 km off the southwest coast of South Georgia, is the largest of several small offshore islands. These, and the ice-free coastal periphery of South Georgia itself, host an abundance of wildlife including several species of penguin, seabirds and seals. South Georgia is administered as a British Overseas Territory (in conjunction with the volcanically active South Sandwich Islands; Figs. 1, 3) but is also claimed by Argentina.

Figure 6 about here
6. The contrasting coast lines of South Georgia.

a. The SW coast at Cape Darnley, Cumberland Bay Formation strata forming steep, ice-bound cliffs plunging dramatically to the sea. Image by Nicholas Bayou.

b. The NE coast at Royal Bay: valley glaciers descend from the mountainous centre of the island (seen here is Smoky Wall 1840 m, formed of Cumberland Bay Formation strata) to a relatively benign coastal fringe. All South Georgia's glaciers are in rapid retreat (Pelto, 2017). This photograph was taken in 1971 and it is doubtful whether the pictured Ross and Hindle glaciers are now visible from this viewpoint near Moltke Harbour. Image by Philip Stone.

The modern geological interpretation, founded on the work of Trendall (1953, 1959), arises from the cooperative efforts of geologists working with the British Antarctic Survey and the United States Antarctic Program, supplemented by the contributions of geologists accompanying small private expeditions. The chronology and bibliography of this scientific endeavour has been reviewed by Stone (2015a). The geological results were summarised by Macdonald et al. (1987) and an up-to-date geological map of the island at a scale of 1:250 000 has been compiled by Curtis (2011), both published by the British Antarctic Survey. A simplified version is shown in Figure 7.

Figure 7 about here
7. Geological maps of South Georgia

a. Geology of the main island of South Georgia showing the onshore position and inferred offshore extension of the Cooper Bay Shear Zone; heavy black line is the thrust separating the Cumberland Bay and Sandegugten formations, teeth on the hanging wall side; codes in italics show: bpm–Barff Point Member of the Cumberland Bay Formation; cbf–Cooper Bay Formation (a correlative of the Sandebugten Formation). Red dashed lines–normal faults with tick on downthrown side.

Place names: CB–Cooper Bay; CI–Cooper Island; DF–Drygalski Fjord; DH–Ducloz Head; GH–Gold

Harbour; GP–Greene Peninsula; HG–Hindle Glacier; LH–Larsen Harbour; LV–Larvik; MC–Macdonald Cove; MH–Moltke Harbour; RG–Ross Glacier; Sb–Sandebugten; SC.–Smaaland Cove; T–Twitcher Glacier

b. Geology of the Annenkov Island area plus other offshore islands and adjacent area of South Georgia (after Tanner et al., 1981); the darker red area on Annenkov Island is the outcrop of the Lower Tuff Member of the Annenkov Island Formation; all other areas of Annenkov Island plus Mislaid Rock and Low Reef are underlain by the Upper Breccia Member

c. Offshore geology of the South Georgia microcontinent from geophysical work by Simpson and Griffiths (1982). Note the extension of the Cooper Bay Shear Zone across the whole block.

d. Palaeocurrent and palaeoslope directions in the Cumberland Bay Formation (unshaded; after Macdonald and Tanner, 1983).

e. Comparison of magnetic profiles for southern Chile and South Georgia (simplified from Simpson and Griffiths, 1982, Fig. 20.4). Top profile, Chile, from Project Magnet Flight 722 southward from Punta Arenas (P on Fig. 4a); bottom profile along NE-SW line shown on Fig. 7c. The magnetic highs correspond to the. Patagonian arc/batholith and Annenkov arc respectively

Much of South Georgia is underlain by a thick succession of Lower Cretaceous volcaniclastic sandstone and mudstone (*Cumberland Bay Formation*) originally deposited by deep-marine turbidity currents. The strata were then asymmetrically folded and thrust towards the north-east to structurally overlie a more intensely deformed and cleaved succession of quartzose turbiditic sandstones (*Sandebugten Formation*), which may nevertheless be of a similar Cretaceous age. Thin doleritic sheets occur locally, intruded prior to thrusting, and additional, post-thrusting deformation episodes have affected both lithostratigraphical divisions; much of the Sandebugten Formation is affected by a pervasive crenulation cleavage. A major shear zone (Cooper Bay Shear Zone) separates the Cumberland Bay and Sandebugten formations plus broadly similar (but more deformed) sedimentary rocks cut by multiple dolerite sheets (*Cooper Bay Formation*) from a very different lithological assemblage that forms the south-east end of the island. This SE area comprises three units:

1. gneiss and metasedimentary rocks intruded by gabbro and possibly Triassic granite, with mafic dykes locally forming up to 80 percent of the outcrop (*Drygalski Fjord Complex*)

2. Upper Jurassic gabbros, sheeted dolerite dykes, basaltic lavas and volcaniclastic rocks (*Larsen Harbour Complex*), and

3. Lower Cretaceous tuff and volcaniclastic breccias (*Annenkov Island and Ducloz Head formations*).

Overall, the geology of South Georgia derives from being located close to a continental margin characterized by opening of a marginal basin system, formation of an active volcanic arc, with the subsequent closure of the basin and deformation of its sedimentary fill during inversion. This history is remarkably similar to that of the missing hinterland of the southernmost Andes as pointed out by Dalziel et al. (1975) and documented in more detail by Macdonald et al. (1987). This theme is further developed in the following sections.

3. Marginal basin framework

The sedimentary basin strata forming most of South Georgia are floored partly by pre-Jurassic continental basement and partly by mafic rocks formed during the Late Jurassic rifting event that resulted in formation of the marginal basin. The continental basement is named the Drygalski Fjord Complex and the mafic rocks the Larsen Harbour Complex.

3.1 Drygalski Fjord Complex

The Drygalski Fjord Complex (DFC) is an area of igneous and metamorphic rocks comprising deformed paragneisses and metasedimentary sequences intruded by large gabbro and small granitic plutons and migmatites cut by mafic dyke suites (Fig.8a). It has been interpreted as a fragment of pre-Jurassic continental crust (basement) that was intruded by a wide variety of mafic and felsic intrusions prior to and during the early stages of formation of the basin system of South Georgia (Storey, 1983). The DFC forms the southeastern end of the central mountain range of southern South Georgia; it is separated from the Larsen Harbour Complex (LHC) by a prominent topographic low that includes Drygalski Fjord and is interpreted as a fault zone (Fig. 7). On its north eastern side, the DFC is separated from the deformed basin-fill sedimentary rocks of the Cumberland Bay and Cooper Bay formations by a major shear zone, the Cooper Bay Shear Zone (*sensu* Curtis et al., 2010; formerly the mylonite zone of Storey et al., 1977, and the Cooper Bay Dislocation Zone of Tanner, 1982a and Macdonald et al., 1987).

Although both igneous and metamorphic rocks were known from this area since Filchner's German South Polar Expedition (Heim 1912), it was Trendall (1959) who first mapped the area and referred to it as the South-Eastern Igneous Complex. More extensive mapping was carried out in 1974-76 by the British Antarctic Survey (Bell et al., 1977; Storey et al., 1977; Tanner and Rex, 1979; Storey and Mair, 1982; Storey, 1983) and the complex was renamed the Drygalski Fjord Complex (Storey, 1983). The DFC extends 65 km offshore to the ESE to include the Clerke Rocks (Fig. 7c; Tanner, 1982a).

The age of the various lithologies is not well constrained. The sedimentary and metasedimentary rocks predate all the intrusive rocks and are intruded by granites and gabbros which have yielded early Jurassic Rb-Sr and K-Ar ages (Tanner and Rex, 1979). These are the oldest known rocks on South Georgia. K-Ar mineral ages for the migmatite granite and dykes fall within the 120-149 Ma age range.

Three spatially distinct sedimentary and metasedimentary formations (Salomon Glacier, Cooper Island and Novosilski Glacier formations) that form part of the pre-Jurassic basement of South Georgia have been identified within the DFC (Storey, 1983).

Metasedimentary rocks of the Salomon Glacier Formation (Storey,1983) range from fine-grained biotite paragneiss showing relict sedimentary structures (cross lamination, slump folds, disturbed bedding and load and flame structures) to intensely deformed banded gneisses and layered migmatites regionally metamorphosed up to amphibolite facies. The Cooper Island Formation (Storey, 1983) is a flat-lying inverted sequence of laminated mudstone, siltstone and massive sandstone. Convolute lamination, cross lamination, graded beds, and bottom structures are well preserved in the laminated units. Rare plant stem impressions, trace fossils and irregular burrows are preserved in mudstone horizons. The Novosilski Glacier Formation (Storey, 1983) consists of epiclastic and volcaniclastic sandstones with interlaminated mudstone and shale, and porphyritic felsites.

Sandstones of the Novosilski Glacier and Cooper Island formations are moderately to poorly sorted immature arkosic greywacke and arkosic arenite derived from a granitic, metamorphic and felsic volcanic terrane.

Early Jurassic magmatism (Storey, 1983) led to emplacement of a sub-alkaline tholeiitic magma. Differentiation of this magma produced a wide variety of layered gabbros and intermediate and granitic rocks; continued emplacement of mafic magma formed a mafic dyke suite cutting the gabbroic and granitic rocks. A migmatitic aureole surrounds the plutonic rocks as a result of emplacement of the large volumes of mafic magma into the metasedimentary rocks. The igneous rocks are injected and net-veined by anatectic granitic veins and dykes and by emplacement of intrusive breccias.

The gabbroic rocks vary in composition from feldspathic leucogabbro to ultramafic and are the most abundant igneous rocks of the complex. They range from small intrusive bodies to large irregular plutons. Larger plutons commonly exhibit cumulate layering and variable composition (including troctolite, olivine-gabbro, hornblende pyroxene gabbro, two-pyroxene gabbro, and norite). Ultramafic inclusions up to 40 m across of lherzolite, peridotite and picrite are present within a massive leucogabbro. Dioritic granitic rocks form a small proportion of the igneous rocks within the DFC. There are two large granite intrusions, the Trendall Crag granite and the Cooper

Island granophyre (Storey, 1983). The Trendall Crag granite is a composite unit of tonalite, granodiorite and granite that intrudes and assimilates earlier gabbro plutons. The Cooper Island granophyre is a pale grey leucocratic trondhjemite and tonalite with a characteristic granophyre texture that intrudes and assimilates the Cooper Island Formation. In addition, trondhjemite, tonalite and alkali granite dykes intrude older igneous and basement rocks throughout the complex.

Numerous mafic dykes (Storey, 1983) were intruded during the history of the complex. Most were intruded prior to and were affected by the migmatization event. The concentration of mafic dykes increases towards the western margin of the complex bordering the Larsen Harbour Complex where they form up to 80 percent of the DFC. The dykes occur both singly and in multiple units up to 10 metres wide. The dykes form a bimodal sequence with north-west and north-east orientations.

The mafic dykes and gabbros have similar geochemistry (Storey, 1983) and are derived from the same magmatic source, a sub-alkaline tholeiitic magma. The mafic dykes are olivine- and quartz-tholeiites and the gabbros are sub-alkaline olivine tholeiites. The granitic rocks are variable in origin. Some are cogenetic with the mafic magmas and formed by differentiation of the tholeiitic magma, whereas others formed by partial melting of the metasedimentary rocks during emplacement of the mafic rocks; a small proportion are calc-alkaline and may be related to subduction processes.

A heterogeneous migmatite complex cut by aplite, pegmatite and late mafic dykes surrounds the gabbro and granitic plutons. The older rocks are injected and net-veined by a fine grained granitic neosome with angular and partially assimilated gabbroic and mafic dyke enclaves enclosed within a granitic matrix.

In summary, the DFC represents pre-Jurassic continental basement that was rifted and intruded by a mafic igneous complex during the initial stages of formation of what became a marginal back arc basin in the Jurassic and Early Cretaceous (Storey et al., 1977; Storey, 1983).

3.2 Larsen Harbour Complex

The Larsen Harbour Complex (LHC) covers an area of about 120 km^2 in the southwestern part of South Georgia. Volcanic rocks were first recorded in this area by Heim (1912). Tyrrell (1915, 1918) classified various igneous rocks acquired by Ferguson in 1912 (Ferguson et al., 1914) as dolerite and trachyte. Wordie (1921) visited the Larsen Harbour area and recorded two distinct sets of vertical and inclined doleritic dykes. Various igneous rocks were subsequently described by Douglas (1930), Holtedahl (1929), Barth and Holmsen (1939), and Trendall (1953, 1959). Detailed geological mapping was later carried out by the British Antarctic Survey (Bell et al., 1977; Storey et al., 1977; Mair, 1983, 1987). The igneous rocks in this area were named the Larsen Harbour Complex by Macdonald et al. (1987).

The LHC comprises a stratiform sequence, up to 2 km thick, of pillowed and massive lavas and breccias with interbedded chert and volcanogenic sedimentary rocks, cut by mafic and felsic dike swarms and small, mainly mafic intrusions (Mair, 1987). At Smaaland Cove in the southern part of the complex, a small (6 km^2) composite pluton dated at 150 ± 1 Ma (U-Pb zircon from quartz-gabbro, Mukasa and Dalziel, 1996) intrudes the base of the lava sequence (Fig 8b). Five groups of basalts (Groups I to V) have been recognized in Smaaland Cove (Alabaster and Storey, 1990) based on lava stratigraphy and cross cutting dyke relationships. Groups I, II, and III are part of the early lava sequence and predate the 150 Ma intrusion. Chemically identical feeder dikes for these three groups cut a ca. 180 Ma granite pluton in the Drygalski Fjord Complex (Tanner and Rex, 1979). Group IV and V dikes cut the Smaaland Cove intrusion and fed lavas higher in the succession; they are thus younger than 150 Ma.

Petrographically each basalt group is aphyric (Alabaster and Storey, 1990), although some Group II lavas contain microphenocrysts of olivine (now pseudomorphs) and diopside (Mg#–82), and some Group IV lavas contain microphenocrysts of forsteritic olivine (Fo90) with 0.34 wt percent NiO, and diopside (Mg#–84). Initial Nd isotopic compositions suggest that the basalts are derived from two different sources. The older Groups I-III magmas were derived by varying degrees of partial melting and fractional crystallization from a large ion lithophile element (LILE)-enriched, low-eNd mantle source relative to normal (N)-type mid-ocean ridge basalt (MORB) during the early stages of continental lithosphere attenuation. Magmas produced during later stages of rifting (Groups IV and V) were, however, derived from a high-eNd asthenospheric mantle source similar to N-type MORB, unaffected by earlier LILE enrichment. The lack of a clear subduction related signature in the basalts led Alabaster and Storey (1990) to suggest that the Larsen Harbour Complex and, by implication, the Rocas Verdes did not form in a supra-subduction-zone setting, but instead along an oblique-slip margin akin to that of the Gulf of California. The relationship of the basin to subduction-related arc magmatism will be discussed in more detail below.

3.3 Comparison with southern South America

The two major Mesozoic igneous complexes, Sarmiento and Tortuga, of the Rocas Verdes belt described by Katz (1964) consist predominantly of mafic pillow lavas and breccias, sheeted dykes and gabbros that are interpreted as the upper parts of ophiolites formed along mid-ocean ridge type spreading centers (Dalziel et al., 1974a; Dalziel 1981; Stern and de Wit, 2003; Calderón et al., 2007).

3.3.1 Sarmiento Complex

The Sarmiento Complex (SC) is the northernmost ophiolitic complex in the Rocas Verdes belt (Fig. 4a). It has been studied in detail by Allen (1982) and by Calderón et al. (2007, 2013). Approximately 2 km of extrusive pillow basalts and breccias are underlain by a sheeted dyke complex which exhibits a gradation from screens of pillow lavas cut by dykes to 100 percent mafic dykes. A leucocratic granite is intruded at the base of the sheeted dyke complex with an abrupt margin similar to the one at Smaaland Cove on South Georgia (Fig 8b). Gabbro and ferrogabbro cut by mafic dykes form the deepest part of the ophiolite sequence. The basalts of the Sarmiento Complex exhibit a tholeiitic differentiation trend, and the least differentiated basalts are olivine normative. (La/Yb) ratios of both basalts and more differentiated rocks are >1 (Saunders et al., 1979; Stern 1980), similar to mafic magmas produced during the early stages (Basalt Groups I to III) of evolution of the Larsen Harbour Complex on South Georgia (Alabaster and Storey, 1990; Storey and Alabaster, 1991). Zircon U/Pb ages of 139+/-2 Ma for silicic rocks within the Sarmiento complex, indicate a slightly younger age than the LHC (Stern et al. 1992), but more recent studies have indicated the mafic magmatism started as early as 150 Ma (Calderón et al., 2007)

3.3.2 Tortuga Complex

The Tortuga Complex (TC), on Isla Navarino (Fig. 4a), has a similar overall stratigraphy to the SC with the following exceptions; neither silicic plagiogranites nor trondhjemites have been observed, and the lower contact of the sheeted dyke complex is gradational. Dykes grade downwards into medium grained diabase cut by later dykes and finally cumulate plagioclase rich and olivine rich gabbros. The igneous rocks of the Tortuga Complex exhibit a more restricted chemical range than those of the Sarmiento Complex with no ferrobasalts or silicic plagiogranites. The tholeiitic basalts of the Tortuga Complex have (La/Yb) N < 1, similar to MORB, and they must have formed from melting of a MORB type asthenospheric mantle depleted in incompatible large ion lithophile elements (LILE), including LREE (Stern 1980) similar to the younger Group IV to V basalts in the Larsen Harbour Complex. Dykes also have similar Nd isotopic composition to MORB, but significantly higher measured Sr isotopic ratios because of the extensive interaction and exchange of Sr with sea water that they experienced during the ocean-floor metamorphism of the Tortuga Complex. In addition, the TC contains high-MgO (>10 wt percent MgO) dykes which cut the deeper gabbro and massive diabase level below the sheeted dyke complex.

4. Volcanic arc – marginal basin system

4.1 Volcanic arc rocks

All of the geological units ascribed to a volcanic arc are found in three scattered island groups SW of the central part of South Georgia: Annenkov Island and its offshore islets; the three islands of the Hauge Reef; and the Pickersgill Islands (Fig. 7). The rocks here are geologically distinct from adjacent parts of the main island and are separated from it by a continuation of the Cooper Bay Shear Zone (Curtis et al., 2010). This structure has been identified by geophysical mapping offshore, traversing the whole southwestern margin of the South Georgia continental block (Simpson and Griffiths, 1982; Fig. 7c). The shear zone is assumed to separate the basin-fill units from arc and basin-floor rocks to the SW (Macdonald et al., 1987).

Annenkov Island was briefly examined by Trendall (1959), but it was T.H. Pettigrew who carried out the first detailed work in 1974 (Pettigrew, 1981). The other islands were investigated in the 1975-76 season in a series of ship-supported landings (Tanner et al. 1981). Later work, such as the dating by Carter et al. (2014) relied on samples collected by these earlier investigations.

Figure 8 about here

8. Aspects of the geology of South Georgia

a. Drygalski Fjord Complex: dykes cutting layered gabbro on the NE coast of Drygalski Fjord. Image by Bryan Storey.

b. Larsen Harbour Complex: pale-coloured plagiogranite intruding basaltic pillow lavas and dykes at Smaaland Cove. Image by Bryan Storey.

c. Larsen Harbour Complex: pillow lavas on the Hauge Reef. Image by Geoff Tanner.

d. Cumberland Bay Formation: turbidite beds at Gold Harbour showing two of the facies defined by Macdonald (1986); transitional Facies at the base of the cliff passing upwards into the Sandstone Facies. Image by Tom Sharpe.

e. Sandebugten Formation: D2 deformation at Moltke Harbour. Image by Philip Stone.

f. Cumberland Bay Formation: large-scale D1 asymmetric folding at Stromness Bay. Image by Kim Crosbie.

4.1.1 Field relationships and age

The pillow lavas and interbedded tuffs of the Larsen Harbour Complex have been interpreted as equivalent to the floor of the marginal basin below the Cumberland Bay and Sandebugten formations (Tanner et al., 1981; Mair, 1983, 1987).

The outcrop of the Larsen Harbour Complex is mostly on the main island but also, importantly, on Pillow Rock, the easternmost island of Hauge Reef (Fig. 7b, 8c), where the uppermost part of the complex is exposed. The pillow lavas have an overall westward dip which is near-flat in places in the highest part of the stratigraphy. Upper parts of the complex exposed at Leon Head

(Mair, 1987) and Pillow Rock (Tanner et al., 1981) contain significant units of interbedded andesitic tuff which suggest a gradational boundary with the overlying tuffaceous rocks (Tanner et al., 1981). Although this apparently gradational boundary is only partially exposed, it is the only stratigraphic boundary between sedimentary units exposed anywhere on South Georgia.

The overlying rocks are exclusively andesitic in composition: crystal-lithic tuffs and rarer vitric tuffs; andesitic lithic greywackes with subordinate feldspathic wackes; and breccias of andesite with some dacite and microdiorite clasts. They were originally regarded as a facies variation of the Cumberland Bay Formation on the basis of their comparable composition (Trendall, 1959). However, since they differed from the Cumberland Bay turbidites in field appearance and sedimentology, Pettigrew (1975, 1981) assigned them to the Annenkov Island Formation. He recognised two members: a Lower Tuff Member and an Upper Breccia Member, which has a thick basal sandstone unit. The lower member crops out on the middle and western islands of the Hauge Reef and in the eastern part of Annenkov Island. The upper member forms all of the rest of Annenkov Island plus Low Reef (SE of Annenkov) and Mislaid Rock (to the west). Pettigrew (1981) thought that the two members were conformable, but Storey and Macdonald (1984) suggested that their dips were discordant and that the upper member could be erosive into the lower unit.

The Lower Tuff Member is fossiliferous, yielding a fauna of ammonites and bivalves (Wilckens, 1932, 1937; Trendall 1959; Thomson et al., 1982). Although fossils are relatively abundant, they seem to represent endemic species and it is not possible to ascribe a date more precise than Early Cretaceous (Thomson et al., 1982).

4.1.2 Sedimentology

The Lower Tuff Member consists of interbedded mudstone and fine tuff. Mudstones are commonly tuffaceous, but a significant minority are radiolarian-bearing; in places there are lenses of radiolarite (Macdonald et al., 1987). The percentage of tuff seems to decline upwards, from about 70 percent to 10 percent near the top (Pettigrew, 1981; Storey and Macdonald, 1984). Tuff beds are generally graded and could be airfall deposits, however, many exhibit sharp bases with sole marks, grading, parallel lamination, and cross lamination suggesting Bouma sequences indicative of deposition from turbidity currents (Pettigrew 1981; Storey and Macdonald, 1984). There are also slump folds, synsedimentary faults and sediment injection structures, all indicating deposition on a slope; the presence of abundant radiolarians and the lack of current-generated structures points to deposition in relatively deep water (Storey and Macdonald, 1984).

The Upper Breccia Member comprises two facies associations. The Basal Sandstone is 91 m thick and consists of beds of coarse volcaniclastic sandstone, up to 2 m thick with sharp, flat bases and tops. Beds tend to be amalgamated or separated by thin mudstones (less than 10 cm thick). Most beds are structureless; a few exhibit coarse-tail grading and faint parallel lamination towards the top. This unit was probably deposited by high-density turbidity currents (*sensu* Lowe, 1982). The bulk of the upper member is formed of matrix- and clast-supported andesitic breccia with clasts up to boulder size. These are crudely bedded on a 1-5 m scale and are either amalgamated or contain lenses of very coarse sandstone (Tanner et al., 1981). Deposition was probably from debris flows (Pettigrew, 1981; Storey and Macdonald, 1984).

The overall sedimentary succession seems to be clearly related to marginal parts of a volcanic arc. Pettigrew (1981) suggested that the coarsening-upward motif was due to growth of the flanks of an andesite volcano. Storey and Macdonald (1984) favoured a deep shelf setting, cut by a large submarine canyon feeding volcanic material into deeper water (Fig. 7d).

4.1.3 Igneous intrusions

A wide variety of igneous intrusions cut the Annenkov Island Formation on Annenkov Island and its surrounding reefs; larger igneous bodies form the Pickersgill Islands (Macdonald et al., 1987). The earliest intrusions are basic sheets intruded into wet sediments, seen on the islands of the Hauge Reef. The volcaniclastic rocks are further cut by irregular stocks, sheets, sills and dykes of hornblende andesite with later biotite andesites. The Pickersgill Islands are composed of hornblende microdiorite and Tanner Island (largest island of the group) is formed of a composite quartz microdiorite intrusion which seems to have an earlier basic phase (Tanner et al., 1981).

Storey and Tanner (1982) analysed both igneous and sedimentary rocks from this area and demonstrated that all the rocks have calc-alkaline characteristics, suggesting a common island arc origin for all the igneous and volcanic rocks except the early basic sheets which have more in common with the Larsen Harbour Complex.

Dating of the igneous rocks using a variety of methods suggested Late Cretaceous ages (81–103 Ma: Thomson et al., 1982). This view is supported by more recent work (Carter et al., 2014) which yielded apatite fission track ages of 79-88 Ma for the same rocks and suggested that emplacement of the Tanner Island granite occurred in the range 85-100 Ma.

4.2 Marginal basin fill

Most of the land area of South Georgia is underlain by thick sequences of clastic sedimentary rocks, clearly differentiating the island from the ‘typical’ oceanic islands such as Ascension and Tristan da Cunha, which have a volcanic origin, with fossiliferous limestones as the main sedimentary rock (Darwin, 1844). This made South Georgia a scientific puzzle from the earliest days of exploration (e.g., Gregory, 1929). The evidence had accumulated through the early

20th century, exemplified by the series of papers written by G.W. Tyrrell based on samples collected by whalers or by expeditions, emphasizing the sedimentary nature of the Scotia Arc in general and South Georgia in particular (Tyrrell, 1915, 1916, 1918, 1930, 1945).

The first systematic surveys of South Georgia identified two principal lithofacies in the South Georgia sedimentary succession (Trendall, 1953, 1959): Cumberland Bay type greywackes rich in detrital andesitic material, cropping out across most of South Georgia, which contrasted with Sandebugten type quartz-rich greywackes restricted to areas around Cumberland East Bay, across Barff Peninsula, and in a coastal strip SE to Gold Harbour (Fig. 7). Trendall used informal stratigraphic nomenclature but the two types were formalized as formations by Stone (1980). Stone also recognized a rock unit of mixed provenance, restricted to the extreme NE of South Georgia, which he defined as the Barff Point Member of the Cumberland Bay Formation.

4.2.1 Age

The Cumberland Bay Formation is sparsely fossiliferous, with a shelly fauna of the bivalve *Aucellina* and a few fragments of large heteromorph ammonites, indicative of an Aptian–Albian (Early Cretaceous) age (Thomson et al., 1982). Belemnite fragments suggesting pre-Aptian deposition (Stone and Willey, 1973) were discounted by Doyle (1985) as indicating only a general Mesozoic age. Fossil wood material is relatively common across the whole island, being found widely in the Cumberland Bay Formation, including the Barff Point Member, and more sporadically in the Sandebugten Formation. Jefferson and Macdonald (1981) summarized the occurrences of this poorly preserved material and concluded that all specimens were indicative only of a Jurassic-Cretaceous age. There are abundant trace fossils which have no age significance, but Macdonald (1982) interpreted the ichnofacies and the sparse shelly fauna as indicators that the Cumberland Bay Formation was deposited in relatively deep water with high environmental stress, possibly partial anoxia within a restricted basin.

Thomson *et al*. (1982) dated 12 samples from the Cumberland Bay Formation using K-Ar whole-rock analysis; their results were in the range 138–151 Ma (Valanginian-Ypresian; Ogg et al., 2016). Older data may indicate provenance ages, while they interpreted a tight grouping of 93–84 Ma (Turonian-Santonian; Ogg et al., 2016) as uplift and cooling ages. Carter et al. (2014) used apatite thermochronometry to show a significant cooling event on the SW coast starting at 90 Ma. Taken together, these data show that the end of sedimentation of the Cumberland Bay Formation was before the start of the Turonian stage at 98.9 Ma; since there is no evidence of Cenomanian deposition, a depositional age range of Aptian-Albian (126-100 Ma; Ogg et al., 2016) is likely.

No fossil evidence indicative of age has been recovered from the Sandebugten Formation. Nevertheless, it is thought to be coeval with at least part of the Cumberland Bay Formation (Early Cretaceous; Thomson et al. 1982). The justification for that correlation arises from relationships at the northern end of the Barff Peninsula where an outlying klippe of the thrust sheet that rests on the Sandebugten Formation has been preserved. Most of the greywackes within this structural outlier have similar composition to those forming the Cumberland Bay Formation, but a significant proportion are of a quartz-rich composition characteristic of the Sandebugten Formation. This unit of interbedded greywacke types was defined by Stone (1980) as the Barff Point Member of the Cumberland Bay Formation. Detrital zircon data from a Sandebugten Formation greywacke reported by Carter et al. (2014) showed marked population peaks at about 275 Ma 171 Ma and 109 Ma. The youngest of these peaks is late Early Cretaceous age (mid–Albian: Ogg et al., 2016), which suggests that the Sandebugten Formation corresponds to the youngest part of the Aptian-Albian Cumberland Bay Formation. Previously, it had been assumed that the Sandebugten Formation was amongst the older parts of the succession (Winn, 1978; Stone, 1980) or even that it conformably underlay the Cumberland Bay Formation (Trendall, 1953; Turnbull and Craw, 1988)

4.2.2 Stratigraphy and sedimentology

Sandebugten Formation

The Sandebugten Formation crops out along the NE coast of South Georgia, extending from Greene Peninsula and Barff Peninsula south to Gold Harbour. Its component strata have experienced polyphase deformation, no base is seen, and the upper limit of the formation is a thrust contact with the structurally superincumbent Cumberland Bay Formation (Dalziel et al., 1975, Fig. 3 therein). Farther south, between Wirik Bay and Cooper Bay, an isolated outcrop of highly deformed metasedimentary rocks (with abundant meta-basaltic intercalations) has been defined as a separate Cooper Bay Formation but has some compositional similarity to the Sandebugten Formation and may be a variation of it (Stone, 1982; Storey, 1983).

The type section of the Sandebugten Formation lies 0.5 km NW of Sandebugten (Stone 1980). The formation mostly comprises a thick succession of alternating greywacke and mudstone, with sporadic granulestone interbeds; it has abundant examples of sedimentary structures characteristic of deposition from turbidity currents (Stone, 1980). Chert and marble bands have also been reported locally (Craw and Turnbull, 1986). The formation is generally thinly bedded but beds up to 1.45 m thick occur locally.

The clast population of the greywackes is dominated by quartz, commonly polycrystalline, and with some clasts derived from quartzose sedimentary rocks; feldspar and felsitic lithic grains are also abundant with accessory grains of trachyte and granite (Winn, 1978; Stone, 1980).

Palaeocurrent analyses from the sedimentary structures of the Sandebugten Formation have given mixed results. Data derived from cross-lamination seemed to favour a current flow towards the SW (Trendall, 1959; Dalziel et al., 1975) whilst small numbers of sole marks led to contradictory conclusions of either flow towards the SW (Stone 1980) or the NNE (Tanner, 1982a). In view of the structural complexity of the Sandebugten Formation, all these data should be treated with caution (Dott, 1974; Macdonald et al., 1987). although it does seem clear that they show a marked difference to the current patterns derived from the time-equivalent Cumberland Bay Formation (see discussion in next section).

The Early Cretaceous continental reconstruction by Dalziel et al. (1975) placed the South Georgia depositional basin adjacent to the termination of the Patagonian Andes in a position that broadly facilitated derivation of eroded silicic volcanic material (Tobífera/Lemaire Formation) from the north or NE. In terms of that reconstruction, Macdonald et al. (1987) commented that the petrographic evidence for a northerly or north-easterly source was much stronger than any of the palaeocurrent data.

Cumberland Bay Formation

Neither base nor top of the Cumberland Bay Formation is exposed. It lies structurally above the Sandebugten Formation, separated from it by a thrust plane which is exposed at Dartmouth Point (dipping SW) and on Barff Point (dipping NE). The same thrust is inferred to separate the Cumberland Bay Formation from the Cooper Bay Formation along the line of the Twitcher Glacier, but it is not exposed (Macdonald et al., 1987). The CBF is separated from the Novosilski Glacier, Ducloz Head and Annenkov Island formations and the Larsen Harbour Complex by the Cooper Bay Shear Zone (Curtis et al., 2010). Tanner and Macdonald (1982) estimated the thickness of the unit at 5-8 km.

The Cumberland Bay Formation is systematically variable across the island, so two type sections were established by Macdonald et al. (1987): at Larvik on the SW coast for the coarser grained deposits and at Macdonald Cove on the north coast for the finer deposits. The rocks are clearly turbidites and four lithofacies associations are recognised: massive, sandstone, transitional, and shale (Macdonald, 1986; Fig. 8d). The massive facies is almost entirely sandstone with most beds greater than 2 m thick, reaching 6 m in places; beds are coarse-very coarse sandstone and ungraded, or with crude coarse-tail grading. The sandstone lithofacies is more than 67 percent sandstone, with bed thickness varying from 0.25-2 m; beds display grading and have generally well-developed T_a, T_b and T_c divisions (Lowe, 1982); sandstone beds can be amalgamated or are interbedded with mudstone and thin beds of fine sandstone and siltstone. The transitional facies is formed of interbedded mudstone and sandstone (50-67 percent); sandstone beds are typically 0.1-0.25 m thick and display the full range of turbidite structures. The shale facies comprises mudstone (more than 50 percent) with thin beds of very fine sandstone and siltstone, generally less than 0.1 m thick.

Palaeocurrents have been measured from sole marks (tool and flute casts; Dott, 1974; Macdonald and Tanner, 1983; Craw and Turnbull, 1986), cross lamination in the T_c division (Trendall, 1953, 1959), and oriented wood (Macdonald and Jefferson 1985). Macdonald and Tanner also analysed syn-sedimentary slump folds to give an independent palaeoslope direction. Results consistently indicate dominant transport towards the NNW, with a subsidiary NE-directed component in coarser beds along the SW coast of the island; Macdonald and Tanner (1983) interpreted these results to show dominant ESE-NNW transport, with local input from a basin margin to the SE, now represented by the Cooper Bay Shear Zone.

Macdonald (1986) put the palaeocurrent, palaeoslope, and sedimentological data together to show strong proximal to distal sedimentological changes. The two finer–grained lithofacies associations are constant in character across the CBF outcrop, although thicker on the SW coast. In the two coarse-grained associations there is a change in character WNW, down the palaeocurrent direction, shown by decreasing sandstone bed thickness and percentage amalgamation, and also a marked proportional increase in T_b and T_c divisions within the bed downcurrent. This suggests that the system was aggradational rather than progradational: tectonic control of the basin margins prevented major migration of the depositional system, and most areas remained in the same position relative to source through time. This interpretation was supported by the observations of Stone (1980) and Craw and Turnbull (1986) who showed that in its southernmost outcrop the Cumberland Bay Formation was coarser, reaching pebble conglomerate grade, than in the areas described by Macdonald (1986).

4.2.3 Provenance

Once the clastic nature of the South Georgia sedimentary succession had been established, the island's isolated position posed the problem of a source for the copious amount of sediment required. The early debate around this issue has been summarised by Gregory (1929) who, like many of his contemporaries, thought the South Georgia sedimentary basin to have been peripheral to an Atlantic continent that had subsequently subsided beneath the ocean. The eventual acceptance of continental drift and the succeeding realisation of plate tectonics has allowed alternative solutions linked with the overall development of the Scotia Arc. In that context, data from a series of petrographical studies (for a summary and bibliography, see Stone, 2015a) demonstrated that the provenance of the Sandebugten Formation greywackes was an area of continental basement, silicic volcanic rocks, and quartz-arenites. Dalziel et al. (1975) and Winn (1978) made strong cases for this being, in geological terms, the Jurassic silicic volcanic and siliciclastic sedimentary rocks of the Tobífera/Lemaire Formation of South America, together with the underlying metamorphic rocks of the south Patagonian basement. Detrital zircon data (Carter et al., 2014; Barbeau et al., 2009a, b) provide a good match for that assemblage, notably the 171 Ma age equivalent of the Jurassic Tobífera/Lemaire Formation, with the addition of the younger 109 Ma zircons from a potential source in the South Patagonian Batholith.

The Cumberland Bay Formation is more straightforward. All the sand grade and coarser material is volcanic in origin: the arenites are lithic volcaniclastic greywackes (*sensu* Pettijohn et al., 1972). Sorting is poor and clasts are varied. Feldspars are dominantly plagioclase (Winn, 1978) and volcanic rock fragments include: andesite, amygdaloidal andesite, quartz-andesite, dacite, felsite, and rhyolite, with 95 percent of the clasts being of intermediate composition (Tanner et al., 1981). The interpretation of derivation from a volcanic arc, first suggested by Trendall 1959, has been supported by every subsequent study, notably Dalziel et al. (1974a, 1975); Dott et al. (1982); Winn (1978); Tanner et al. (1981); and Storey and Macdonald (1984). It is interesting to note that every sandstone sample contains a variety of rock types, suggesting mixing in a shelf area rather than direct volcanic input.

4.2.4 Structure and metamorphism

Deformation of both formations was polyphase, with a southward increase in complexity and intensity in the Sandebugten Formation and a NE–ward increase in the Cumberland Bay Formation.

Stone (1980) recognised three folding episodes in the southern part of the Sandebugten Formation outcrop but the second and third of these are not seen in the northern part of the outcrop, in the Barff Peninsula. There, and on part of the neighbouring Greene Peninsula, tight chevron folds with a NW-SE axial trend have wavelengths from tens of metres up to a few hundred metres. The fold attitude varies, with facing direction to the SW in the west progressively rotating through upright so that facing direction is to the NE in the east. As a result, the overall structure resembles an anticlinorium. These FI folds have a penetrative, axial planar slaty cleavage. In the southern part of the Barff Peninsula, and increasingly southwards into the Royal Bay area, The F1 folds are refolded about a sub-horizontal axial plane (hinges are variable but broadly NW-SE) with the development of a crenulation cleavage that becomes progressively more dominant southward (Fig. 8e). That crenulation cleavage, S2, is itself refolded into large scale open structures with the development of a second crenulation cleavage, S3, which is sub-vertical and trends NNE-SSW. The F3/S3 effects are most pronounced on the south side of Royal Bay. The southward increase in the intensity of deformation continues into the possibly correlative Cooper Bay Formation, where schistose and mylonitic fabrics developed within a series of three principal fold episodes that were approximately coaxial on a NW-SE structural trend; a range of later, minor structures are locally superimposed (Stone, 1982; Storey, 1983). Fabrics merge into the mylonitic Cooper Bay Shear Zone (which forms the SW boundary of the formation's outcrop) with which the structural history is most probably closely related (Curtis et al., 2010).

The Cumberland Bay Formation is deformed across its whole outcrop by large F1 chevron folds; amplitude and wavelength are on a scale of hundreds of metres (Macdonald et al., 1987). These vary from upright to overturned close folds on the SW coast to tight or isoclinal overturned or recumbent folds in the NE (Fig. 8f). There is a penetrative S1 axial planar fabric which gives a strong L1 intersection lineation. Fold axes are parallel to the length of the island and folds face and verge NE; plunge is variable, but dominantly towards the NNW (Tanner and Macdonald, 1982). The F1 deformation occurred by flexural slip (Tanner, 1989). There are two later phases of folding, both restricted to the NE coast. F2 structures are coaxial with F1, and present as upright open folds with wavelengths of metres to tens of metres. Much larger F2 structures are apparent from mapping their effect on S1, which is flattened to near-horizontal in areas between Fortuna Bay and Busen Point (Macdonald et al*.,* 1987). A vertical axial-planar S2 crenulation cleavage is locally developed. F3 structures are mostly centimetre-scale kink bands and rarer box folds oriented N-S and plunging north (Macdonald et al., 1987). Tanner and Macdonald (1982) suggested that all the deformation could be ascribed to a simple shear couple related to SW-directed underthrusting of the basin floor. More recent work by Curtis et al. (2010) has found that there is also a component of vertical-axis sinistral shear affecting the basin fill, which they interpreted as the result of sinistral transpression and kinematic partitioning.

In association with the polyphase deformation, both formations were buried to depths that allowed metamorphism within the prehnite-pumpellyite facies (Tyrrell, 1930; Trendall, 1959; Stone, 1980). This is roughly 200°C and 0.4 GPa (Bucher, 2005), indicating burial to about 10–15 km.

The age of metamorphism was interpreted by Thomson et al. (1982), using K-Ar dating of slate samples from the Sandebugten Formation, as about 82-84 Ma, with post-orogenic uplift continuing until at least 50-60 Ma. More modern data provided by Carter et al., (2014) was derived from apatite and zircon fission track and apatite (U-Th/He) thermochronometry. This showed that metamorphism was ended at about 45 Ma by rapid uplift to relatively shallow crustal levels (<2 km) before Miocene reburial was followed by inversion at about 7-10 Ma. The Cumberland Bay Formation apatite results are less clear-cut than those from the Sandebugten Formation, but seem to show similar reburial in Miocene time with inversion at 7-10 Ma.

4.3 Comparison with southern South America

Close comparisons have been established in age, facies, palaeogeography and provenance between the main arc and basin units of South Georgia and South America, where similar rocks occur, both in outcrop and in the subsurface (Dalziel et al., 1975; Dalziel, 1981; Suarez and

Pettigrew, 1976; Dott et al., 1977; Winn, 1978; Olivero and Mantinioni, 2001). These correlations are discussed here, working from the Pacific margin towards the continent; the correlations are summarised in Fig. 9.

Figure 9 about here
9. Stratigraphic correlation diagram of South Georgia and southernmost South America. The colours correspond to those in the geologic maps of South Georgia (Figs. 7a, b).

Q–quartzose facies of Yahgan Formation; S Fm–Springhill Formation

4.3.1 Volcanic arc

The arc rocks of Tierra del Fuego are dominated by the Patagonian Batholith. This large, complex granitoid body was emplaced in phases from Late Jurassic until Early Cretaceous time (157–137 Ma; Hervé et al. 2007). It is coincident in time with both the silicic volcanics of the Tobífera Formation and the basic rocks of the Rocas Verdes system. No such large batholith occurs in outcrop on South Georgia, but marine geophysical investigations revealed the existence of fragments of a large magnetic batholith on the SW side of the South Georgia continental block (Simpson and Griffiths, 1982; Fig. 7c). Simpson and Griffiths (1982, Fig.20.4) noted the close similarity of magnetic anomaly profiles across the Andean Cordillera at ~70° W longitude and the South Georgia block (Figs. 7e).

On the northern Andean edges of the batholith lie the andesitic volcaniclastic rocks of the Hardy Formation (Fig. 4a), a 1300 m thick sequence of intermediate lavas, tuffs, tuffaceous sandstones, and volcanic breccias, interpreted as shallow marine deposits of dominantly Early Cretaceous age deposited on the shelf of a volcanic arc (Suarez and Pettigrew, 1976; Cunningham et al., 1991; Miller et al., 1994). To the north of the Hardy Formation outcrop is a facies variant named the Tekenika Beds, of uncertain age (Halpern and Rex, 1972; Dott et al., 1977), but most likely Early Cretaceous; the unit comprises coarse sandstone and conglomerate. These have been interpreted as post-orogenic, but Dott et al. (1977) and Winn (1978) interpreted the Tekenika Beds as pre-orogenic deep-water deposits, equivalent in age to both the Hardy and Yahgan formations. The mass-flow conglomerates of the Tekenika Beds closely resemble the Upper Breccia Member of the Annenkov Island Formation and might be more specifically extended to the debris flow breccias that form the Coastal Member of South Georgia's Ducloz Head Formation (Figure 9).

The Hardy Formation and the Tekenika beds encompass all of the facies of the Annenkov Island Formation and are of the same age. Some South American facies are not recognised on South Georgia; this can be ascribed to the much larger outcrop area exposed in Tierra del Fuego compared to the scattered islets around Annenkov Island.

4.3.2 Boundary between arc and basin

The rocks of the arc terrane on South Georgia are separated from the basin-fill by the major tectonic boundary of the Cooper Bay Shear Zone, which is suspected to be close in space and orientation to the original basin margin (Macdonald et al., 1987). No such boundary is found in Tierra del Fuego, Instead, the arc and basin rocks interfinger, allowing their stratigraphic age equivalence to be demonstrated (Olivero and Martinioni, 2001). Despite the lack of a major shear zone, the transition from arc to basin is relatively abrupt, over no more than a couple of kilometres (Miller et al., 1994; see also Olivero and Martinioni, 2001, Fig. 4).

4.3.3 Basin fill: Yahgan and Cumberland Bay formations

The association of the Lower Cretaceous, volcaniclastic turbidites of the Cumberland Bay Formation (South Georgia) with those of the Yahgan Formation (southern Andes) was recognised at an early stage in the geological study of the region and has been reinforced by successive subsequent investigations (e.g., Kranck, 1934; Katz and Watters, 1966; Winn, 1978). The volcaniclastic turbidites of the Yahgan Formation have been studied in some detail (Dott et al., 1977; Winn, 1978). Their age is well-established as Early Cretaceous (Olivero and Martinioni, 1996, 2001). They are pervasively deformed into north-facing and verging chevron folds with a penetrative slaty cleavage developed in finer-grained units and metamorphosed to prehnite-pumpellyite facies, reaching greenschist grade in places (Caminos et al., 1981; Olivero and Martinioni, 2001). They comprise abundant andesitic volcanic fragments (Winn, 1978; Andrews–Speed, 1980) and show three major facies associations: black mudstones with fine-grained turbidites and tuffs; classic sandstone turbidites; and massive to graded sandstones (Olivero and Martinioni, 1996). Winn (1978) documented complex palaeocurrent patterns with both margin-normal (north-directed) and margin-parallel (NW-directed) components.

All of these features are an exact match for the Cumberland Bay Formation of South Georgia: age; provenance; sedimentology; paleogeography; structural style and vergence; and meta-morphic grade (Olivero and Martinioni, 2001).

4.3.4 Basin fill: Beauvoir, Springhill and Sandebugten formations

In Argentina, the Yahgan Formation passes north into the dark mudstones of the Beauvoir Formation, of Early Cretaceous age; their outcrop extends east into northern Isla de los Estados (Fig. 4a) and they are interpreted as a slope deposits from the northern side of the marginal basin (Olivero and Martinioni, 2001). They are laterally equivalent to the Hito XIX and La Paciencia formations of Chile. This unit is deformed by open folding and thrust faulting. While these rocks do not, at first sight, resemble the Sandebugten Formation in either facies or structural style, no coarse-grained basin-centre equivalents are exposed on Tierra del Fuego, so direct comparisons are not available. However, Olivero and Martinioni (2001) make a correlation between the Beauvoir Formation, via the La Paciencia Formation to the Lower Inoceramus Formation in the subsurface. This unit interfingers with the Berriasian–Valanginian, shallow marine Springhill Formation of the Magallanes-Austral Basin, which unconformably overlies and is derived from the Middle Jurassic silicic volcanic rocks of the Tobífera Formation (Schwarz et al., 2011). Thus the Sandebugten Formation rocks, which have the same provenance, can be considered as a deep-water equivalent of the Springhill Formation.

5. Palinspastic restoration of the South Georgia microcontinent

As discussed in sections 3.3 and 4.3, there is a striking similarity between the geological components of South Georgia and those of the Southern Andes; the key stratigraphic correlations are summarized in Figure 9. The close comparison extends to the timing of tectonic events, notably the formation of the ophiolitic floor of the marginal basin, the onset of arc magmatism and of basin closure and inversion (Table 1). Conversely, the geological evidence demonstrates that the South Georgia microcontinent cannot have originated as part of the Falkland Plateau adjacent to the Maurice Ewing Bank, and hence southern Africa prior to opening of the South Atlantic, as advocated by Eagles (2010) and Eagles and Eisermann (2020). Sedimentary strata in that region have dominant Mesoproterozoic and Neoproterozoic zircon populations (Craddock and Thomas, 2011; Ramos et al., 2017; Craddock et al., 2019). These are absent in the strata of South Georgia and southernmost South America (Barbeau et al., 2009a, b; Carter et al., 2014).

Table 1 about here
Correlation of tectonic events in southernmost South America, the South Georgia microcontinent, and the South Atlantic/Weddell Sea region.

These regional patterns encouraged palinspastic restoration of South Georgia into continuity with the southernmost Andes. In the reconstructions of Dalziel et al. (1975) and Tanner (1982a) an additional benefit arose from the positioning of South Georgia to the south of Isla de los Estados and Burdwood Bank. These localities are underlain by the Jurassic, silicic volcanic rocks of the Tobífera/Lemaire Formation (Fig. 4a, b), which thence provided an appropriately positioned provenance for the Sandebugten Formation, one conspicuously absent in the present location of the microcontinent.

The recognition that igneous assemblages in both South Georgia and the southern Andes represented ophiolitic complexes—Larsen Harbour Complex (South Georgia: Storey et al., 1977; Storey and Mair, 1982), Sarmiento and Tortuga Complexes (southern Andes: Dalziel et al. 1974a, Stern and de Wit, 2003; Calderón et al., 2007, 2013)—completed the correlation of both areas as parts of the same Early Cretaceous marginal basin. The basin formed between a sliver of old

South American continental crust that became the foundation of a calc-alkaline volcanic arc and the main part of the South American continent from which the sliver had rifted. In terms of South Georgia's geology, the continental foundation to the arc is now represented by the Drygalski Fjord Complex, an arc-proximal shelf facies by the Annenkov Island Formation, the oceanic crust of the marginal basin by the Larsen Harbour Complex, and the turbiditic basin-fill sediment by the Cumberland Bay, Sandebugten and associated formations. Asymmetrical, continent-vergent, deformation then accompanied closure of the marginal basin in the mid- to Late Cretaceous, as the volcanic arc complex moved back towards the of the arc (Figs. 10, 11).

Figure 10 about here

10. Schematic reconstruction of the Rocas Verdes basin at its supposed maximum width in the Early Cretaceous, ca 110 Ma. The shape and dimensions of the basin are determined by the paleomagnetic data that indicate a linear north-south trending arc prior to post-Early Cretaceous 90 degrees counterclockwise rotation as determined in the arc/batholith of Tierra del Fuego (Poblete et al., 2016) and arc length implied by the palinspastic reconstruction discussed in the text. The outline of the South Georgia microcontinent shows where the Annenkov Island arc and Rocas Verdes basin ophiolitic basement and infill were located after basin closure as shown in Figure 12.

Coordinates are those of present-day South America

The uncertainties imposed on the early attempts at 'pre-drift' reconstructions of the Scotia Arc, for example, the proposals of Matthews (1959) and Hawkes (1962), are well exemplified by the fact that a geological description of *in situ* epidote-actinolite greenschist specimens from Shag Rocks was not available until published by Tanner (1982b). Prior to that, the only geological information available for the Shag Rocks continental block derived from similar schistose rocks dredged from the sea floor in 1930 and described by Tyrrell (1945). Confirmation of the continental lithological character of the Shag Rocks block added credibility to its positioning in the reconstructions of Dalziel et al. (1975) and Tanner (1982a).

This has been strengthened by recent detailed petrographic study of the Lapataia Formation at the eastern extremity of the Cordillera Darwin metamorphic basement complex on the north shore of the Beagle Channel at Bahia Lapataia (Cao et al., 2018; see Fig. 4 for location). It reveals the striking similarity of some components of this complex to the stilpnomelane–bearing greenschists forming Shag Rocks. As described by Tanner (1982a). The Lapataia schists were thrust northeastward over the Tobífera/Lemaire Formation during inversion of the Rocas Verdes Basin (Bruhn, 1979, Fig. 4; Cao et al., 2018, Fig. 10). Restoration of the south-dipping schists of the Shag Rocks Bank between the South Georgia microcontinent and Isla de los Estados and

Burdwood Bank (Figs. 11 and 12) as further discussed below, creates an exactly analogous structural setting, the southern limb of the large north-verging syncline in the Tobífera/Lemaire Formation on the island being overturned (Dalziel and Palmer, 1979, Figs. 10, 13). South-dipping mylonitic shear zones occur within the schists at Lapataia and the siliceous volcanics of the Tobífera/Lemaire Formation at both locations (Fig. 11).

Figure 11 about here

Comparative cross-sections through: top - the schists of the Lapataia Formation and Tobifera/Lemaire Formation in Tierra del Fuego (simplified from Cao et al., 2018, Fig. 10); and bottom–the Shag Rocks schists and Tobifera/Lemaire Formation and overlying sedimentary strata of Isla de Los Estados (Tanner, 1982a ; Dalziel and Palmer, 1979, Fig.13). See text for discussion.

The paucity of marine geophysical data had also disadvantaged the early discussions but was introduced into the debate by Barker and Griffiths (1972) who, following Hawkes (1962; Fig. 5b), proposed the post-Cretaceous fragmentation of the Andes and illustrated a tentative reconstruction that brought South Georgia to the south of the Burdwood Bank, thus matching the magnetic anomalies over the south Patagonian batholith and the southwestern margin of the South Georgia microcontinent, a match enhanced by Simpson and Griffiths (1982, Fig. 20.4; Fig. 7E). This correlation provided an independent basis for the subsequent, more detailed geological correlations and reconstructions that brought South Georgia into much the same position (e.g., Dalziel et al., 1975; Tanner 1982a). More recently Vérard et al., (2012); and Van de Lagemaat et al. (2021) have incorporated similar reconstructions in their plate models.

This palinspastic reconstruction of the South Georgia microcontinent adjacent to the Cape Horn escarpment requires some 1700 km of eastward translation relative to the South American continent to reach its present position. This must have taken place since the mid-Cretaceous deformation and inversion of the Rocas Verdes basin (Table 1). Apatite thermochronology records a cooling event in both Tierra del Fuego and the microcontinent at ca. 45-40 Ma (Carter et al., 2014). This implies that South Georgia remained connected to its parent continent until that time, but need not, however, require it to have remained in its original position immediately east of the escarpment

The elevated banks of the southern Scotia Sea: Terror Rise, Pirie Bank, Bruce Bank, and probably part of Discovery Bank (Fig. 3), appear to have rifted off the northeast-southwest-trending continental margin of South America that lies immediately to the east of Cape Horn (Civile et al., 2012). They are continental in character (Vuan et al. 2005; Galindo-Zaldívar et al., 2014). The intervening basins may have started rifting at the same time as the continental rifts in Tierra del

Fuego. Dove Basin, between Pirie and Bruce Banks, for example, appears on the basis of heat flow data to have rifted between 40 and 30 Ma (Maldonado et al., 1998, Barker et al. 2013), although sea-floor spreading in this small basin is dated as 25-20 Ma (Galindo-Zaldívar et al., 2014). Dredged material from the western side of Pirie Bank is compatible with the geology of Tierra del Fuego, with “gneisses and mica schists” said to have an age of 579 Ma, “granites” an Early Jurassic age of 183 Ma, “rhyolites, liparites and basalts” a Middle Jurassic age of 175–169 Ma, and “siltstone and sandstone” an age of 113 Ma (Kurentsova and Udintsev, 2004). Their sequence and the dredging characteristics (including “fresh separation marks”) led Schenke and Udintsev, (2009) to believe that they were *in situ* and not ice- rafted. Precambrian rocks have not been reported in southern Tierra del Fuego, but pre–Late Jurassic basement gneisses and foliated plutonic rocks have been recovered in drilling operations in the Magallanes-Austral basin foredeep north of the Cordillera and are reliably dated as Cambrian with inherited Precambrian zircons (Hervé et al., 2010). Hence the correlation of the Pirie Bank dredge sample with Tierra del Fuego stratigraphy is plausible even though details of dredging and dating are not available in the English-language literature. More recent dredging (Riley et al., 2018) suggests that Pirie Bank is dominantly formed of low-grade metasedimentary rocks, Bruce Bank has sedimentary and plutonic (granitoid) rocks, and Discovery Bank is formed of intermediate volcanic and volcaniclastic rocks.

In their geophysically based reconstruction for 50 Ma, prior to opening of Drake Passage, Eagles and Jokat (2014) restore these banks against Tierra del Fuego, naming them collectively Omond Land, after the meteorological station of the Scottish National Antarctic Expedition on the South Orkney Islands. They incorrectly state, however (p. 38), that when reconstructed against Tierra del Fuego, these banks occupy the space commonly taken up by South Georgia in geologically based reconstructions. In fact, their reconstruction for 50Ma (their Fig.6) leaves a gap to the NE of their Omond Land in precisely the part of the Cape Horn escarpment where the palinspastic reconstruction of the South Georgia microcontinent indicates it would have been located as discussed here.

Figure 12 about here

12. Satellite-altimetry derived gravity image of the Scotia Arc showing the present and palinspastically restored positions of the South Georgia microcontinent and other former fragments of the southernmost Andean Cordillera in the Late Cretaceous, ca 90 Ma, as discussed in the text. The red rectangle schematically encloses the arc/batholith in Tierra del Fuego, the South Georgia microcontient and the continental blocks of the southern Scotia Sea. The dominant strike in Tierra del Fuego, Isla de los Estados, and the palinspastically restored South Georgia microcontinent and Shag Rocks are shown by the gold lines.

AP–Antarctic Peninsula; BB–Bruce Bank; BP–Barker Plateau; BwB–Burdwood Bank; DB–Discovery Bank; DvB–Davis Bank; IE–Isla de los Estados; PB–Pirie Bank; SGM–South Georgia microcontinent; SRB–Shag Rocks Bank; SOI–South Orkney Islands; SShI–South Shetland Islands; T del F–Tierra del Fuego; TR–Terror Rise

As also noted above, over the years several workers have reconstructed the South Georgia micrcontinent within the Staten embayment (Position #1, Fig. 3; e.g., Dalziel et al., 1975; Tanner, 1982a; Macdonald et al., 1987; Torres-Carbonell et al., 2014; Poblete et al., 2016), and Eagles and Jokat (2014) have restored the banks of the southern Scotia Sea against the Cape Horn escarpment. However, there have been no attempts to reconstruct the original positions of all the isolated continental blocks within the Scotia arc since the pioneering efforts of Barker and Griffiths (1972). Despite the paucity of data from the blocks that are totally submerged, we present a tentative reconstruction here (Fig. 12). The palinspastic restoration involves three steps.

First, we re-position the blocks located in the southern Scotia Sea, the Terror Rise, Pirie Bank, Bruce Bank and Discovery Bank (Figs. 3 and 12). As noted by Eagles and Jokat (2014), these were displaced from the Cape Horn escarpment by seafloor spreading on the west Scotia Ridge between ca. 30 and 6 Ma and subsequently separated by the formation of small rift basins (Galindo–Zaldivar et al., 2014). We restore them against Tierra del Fuego, with which their geological affinity from dredge samples has been discussed above, closing the intervening basins and moving the blocks northwestward along the transform fault separating the first and second segments of the west Scotia Ridge (W1 and W2 of Eagles et al. (2005). The blocks shown restored in this manner in Figure 12 have been reduced in width (E-W) by approximately 50 percent to account for extension during rifting but retain their original N-S length. As also noted by Eagles and Jokat (2014), this restoration aligns the high magnetic susceptibility anomaly over the Patagonian arc and batholith with comparable anomalies over the southern parts of all four continental blocks. Only the southernmost part of the Discovery Bank is shown as the northern part appears from dredging and dating to be part of the Neogene ancestral South Sandwich arc (Dalziel et al., 2013a,b; Pearce et al., 2014; Fig.3).

With the southern Scotia Sea blocks restored through control from the seafloor data, the second step in the overall reconstruction can proceed. That is the positioning of the South Georgia microcontinent itself, by far the largest continental fragment displaced from the Staten embayment. This is accomplished by locating the microcontinent with the strike of the Cretaceous sedimentary strata, that form most of the island and parallel the long axis of the block, in line with that of the west-east trending correlative strata in Tierra del Fuego. This in turn places the high magnetic susceptibility anomaly over the arc on the southern part of the microcontinent (Barker and Griffiths, 1972; Simpson and Griffiths, 1982) in line with that over the

Fuegian arc and batholith (Fig. 12). As the SGM was apparently transported east relative to South America on strike-slip faults during some form of 'escape tectonics' (discussed below), there seems no need to adjust for major change in dimensions. We merely suggest that a rather 'close fit' reconstruction is appropriate for all the blocks of the North Scotia Ridge to account for likely marginal deformation during separation and transport eastward relative to South America.

Third, with the position of the South Georgia microcontinent established, we can consider the former location of the Davis Bank, Barker Plateau, and Shag Rocks Bank that form the North Scotia Ridge between the autochthonous Andean Burdwood Bank and the SGM. Notably, although at least the Davis Bank and Barker Plateau have, from dredged samples, including an *in situ* tonalite dated at 49.6 ± 0.3 Ma and sandstones dominated by Cretaceous peaks and an age structure similar to the Sandebugten Formation (Pandey et al., 2010; Riley et al., 2019), affinity with the Fuegian arc and batholith, as well as with South Georgia and Burdwood Bank, they lack characteristic magnetic signature of the arc and batholith (Benniest and Schellar, 2020, Fig. 4). They are apparently, therefore, more closely associated with the northern part of the SGM and Burdwood Bank, apparently with calc-alkaline intrusives and volcanics like the Beagle granite suite of Tierra del Fuego (Mukasa and Dalziel, 1996) reflecting continentward motion of the focus of arc activity following closure and inversion of the Rocas Verdes basin in the mid– to Late Cretaceous, as is apparent in the Fuegian Andes. We therefore suggest they were originally located between the SGM and the eastern end of Burdwood Bank, while the stilpnomelane-bearing greenschists of the Shag Rocks Bank were located immediately south of Isla de Los Estados. That places them in a position structurally above the inverted Tobifera/Lemaire Formation of the island (Dalziel and Palmer, 1979) and thus comparable to that of the petrologically similar schists of the Lapataia Formation along strike to the west in on the north shore of the Beagle Channel in Tierra del Fuego as discussed above. And, as shown in Figure 11, it is worth observing that this tentative reconstruction incorporates all the continental blocks of the Scotia arc within the Staten embayment without significant unaccounted space.

Regarding a mechanism for translation of the South Georgia microcontinent eastward relative to South America, this remains tantalizingly obscure. Part of this uncertainty stems from ambiguity in the age of parts of the floor of the Scotia Sea and the fact that older, Mesozoic segments have been subducted (Dalziel et al., 2013a; Riley et al., 2019). However, confirmation of the post-Early Cretaceous counterclockwise rotation of the arc and batholith during closure and inversion of the marginal basin by Poblete et al. (2014, 2016) makes it likely that oroclinal bending played a significant role as discussed in the following section.

6. Geodynamic considerations

The geodynamic puzzle posed by our reconstruction, as pointed out by Eagles (2010), is that only about half of the necessary translation of South Georgia with respect to South America can reasonably be ascribed to seafloor spreading on the northeast-southwest oriented west Scotia Sea spreading centre. This led to his assertion that the other half of the necessary motion "cannot have occurred", claiming that "plate kinematic data show the South Georgia microcontinent as having originated at the eastern end of the North Scotia Ridge", in spite of its "tectonic likeness to Tierra del Fuego". He did not consider that South Georgia exactly comprises one major segment of the Andean hinterland in Tierra del Fuego that terminates abruptly at the Cape Horn escarpment and is missing from the Staten embayment, namely the Rocas Verdes basin basement and infill. Nor did he take into account the characteristically Andean detrital zircon populations of the sedimentary strata forming most of the island of South Georgia. We therefore continue to seek a mechanism for the separation of the microcontinent from the southernmost Andes while contending that the absence of a clear solution does not negate the extremely robust and compelling geologic evidence for the palinspastic reconstruction presented above.

Clearly the South Georgia microcontinent was displaced from the South American continent by a radically different process than the comparatively straightforward seafloor spreading on the West Scotia Ridge that moved the several components of Omond Land to their present location along the southern margin of the Scotia Sea. To understand the tectonic history of the micro-continent it is necessary to consider the origin of the prominent bend in the southernmost Andes in the vicinity of the western end of the Straits of Magellan where they swing through 90° from their predominantly north-south trend to strike eastward into the North Scotia Ridge: the Patagonian orocline of Carey (1955, 1958).

Paleomagnetic studies going back almost 50 years have indicated that substantial counterclockwise vertical axis rotation occurred along the Pacific margin of the Cordillera since the Early Cretaceous (Dalziel et al., 1973; Burns et al., 1980; Cunningham et al., 1991). Comprehensive modern studies of paleomagnetism and magnetic susceptibility anisotropy have not only confirmed the early results, but have demonstrated that while the magmatic arc oceanward of the Rocas Verdes basin was rotated counterclockwise even in excess of 90°, the curvature of the foreland fold and thrust belt is a primary arc reflecting the original configuration of the continental margin (Poblete et al., 2014, 2016).

The counterclockwise rotation of the magmatic arc towards the convex continental margin (Figs. 10, 12) occurred during the mid-Cretaceous closure and inversion of the Rocas Verdes basin in the initial compressive phase of the Andean orogeny that affected the rocks of the South Georgia microcontinent as part of the Pacific margin of South America at that time. The Weddell Sea had started to open in the Early Cretaceous, (Riley et al., 2020), possibly as a continuation of the Rocas Verdes basin (Mukasa and Dalziel, 1996; Poblete et al., 2016), or perhaps as a 'twin' marginal basin behind the Antarctic Peninsula arc across an embayment in the Gondwana margin. Regardless, this meant that the South Georgia microcontinent was located at a 'free' margin of the South American continent. We suggest that much of the translation from the South American continent took place along a strike-slip fault or faults initiated in a process of 'escape tectonics' during Rocas Verdes basin closure in a manner comparable to the westward motion of Anatolia during collision of the Arabian Peninsula with Asia (Burke and Sengor, 1986). Indeed, the entire North Scotia Ridge east of Burdwood Bank (terminating at about 56° W, and perhaps stretched and submerged but still an autochthonous segment of the Andean Cordillera) is likely to represent an 'orogenic stream' of crustal blocks similar to the North Pacific rim in Alaska, which was formed by escape towards the 'free'margin of the Bering Strait from the collision of the Yakutat block with the eastern end of the Aleutian subduction zone (Redfield et al., 2007). Our palispastic reconstruction suggests that the Davis Bank and Barker Plateau may have been displaced eastward early in the 'escape tectonics' regime. This would have allowed the SGM and the Shag Rocks Bank to 'slide by' *en route* to their present more easterly location as part of a strike-slip duplex , and as the SGM is apparently moving today (Dalziel et al., 2019). independent of both the major adjoining South American and Scotia plates.

7. Conclusions

The geological evidence that the South Georgia microcontinent was formerly a continuation of the Andean Cordillera located immediately east of Tierra del Fuego is overwhelming. Not only do the rock units and tectonic history of South Georgia Island correlate in every significant respect with those of the Andean hinterland south of the Beagle Channel, exact matches for sources of sedimentary strata forming much of the island occur in the easternmost extremity of the Cordillera. Critically, moreover, South Georgia is an exact geologic match for the part of the hinterland immediately south of the Beagle Channel that is truncated at the cross-strike Cape Horn escarpment. These units, so spectacularly exposed in southern Tierra del Fuego, are missing from the Staten embayment south of the main range of the Cordillera which trends east from Peninsula Mitre through Isla de los Estados to the submerged Burdwood Bank. Although now far apart, both South Georgia and the southern Andes preserve the geological record of the expansion, infill and deformation of contiguous parts of the same Late Jurassic to Early Cretaceous marginal basin.

While acknowledging that the structure and history of the oceanic lithosphere in Drake Passage and the Scotia Sea cannot fully explain the present-day separation of the microcontinent from its parent continent, we suggest that a significant portion of that separation may have occurred through a form of eastward sinistral 'escape tectonics' along a transcurrent fault or faults during the mid– to Late Cretaceous closure and inversion of the Rocas Verdes basin that formed the orocline in the Patagonian arc and batholith. Indeed the submerged banks of the North Scotia

Ridge and the South Georgia microcontinent may together form a complex strike-slip 'duplex' system with South Georgia now occupying a restraining bend as it collides with the Northeast Georgia Rise along the South America-Scotia sinistral transform plate boundary.

Restoration of the South Georgia microcontinent along strike from the Andean hinterland in Tierra del Fuego has the paleogeographic implication that this extension of the Andean margin was undergoing subduction during the Early Cretaceous. Thus, if the Antarctic Peninsula was ever located 'outboard' of southern Patagonia, as shown in many reconstructions of the Gondwana supercontinent, including some by the present authors, it must have been translated south of the Andean margin by the earliest Cretaceous, approximately 140 Ma. The original position of the Antarctic Peninsula with respect to southernmost South America remains enigmatic. The early opening of the Weddell Sea embayment between the Antarctic Peninsula and the East Antarctic craton appears to have been contemporaneous with the opening of the Rocas Verdes basin (Table 1) and an alternative possibility to 'overlapping' the peninsula with southern Patagonia might be a cuspate relationship between the two with an embayment between them in the area of the present Scotia Sea. This is beyond the scope of the present contribution.

A further implication of our palinspastic restoration of the South Georgia microcontinent immediately east of Tierra del Fuego is that it must have been present along the evolving North Scotia Ridge throughout the Late Cretaceous and Cenozoic. Thus it needs to be taken into account in any consideration of the effect of submarine topography on the developing Antarctic Circumpolar Current. It was part of the initial compressive Andean Cordillera in the mid–Cretaceous, still attached to South America at 40 Ma, apparently subsided below sea level after separation, and was uplifted again on Miocene collision with the Northeast Georgia Rise. As a fragment of Andean continental crust; however, it must have been a barrier to deep oceanic circulation throughout this history.

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Abbreviations Footnote:

ACC–Antarctic Circumpolar Current; DFC–Drygalski Fjord Complex; LHC–Larsen Harbour Complex; NSR–North Scotia Ridge: SC–Sarmiento Complex; TC–Tortuga Complex; SGM–South Georgia microcontinent

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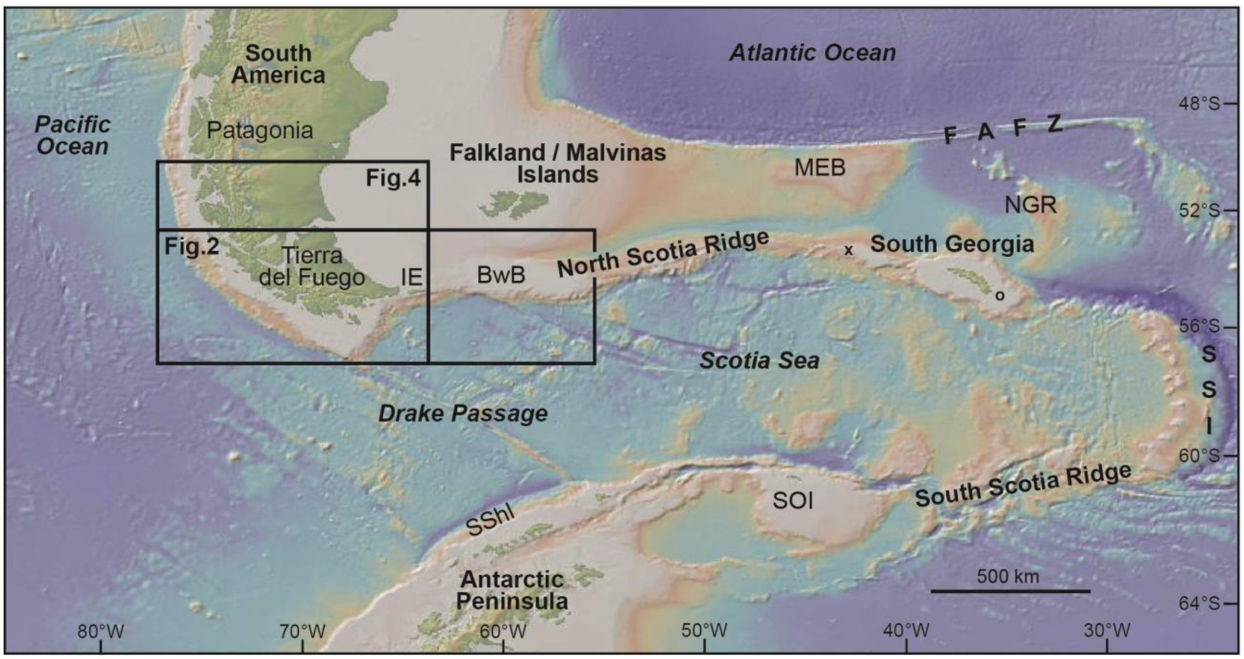
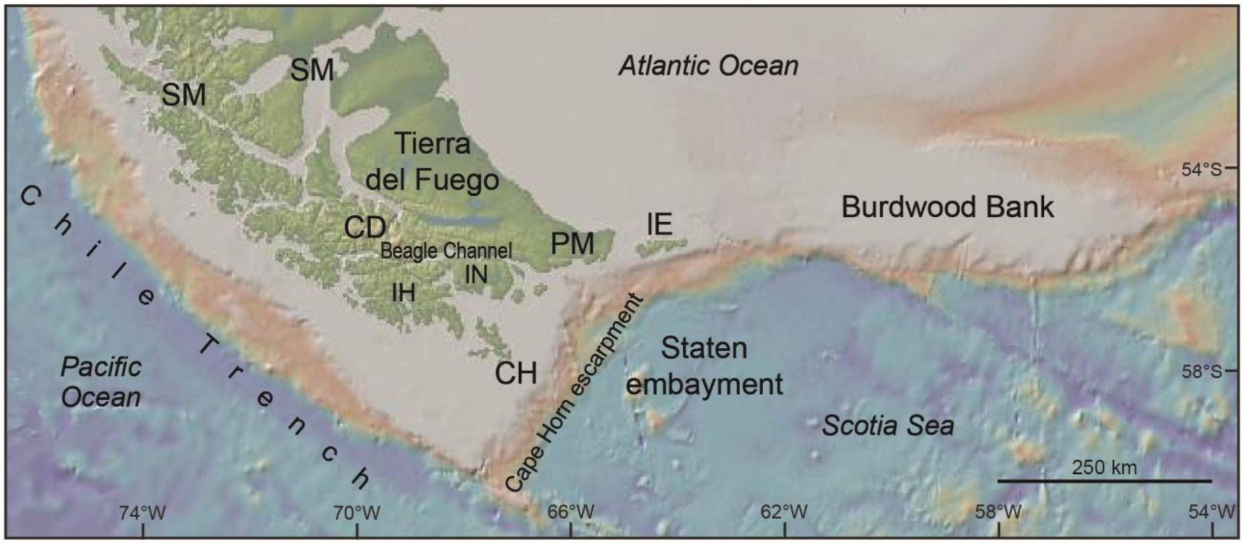


Figure 1



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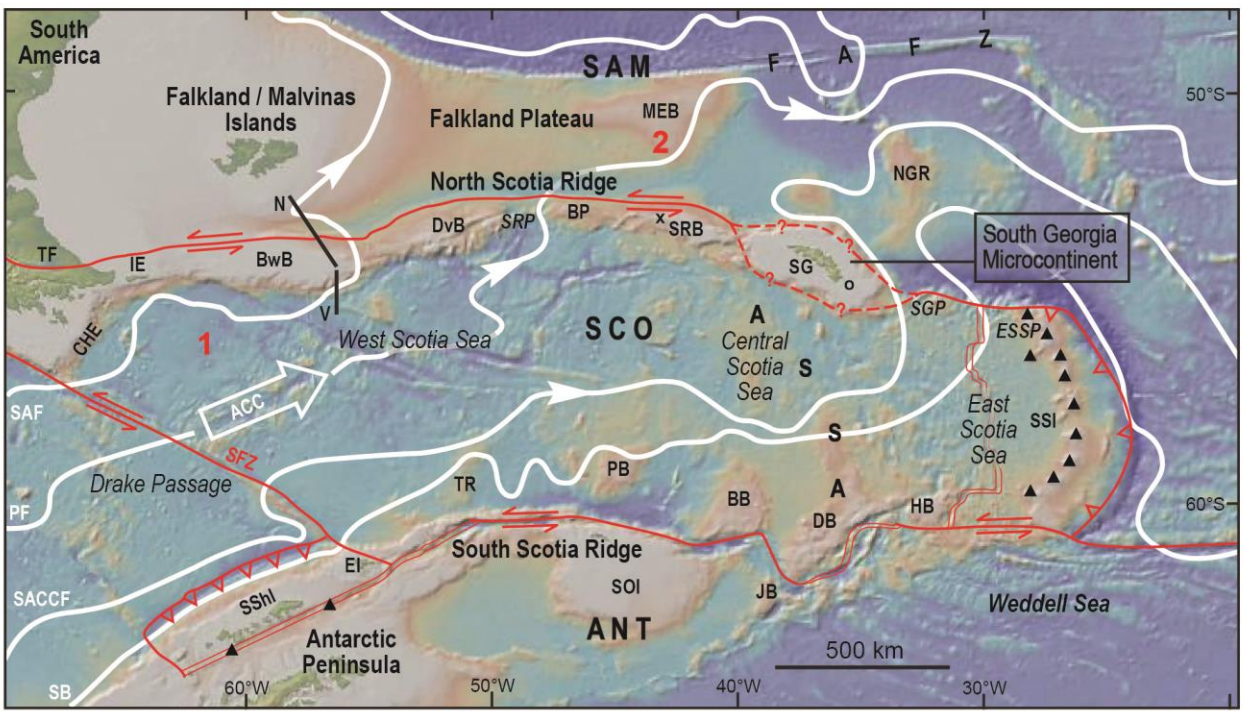

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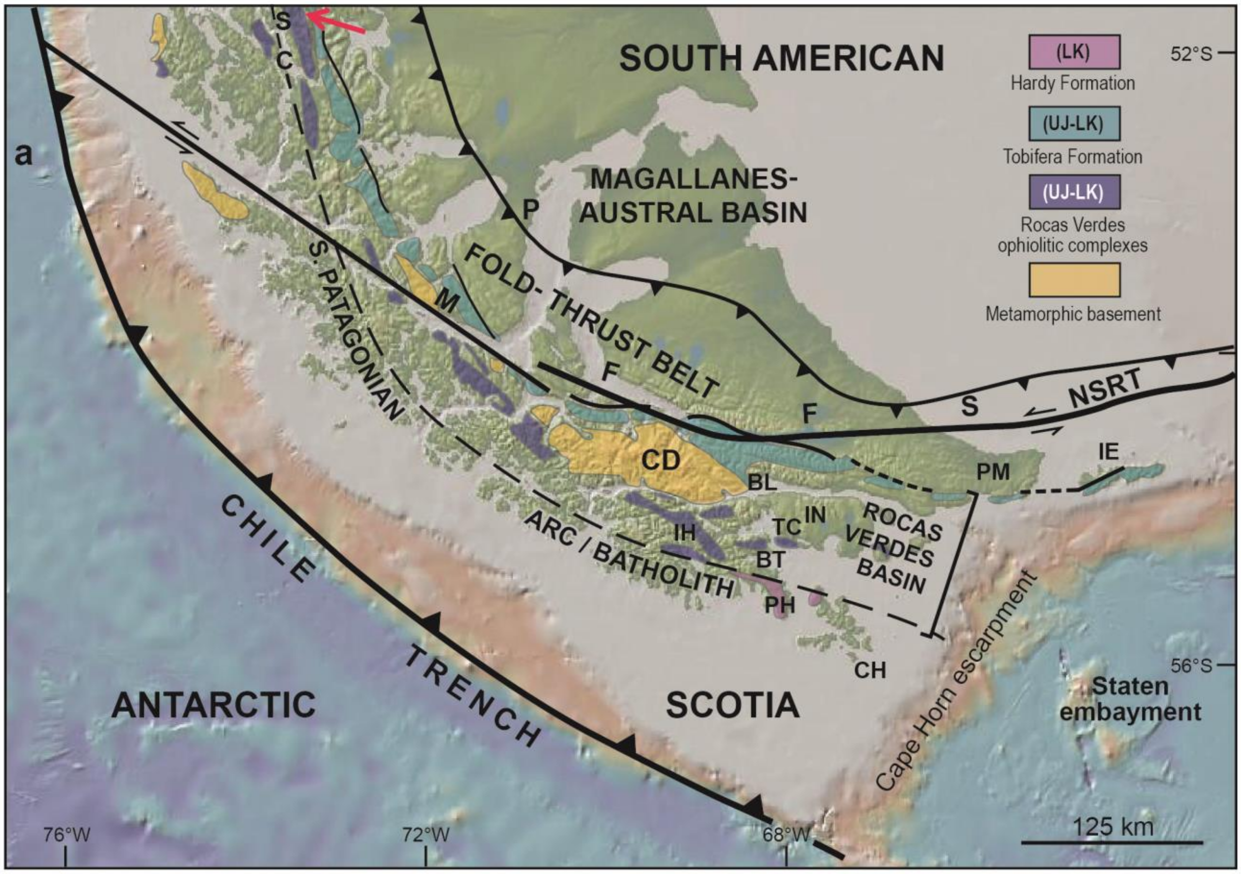

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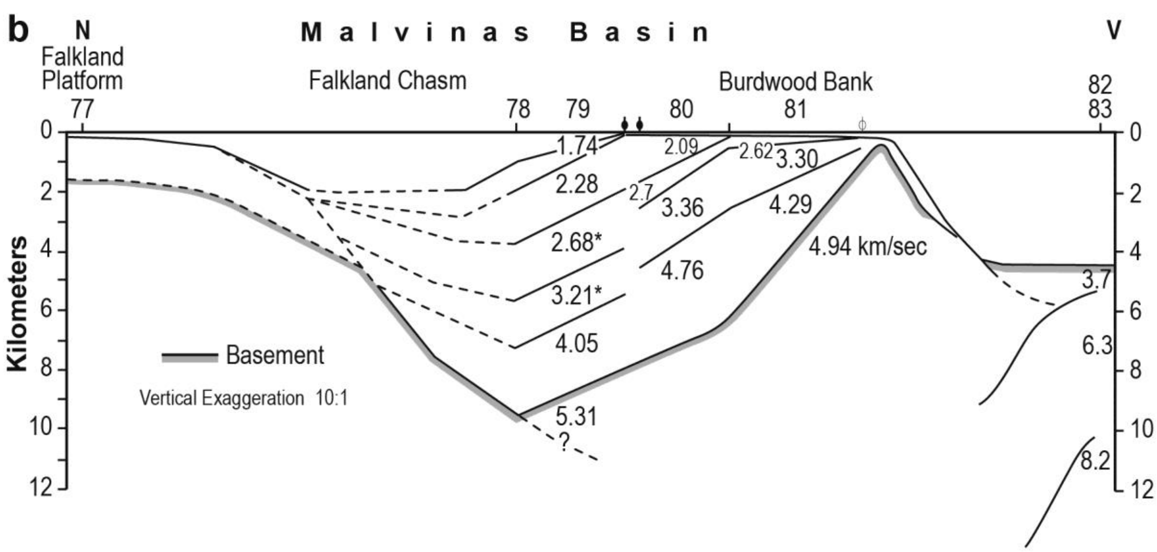

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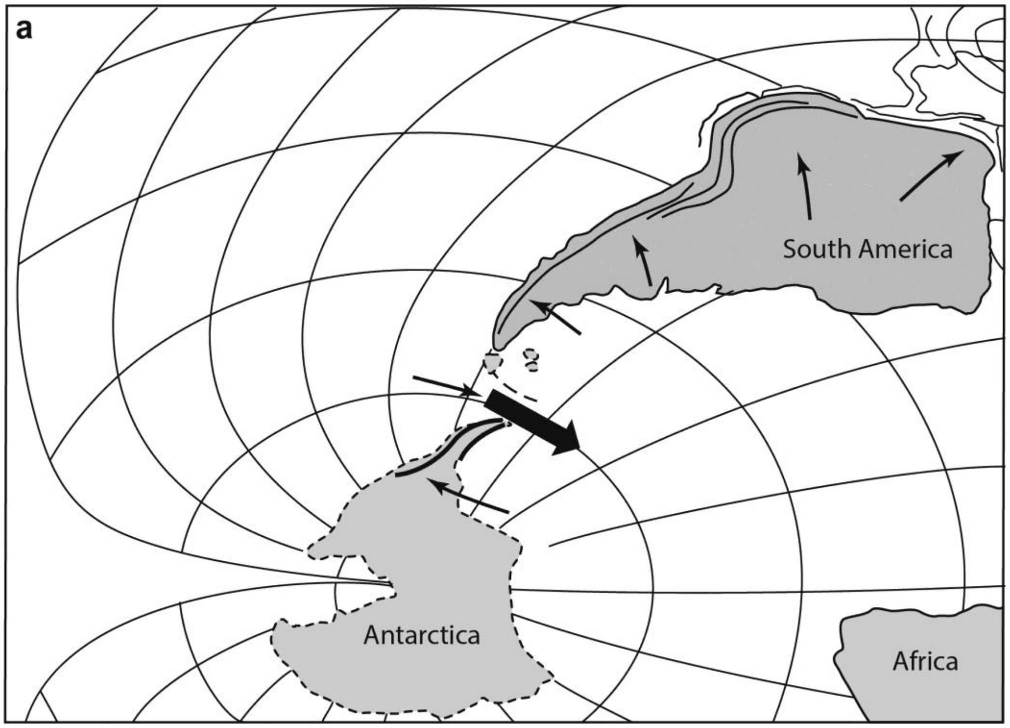

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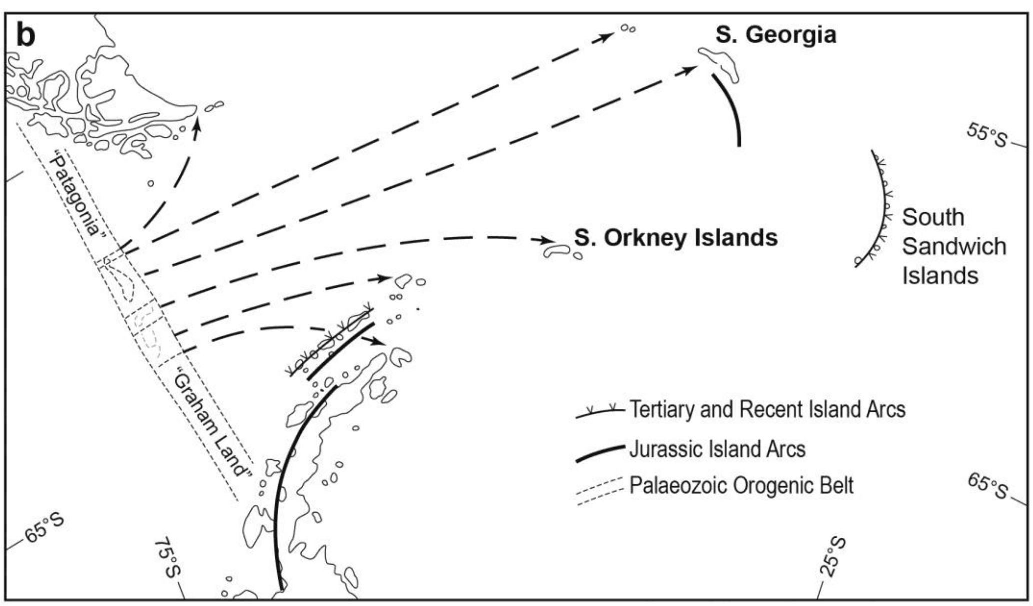


Figure 5b

Figure 6

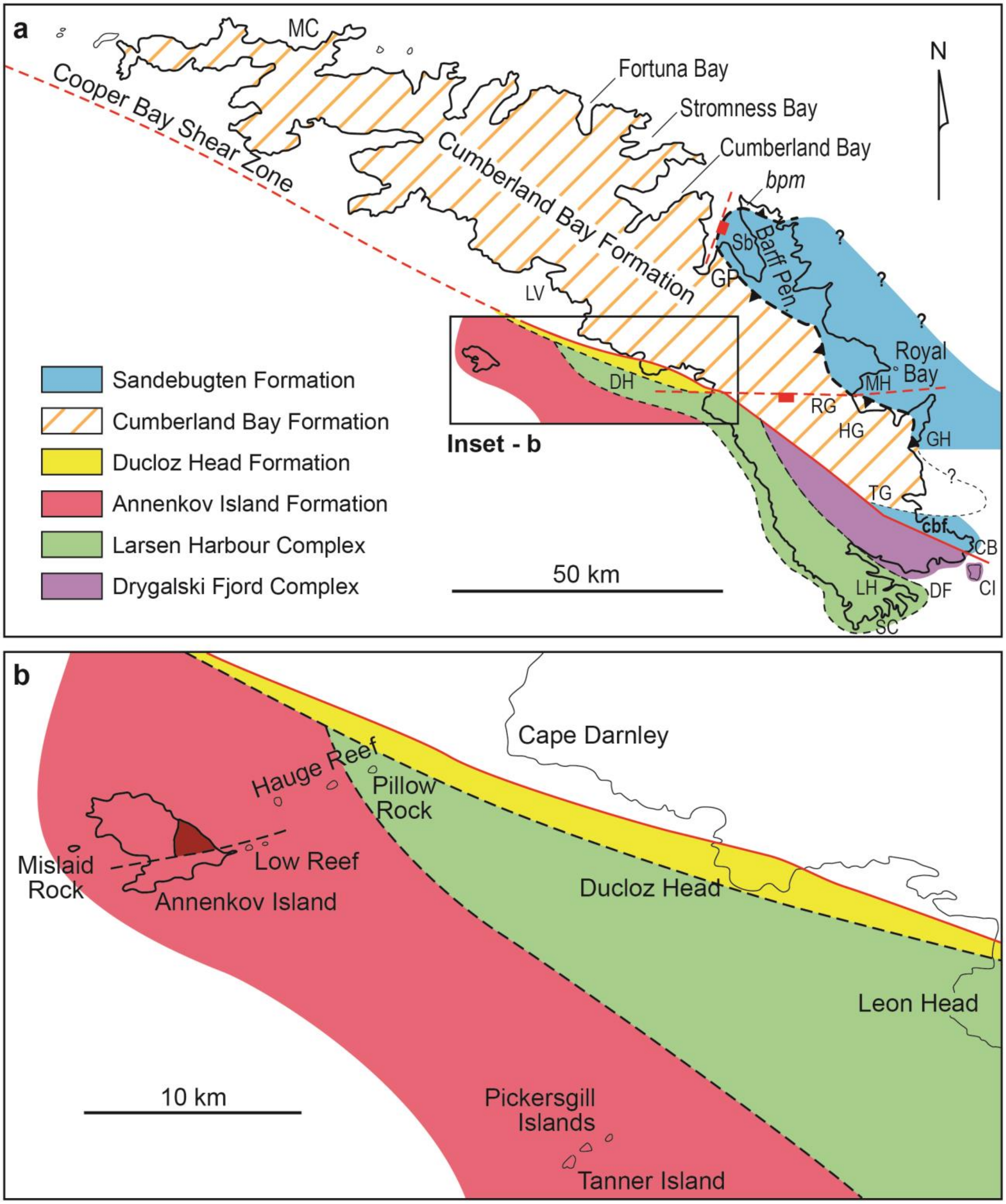


Figure 7a and b

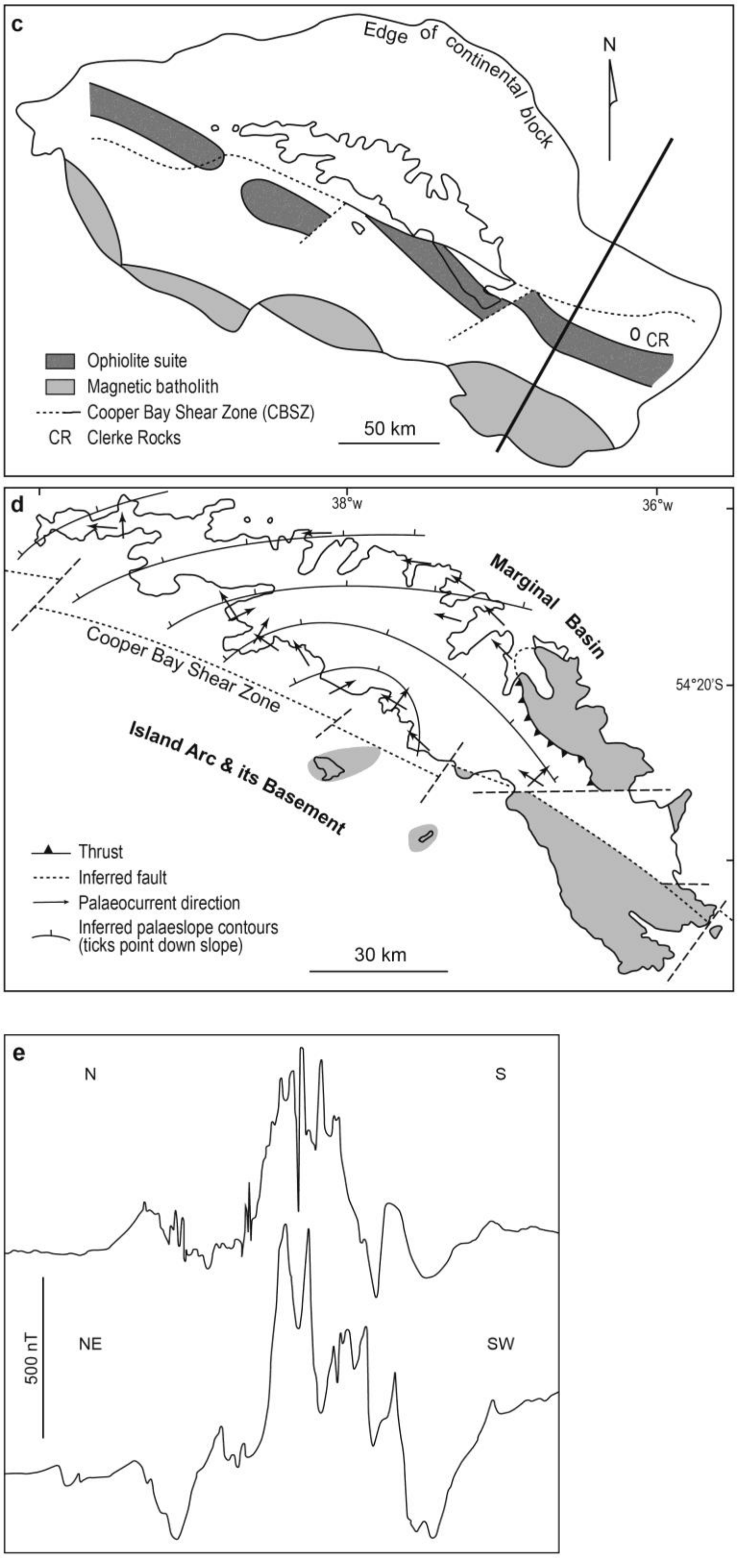
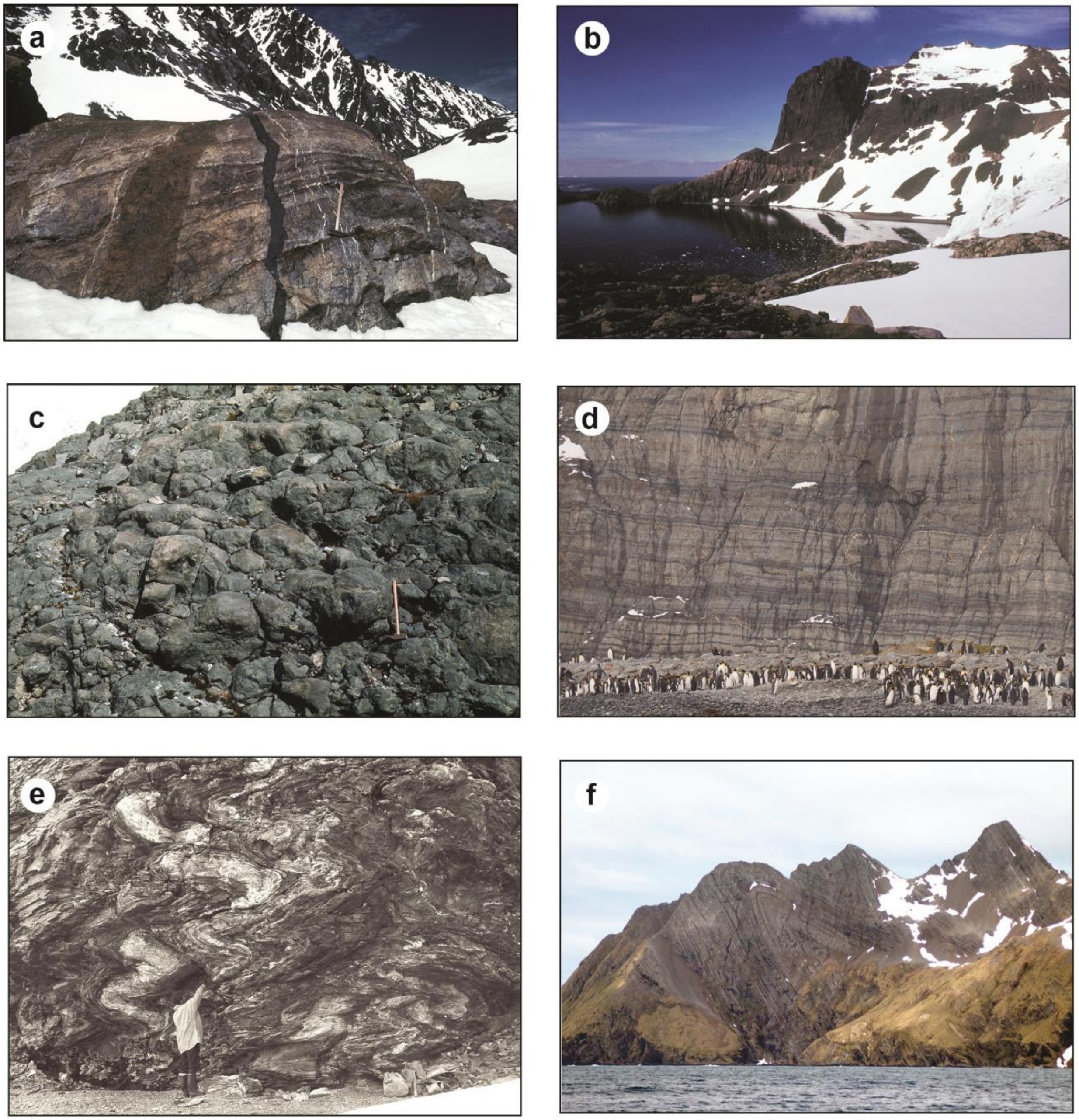


Figure 7 c, d and e



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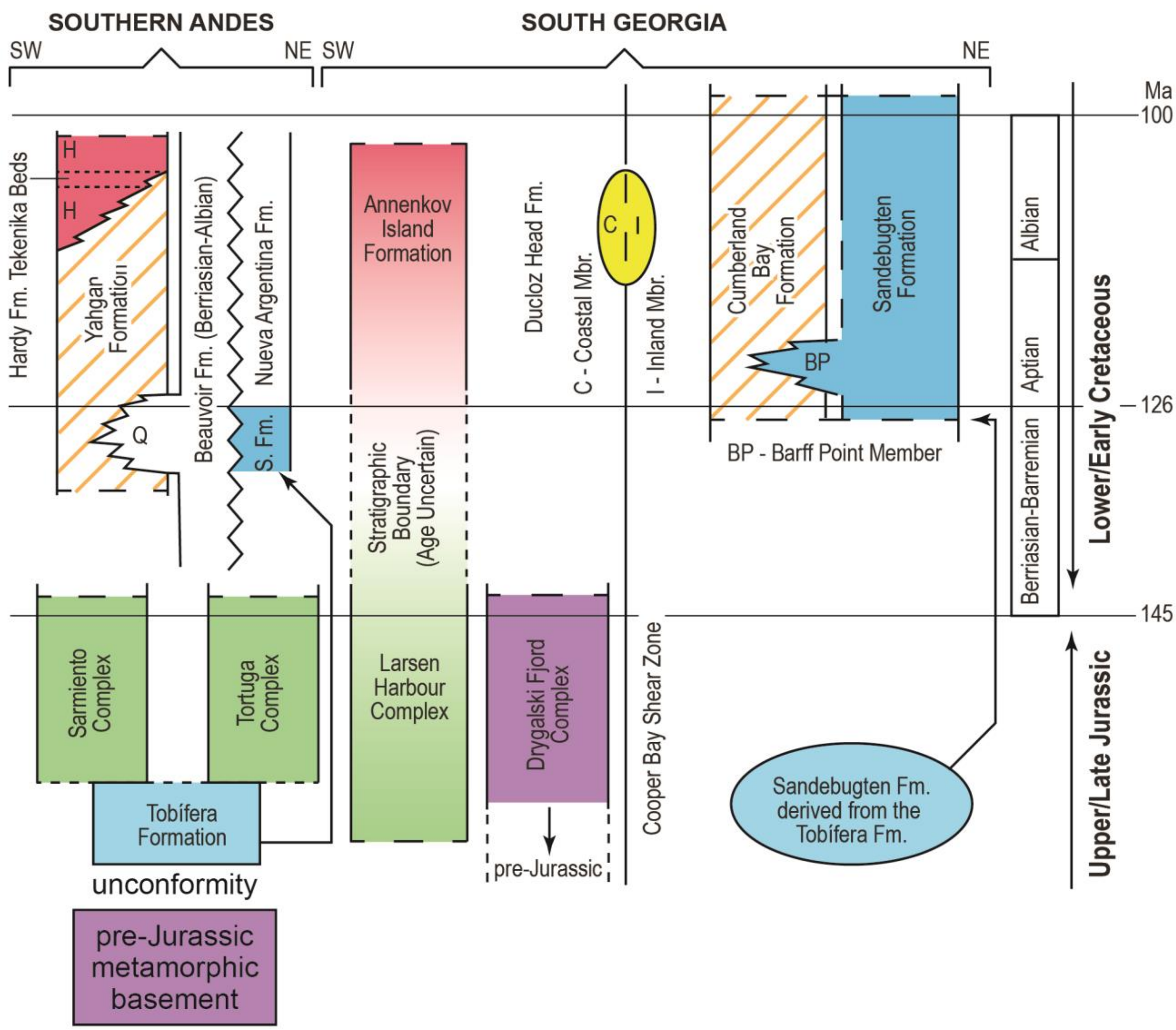
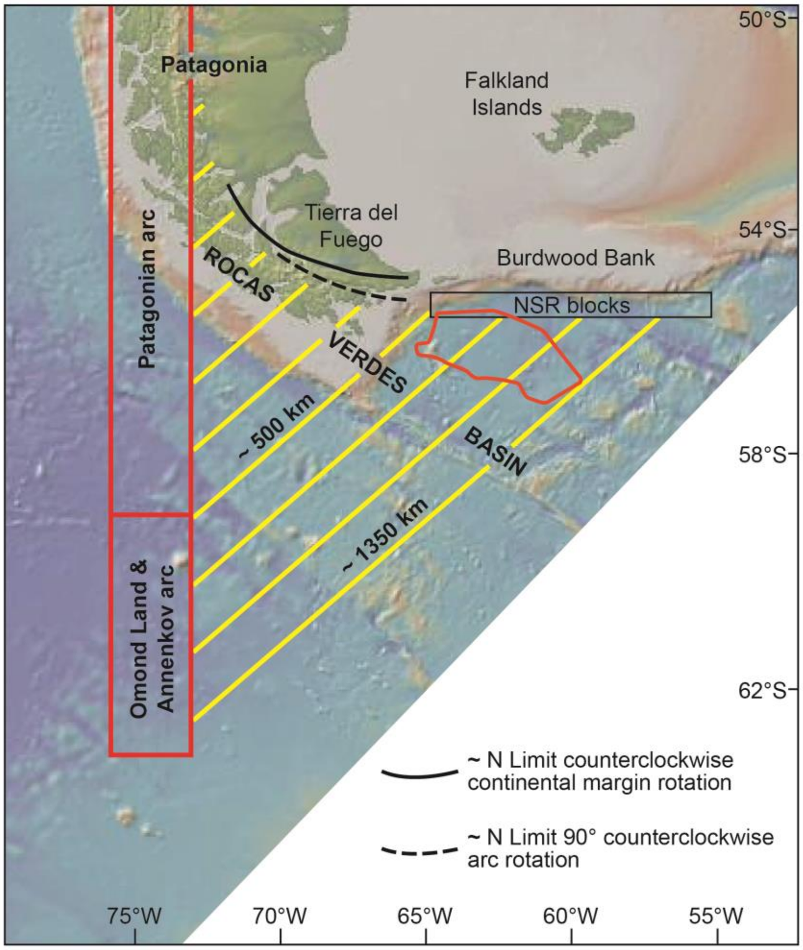


SOUTHERN ANDES
SW
NE
SOUTH GEORGIA
SW
NE
Ma
100
126
145
Hardy Fm. Tekenika Beds
H
H
Yahgan Formation
Q
Beauvoir Fm. (Berriasian-Albian)
Nueva Argentina Fm.
S. Fm.
Annenkov Island Formation
Stratigraphic Boundary (Age Uncertain)
Ducloz Head Fm.
C - Coastal Mbr.
C
I
I - Inland Mbr.
Cumberland Bay Formation
BP
Sandebugten Formation
BP - Barff Point Member
Albian
Aptian
Berriasian-Barremian
Lower/Early Cretaceous
Upper/Late Jurassic
Sarmiento Complex
Tortuga Complex
Larsen Harbour Complex
Drygalski Fjord Complex
Cooper Bay Shear Zone
Tobífera Formation
pre-Jurassic
Sandebugten Fm. derived from the Tobífera Fm.
unconformity
pre-Jurassic metamorphic basement

Figure 9



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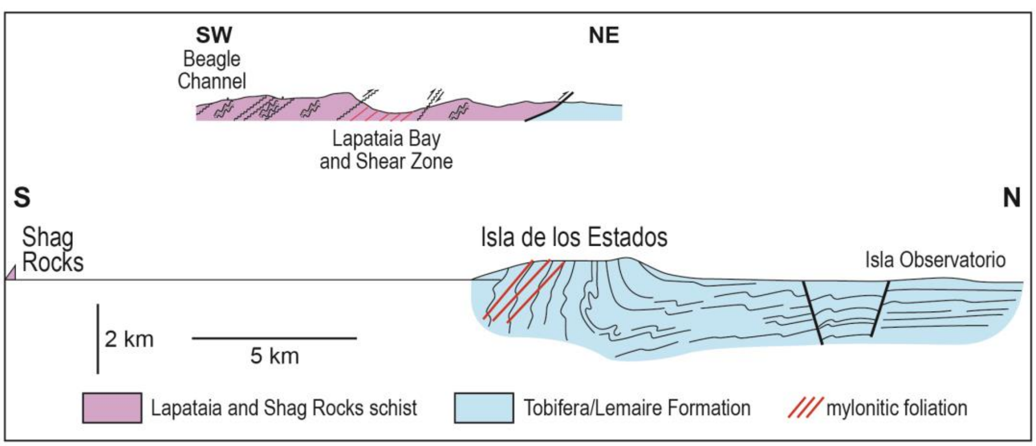

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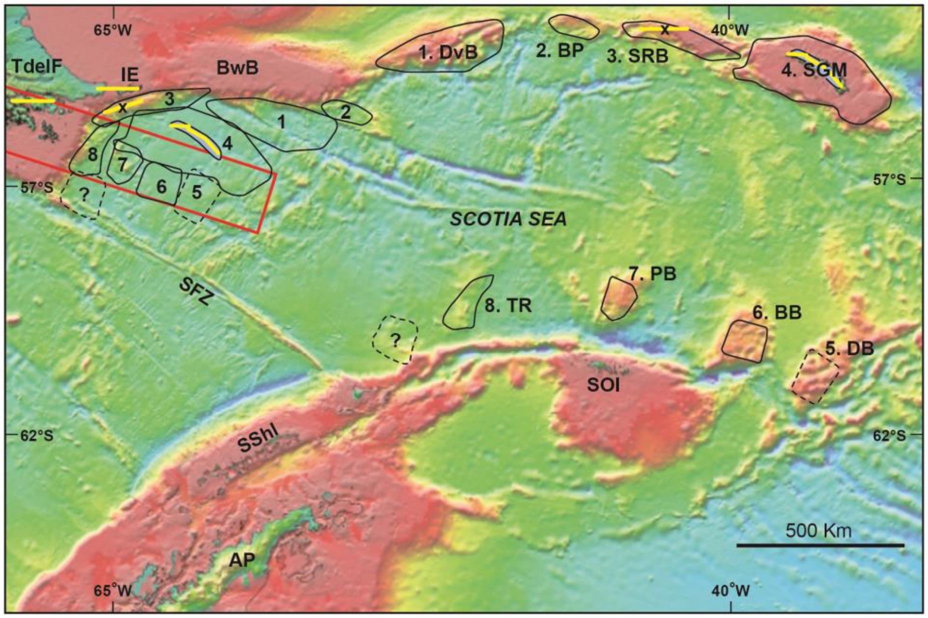

Figure 12

Southern South America	South Georgia	South Atlantic/Weddell Sea
Arc magmatism	Tectonic uplift, Collision with NE Georgia Rise (12-10 Ma)	Seafloor spreading
Arc magmatism Cooling event (45-40 Ma)	Subsidence (40-10 Ma) Final separation from South America (~45-40 Ma) Cooling event (45-40 Ma)	Seafloor spreading
Initiation: Andean compresseion/uplift closure/inversion Rocas Verdes Basin, Patagonian arc orocline (mid-Late Cretaceous, ~95 Ma)	Initiate eastward tectonic escape (?); orogenic compression/ uplift (~95 Ma)	Magnetic quiet interval; increase spreading rate S. Atlantic and subduction rate Andean margin (~121-83 Ma)
		Seafloor spreading
Rocas Verdes Basin infill (Yahgan Fm.); Arc volcanism (Hardy Fm.) (L. Cretaceous)	Basin infilling (Cumberland Bay and Sandebugten fmns.); arc volcanism (Annekov Island Fm.) (L. Cretaceous)	Seafloor spreading
Establish: Andean margin subduction, arc magmatism (~140 Ma)	Initiate subduction, arc magmatism (~140 Ma)	Extension and dykes Falkland Is. (~135 Ma) Rift-drift transition S. Atlantic (~135 Ma) Parana-Etendeka LIP (139-127 Ma)
		Initial extension/rifting S. Atlantic Falkland Islands and north
Rocas Verdes ophiolitic magmatism (~140-150 Ma)	Larsen Harbour ophiolitic magmatism (~150 Ma)	Rift-drift transition Weddell Sea (~147 Ma)
?Slab roll back at Andean margin Extension/rifting, Tobifera Fm. (~170 Ma)	Extension/rifting	Karoo-Ferrar LIP (~182 Ma) initial extension/ rifting south of Falkland-Agulhas Fracture Zone
		Extension/dykes Falkland Islands (~190-170 Ma)
Gondwanide orogeny (Late Permian-Triassic)		Gondwanide orogeny (Late Permian-Triassic)

Table 1