An open access geological map of the South Shetland Islands, Antarctica: a lithostratigraphic compilation

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Abstract

Over the last few decades, numerous geological studies have been carried out in the South Shetland Islands, which have greatly contributed to a better understanding of its geological evolution. However, few attempts have been conducted to correlate the geological units throughout this archipelago, with the most important work to date being that of Smellie et al. (1984). We present herein a lithostratigraphical correlation for the main Mesozoic and Cenozoic units of the South Shetland Islands along with a new geological map. The lithostratigraphical correlation shows that the geological evolution comprises three main stages: (i) deep marine sedimentation from ~164 to 139 Ma, (ii) subaerial volcanism and sedimentation with a proliferation of plant fauna from ~139 to 34 Ma and (iii) glacial and interglacial deposits from ~34 Ma. The shapefile containing the geological information was developed in an open-source GIS software (QGIS 3.24) and is possible to download herein. Additionally, the full project with the shapefile is available to download from the servers of the University of Geneva to serve as a resource for future work by geoscientists on the South Shetland Islands.

1. Introduction

The South Shetland Islands extend for ~300 km in a north-eastern direction parallel to the northern tip of the Antarctic Peninsula. The archipelago is separated from the Antarctic Peninsula by the Bransfield Strait (Fig. 1). It was attached to the peninsula until the Pliocene (~5 Ma), when back-arc rifting opened the Bransfield Strait which represents an active marginal basin (e.g. Barker, 1982). Thus, the archipelago is a recently detached crustal block (a young microplate) and is comprised of late Palaeozoic metasedimentary basement (e.g. Castillo et al., 2015) overlain by Jurassic and younger magmatic and sedimentary rocks (e.g. Smellie et al., 1984; Haase et al., 2012; Bastias et al., 2019). The rock exposures in the peninsula exhibit geological evidence of Mesozoic to Cenozoic tectonic, global sea-level and climate change, which are notably recorded in the South Shetland Islands by the emergence of submarine marginal basins as part of a continental island volcanic arc (e.g. Hathway and Lomas, 1998; Riley et al., 2012; Bastias et al., 2019) and the consequent proliferation of Cretaceous plant species (e.g. Falcon-Lang and Cantrill, 2001; Torres et al., 1997; 2015; Philippe et al., 1995; Leppe et al., 2007; Warny et al., 2019). This was followed during the Eocene by the development of a mountainous landscape with stratovolcanoes and lava fields that were covered by Valdivian-type forest (Poole et al., 2001: Hunt and Poole, 2003), which are similar to wetlands and freshwater environments still present today in Chile and Argentina. Later, approaching the Eocene/Oligocene boundary, northward propagation of the Antarctic icesheet and marine transgression occurred (Baker, 2007; Davies et al., 2021), which led to glacial, glaciomarine and marine sedimentation (Troedson and Riding, 2002; Troedson and Smellie, 2002). This was followed by a period of glaciations and interglacials that are recorded at the northwestern end of the archipelago on King George Island (e.g. Birkenmajer et al., 1985; Birkenmajer, 2001; Whittle et al., 2015; Troedson and Smellie, 2002; Smellie et al., 2021). These paleoenvironmental changes events are related to globally significant climate events (e.g. Kennett, 1977; Zachos et al., 2001; De Conto & Pollard, 2003), and the South Shetland Islands are one of the few regions in Antarctica where these processes are exposed in the geological strata. Therefore, a better understanding of the lithostratigraphic evolution of the South Shetland Islands is key to determine the timings and extent of these events. However, after decades of geological research, our geological knowledge of the South Shetland archipelago is limited, with most studies concentrated on the more accessible islands (King George and Livingston islands). This poses a significant problem when regional correlations between geological units are required. Additionally, while there are a plethora of local-scale geological maps, these are mostly located near research stations and unfortunately often introduce new geological units as opposed to attempting regional correlation. The most important geological map of the South Shetland archipelago was presented in the work of Smellie et al. (1984), and now after almost four decades, significant new information is available (e.g. Bastias et al., 2019; Smellie et al., 2021). This study presents a review of the main geological units of the nine major islands of the South Shetland archipelago. In doing so, we correlate these units across the archipelago to elucidate changes in paleoenvironmental conditions (i.e. submarine *vs* subaerial settings, the presence of glacial episodes). Finally, in addition to the review and correlation of the lithostratigraphical units of the South Shetland Islands, we present a polygon shapefile developed in open-source software (QGIS v3.24) that the reader can download and use freely. This should serve as a valuable tool for earth and environmental sciences researchers working in the region and act as a framework for future work on the geological evolution of the archipelago.

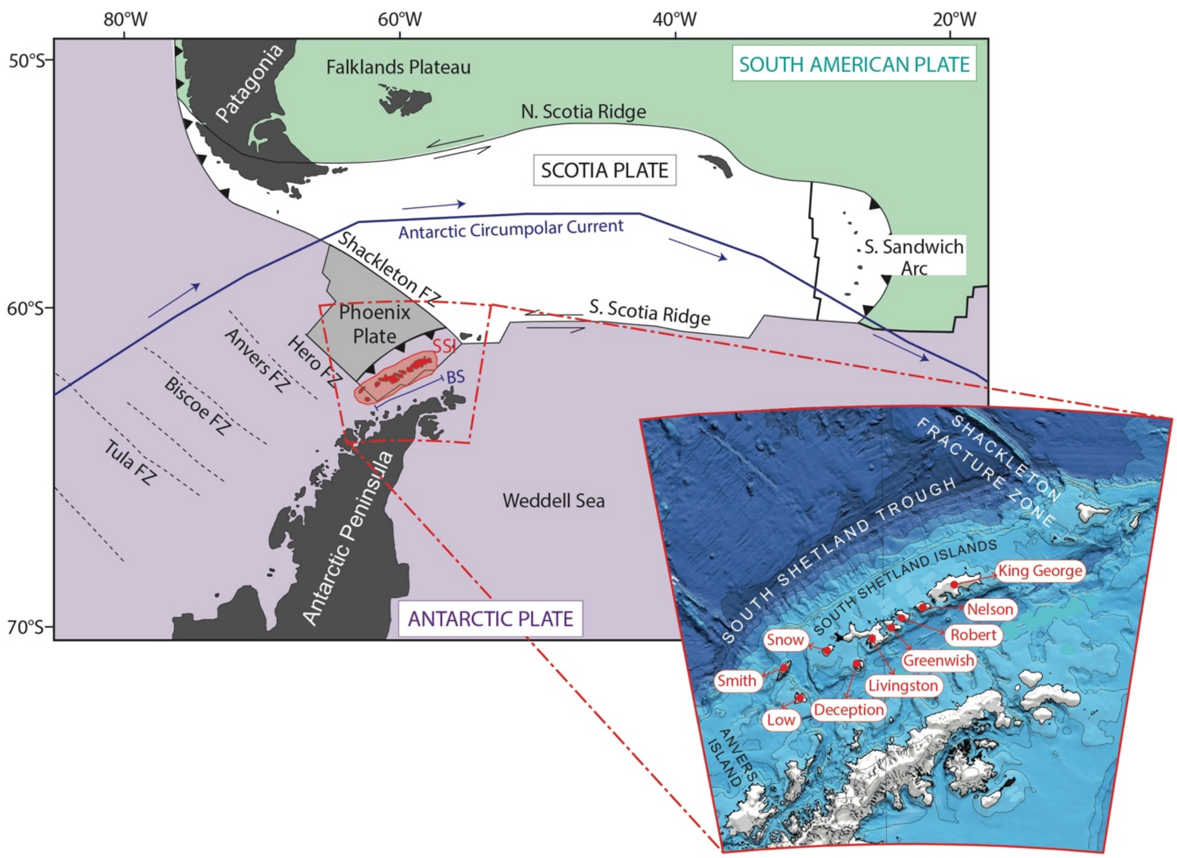

Figure 1. Tectonic configuration of the region encompassing the Scotia Plate, Patagonia and Antarctic Peninsula and showing the location of the South Shetland Islands and the Bransfield Strait. BS: Bransfield Strait. SSI: South Shetland Islands. Red box in inset figure shows the location of the archipelago and its main islands.

2. Geological history

The geological evolution of the Antarctic Peninsula is dominated by an active margin setting during the late Palaeozoic, Mesozoic and Cenozoic resulting from the subduction of the paleo-Pacific Plate (the so-called Phoenix Plate) beneath the Antarctic Peninsula in the southwestern coast of Gondwana (e.g. Burton-Johnson and Riley, 2015; Bastias et al., 2020, 2021, in review; Jordan et al., 2020). During the early Mesozoic the breakup of Gondwana occurred, which caused the separation of the Antarctic Peninsula from Patagonia and Africa by the opening of the Weddell Sea in the South Atlantic (e.g. König and Jokat, 2006). Breakup was associated with magmatism (Pankhurst et al., 2000; Riley et al., 2001; Bastias et al., 2021), deep-marine sedimentation (Trinity Peninsula Group; e.g. Castillo et al., 2015), and the development of volcano-sedimentary basins on the eastern (Larsen Basin; Hathway, 2000) and western flanks of the peninsula (Byers Basin; Bastias et al., 2019). With the exception of the high pressure-low temperature metamorphic accretionary complex on Smith Island (e.g. Truow et al., 1998), the South Shetland archipelago is comprised of Mesozoic and Cenozoic volcanic and volcaniclastic rocks (e.g. Smellie et al., 1984). While waning subduction off the South Shetland Islands - northwestern tip of the peninsula may still be occurring based on seismic activity (Robertson-Maurice et al., 2003), cessation of subduction along most of the margin progressively occurred from south to north by collision of spreading ridge segments separated by fractures zones (Larter and Barker, 1991). The segment of the margin where the South Shetland Islands lies is flanked by Hero and Shackleton fractures zones to the south and north, respectively (Fig. 1; e.g. Galindo-Zaldívar et al., 2004). Subduction is ceasing (Lawver et al., 1995), while the archipelago has lacked arc volcanism for the last ~20 Myr (Birkenmajer et al., 1986). Subduction-related convergence of the Phoenix Plate with the South Shetland Islands stopped at ~3 Ma when seafloor spreading in the ridge ceased (Eagles, 2004). This has led to slab rollback and transtensional motions between the Scotia and Antarctic plates, which caused extension and the opening of the Bransfield Strait (Fig. 1; Gonzalez-Casado et al., 2000; Maestro et al., 2007; Almendros et al., 2020).

3. Methodology

The geological maps were developed through different stages, which are explain herein.

1. Several contributions were consulted to build the geology presented in this work. Because our aim is to present regional lithostratigraphical correlations across the South Shetland archipelago, we have worked at a formation level of detail, when was present in the literature (Fig. 2a). This form of presenting the information allows to observe the evolution of the geological units throughout the Mesozoic and Cenozoic without the noise generated by the several units that may be present within a given geological formation. Furthermore, the lithological descriptions presented in this work are mainly derived from the geological maps published by Smellie (1979), Smellie et al. (1984,

2021), Grunow et al. (1992), Li et al. (1996), Hathway and Lomas (1998), Lee et al. (2002), Machado et al. (2005), Hervé et al. (2006), Kraus and del Valle (2008), Israel (2015), Warny et al. (2015), Bastias (2015) and Bastias et al. (2019). The geological symbols were obtained from the library presented by Frigeri (2020). Finally, it is noteworthy that we acknowledge that this compilation of maps certainly may not include all the geological information published in the Shetlands archipelago and thus this geological compilation is presented as a starting point for further development. Hence, we consider that this material has to be corrected and complemented by the geoscientific community working in the northern Antarctic Peninsula, which will leave to the improvement of this material. We therefore look forward to being contacted. Local geological map of each island presented in Fig. 3 throughout Fig. 13 were produced in Adobe Illustrator 2021.

2. The digitalization of these geological maps was performed in QGIS v3.24 using the EPSG:3031 - WGS 84 / Antarctic Polar Stereographic projection, based on the World Geodetic System 1984 ensemble (EPSG: 6326), which has a limited accuracy of at best 2 meters. To digitalise the maps, first was rasterised each geological map presented in Fig. 3 throughout Fig. 13 (Fig. 2b). The coastline from high-definition model presented for Antarctica by Gerrish et al. (2021) was used as reference guide and is part of the data presented in the SCAR Antarctic Digital Database, accessed in May 2022.
3. Geological maps were fully digitalised within a shapefile polygon type, which includes 321 different polygons (Fig. 2c). This shapefile has the following fields: (i) Name, (ii) Island, (iii) Formation, (iv) Age, (v) Environment, (vi) Type, (vii) Age Max, (viii) Age Min and (ix) Comment. This information it was used to generate maps for each island, which are present in the Supplementary Material.
4. Finally, the scale of the maps here presented are noteworthy. While the archipelago has an extension of more than 250 km, the largest outcrops have a dimension of a few kilometres long and wide. This results in local maps ranging in 1:0.5 km to 1:5 km in scale, which are often not possible to properly be seen at the scale of the archipelago, which is mostly covered by the icecap. Therefore, this work should be considered as a compilation of local geological maps, rather than a consolidated unique map.

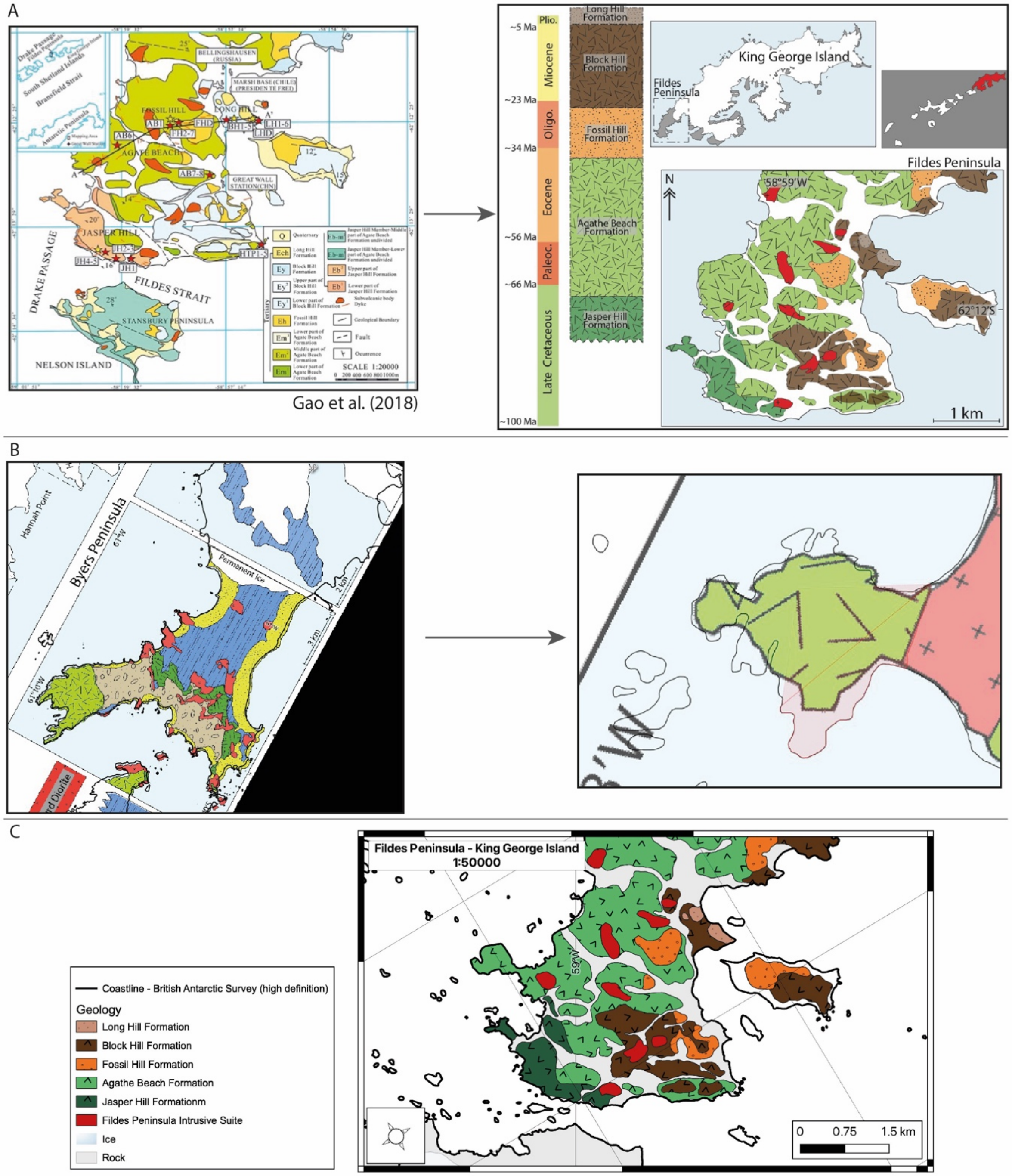

Figure 2. Used methodology. (A) shows the development of a geological map from the work presented in the literature (e.g. Gao et al., 2018). (B) Rasterization and digitalisation in QGIS v3.24 of the units in the polygon shapefile containing the geological information. (C) Final map showing the geological units layout of Fildes Peninsula, King George Island, the rest is presented in the Supplementary Material.

4. Geological units

The geology of the archipelago is presented here from southwest to northeast.

4.1. Low Island, Cape Wallace: Jurassic deep marine sedimentation, Early Cretaceous volcanism and Valanginian plutonism

Located at the southern of the South Shetland Islands (Fig. 1), Low Island is almost entirely covered by ice. Cape Wallace is the exception, where about 4 km^2 of rock outcrop is exposed (Fig. 3). The exposures consist of a fine-grained sedimentary unit overlain by a volcanic succession, both of which are intruded by a granodiorite pluton (Araya and Hervé, 1966; Smellie, 1979; Bastias, 2014). Smellie (1979) first described the bedded sequence at Low Island, later redefined by Bastias and Hervé (2013) as the Cape Wallace Beds (Fig. 3). Bastias et al. (2019) defined two members for the Cape Wallace Beds, the Pencil Beach and the Albatross Hill members, which are equivalent to the 'sedimentary member' and 'volcanic member' defined by Smellie (1979), respectively. Further details of these units are presented below.

4.1.1. Pencil Beach Member: deep marine Jurassic sedimentation

Outcrops of the Pencil Beach Member are located in the southeast and the northeast parts of Cape Wallace (Fig. 3). This sequence is well-bedded and includes fine-grained volcaniclastic claystones, crystal tuffs and lapillistones. Individual beds are generally 4-30 cm thick, laterally uniform and continuous, and have been interpreted as turbiditic deposits (Smellie et al., 1984). Primary sedimentary structures are uncommon, except for some cross-lamination, disrupted beds, and graded bedding (Smellie, 1979). The strata strike between NW-SE and E-W, with a dip varying from sub-horizontal to 30°. Thomson (1982) reported in-situ trace and molluscan fossils that suggest a Late Jurassic age, with an ammonite identified as *Epimayites aff. transiens* (Waagen) corresponding to the Oxfordian stage. Castillo et al. (2015) and Bastias (2014) worked on the provenance (U-Pb in zircon) of this unit and indicated that the detrital material of Pencil Beach prominently Permian, which coincides with other turbidite-like sequences in the northern of the Antarctic Peninsula (the Triassic Trinity Peninsula Group).

4.1.2. Albatross Hill Member: subaerial Valanginian volcanism

The Albatross Hill Member unconformably overlies the Pencil Beach Member and consists of a succession of subaerial lava flows. Lavas and hyaloclastites are aphyric or feldspar-phyric augiteandesites, dacites, hornblende-andesites and rare basalts (Smellie, 1979). Both lithostratigraphic units are intruded by the Cape Wallace pluton (Smellie, 1979; Bastias, 2014), a granodiorite from which Smellie et al. (1984) obtained a whole-rock K-Ar age of 120 ± 4 Ma (Fig. 3). However, Bastias et al. (2019) presented $^{40}Ar/^{39}Ar$ plateau ages of ~140–136 Ma, which now constrain the age of this volcanic unit.

4.1.3. Cape Wallace Granodiorite: Valanginian acidic plutonism

This pluton comprises most of the outcrop on Cape Wallace Island and is a granodiorite associated with minor aplite bodies. This pluton intrudes the Pencil Beach and Albatross Hill members (Fig. 3). While Smellie et al. (1984) obtained a whole-rock K-Ar age of 120 ± 4 Ma, Bastias et al. (2019) reported an age of 137 Ma ± 2 from a U-Pb SHRIMP zircon age.

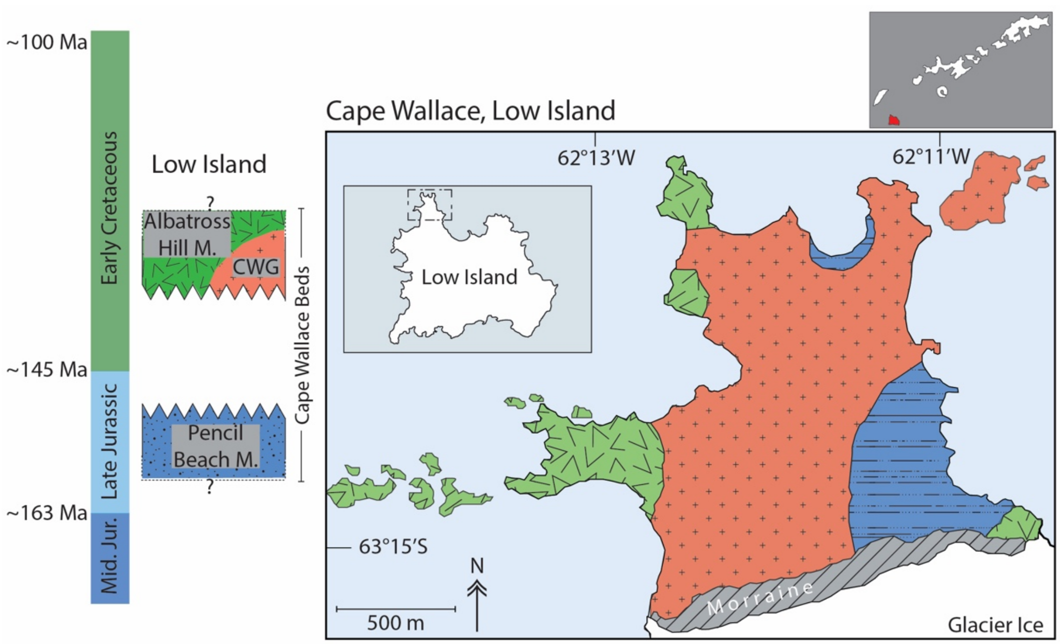

Figure 3. Revised geological map of Cape Wallace, Low Island, modified from Bastias (2014).

4.2. Smith Island: high pressure-low temperature Palaeocene–Eocene metamorphism

A group of blueschist/greenschist rocks crop out on Smith Island, a 40 km long and 8 km wide island. It has a mountainous interior (the Imeon Range) that extends the entire length of the island and has peaks over 2 km in elevation. The exposures are comprised mainly of green and blue phyllites with intercalations of metachert, grey phyllite and marble (Fig. 4; Dalziel, 1984; Grunow et al., 1992; Trouw et al., 1998a). Valeriano et al. (1997) on the base of chemical analyses suggested these rocks had a mid ocean ridge basalt protolith, along with hemipelagic sediments from an ocean floor environment (Grunow et al., 1992; Truow et al., 1998a). Grunow et al. (1992) presented $^{40}Ar/^{39}Ar$ ages of ~58–47 Ma in white mica and Truow et al. (1998b) determined PT conditions of 8 kbar, 300 - 350 °C. These authors interpreted these rocks as part of a subduction complex composed of ocean floor material mixed with arc-derived sediments. Bastias et al. (in review) suggested that the uplift of these rocks may have been associated with the resumption of the subduction during the Cretaceous. While the Palaeocene–Eocene age of deformation has been stablished by Grunow et al. (1992) for these rocks, the age of its protholith remains unknow. However, considering the late Palaeozoic and Mesozoic active margin evolution of the Antarctic Peninsula, we can speculate a Palaeozoic to Mesozoic age for the protolith.

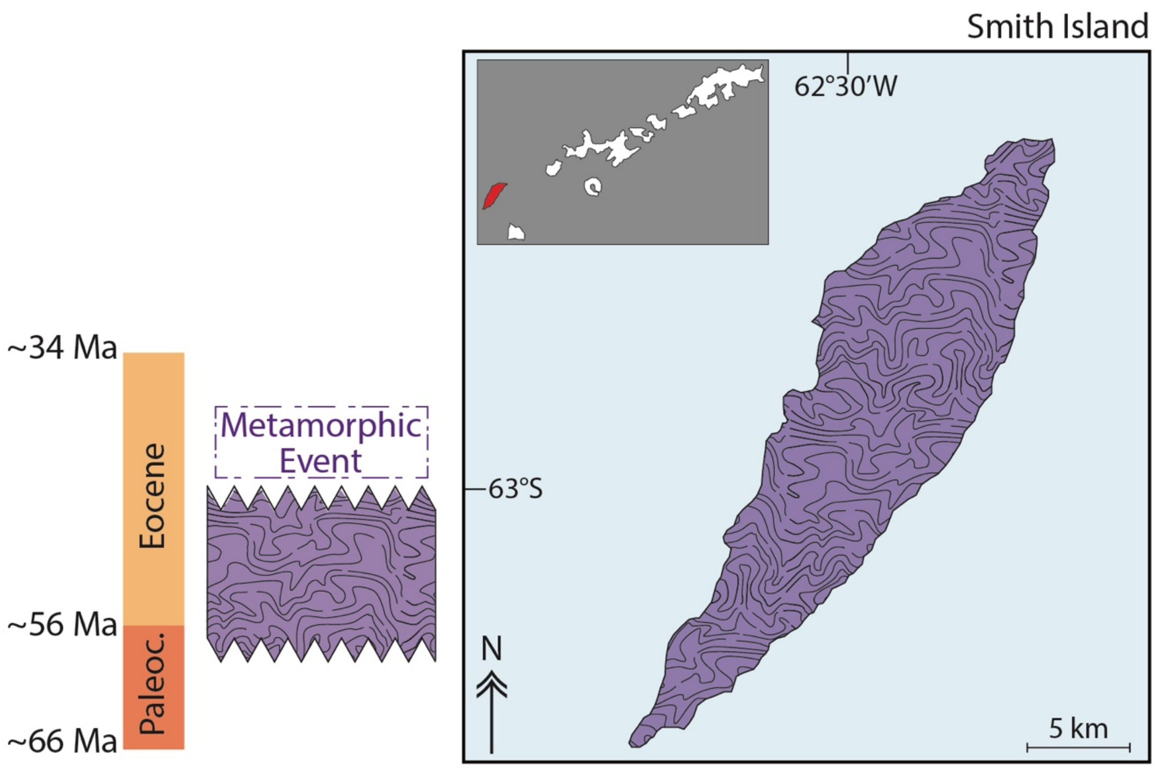

Figure 4. Revised geological map of Smith Island, modified from Grunow et al. (1992).

4.3. Snow Island: Early Cretaceous marine sedimentation, subaerial Late Cretaceous volcanism and Eocene plutonism

The outcrops of Snow Island are mostly restricted to President Head Peninsula (Fig. 5), ~10 km south of Byers Peninsula on Livingston Island (Fig. 7), although other minor shore outcrops are also present to the west and east on Snow Island. On President Head Peninsula, two distinctive units crop out -

a sedimentary unit overlain by a volcaniclastic unit (Fig. 5; e.g. Israel, 2015; Bastias et al., 2019). Due to their proximity and based on litho-, bio- and tectono-stratigraphic similarities, Crame et al. (1993) suggested correlations between these units and those present on Livingston Island to the north (section 4.5, Fig. 7). Several authors have suggested that the marine ammonite-bearing sedimentary succession at the western end of President Head could be correlated with the Sealer Hill Member of the Chester Cone Formation of Livingston Island (Crame et al., 1993; Torres et al., 1997; Hathway and Lomas, 1998; Israel et al., 2015). Based on the marine invertebrate assemblages, the sedimentary succession on President Head has been assigned to the mid-Valanginian (Ugalde et al., 2013; Israel, 2015). Plant remains found in the terrestrial volcaniclastic rocks at the top of the President Head succession on Snow Island show the emergence of the arc (Torres et al. 1997; Cantrill, 1998; Hathway and Lomas, 1998). Furthermore, Bastias et al. (2019) presented a U-Pb zircon age of 109 Ma ± 1 from the upper volcaniclastic succession at President Head, constraining its age to Aptian–Albian. This age was used to refine lithostratigraphic correlations with the units on Livingston (section 4.5, Fig. 7) and Low (section 4.1, Fig. 3) islands and to propose a common evolution within the Byers Basin.

Intrusive bodies constitute a major part of the outcrops on President Head, Snow Island (Fig. 5). They are mostly dykes, hypabyssal intrusives of basaltic to andesitic composition (e.g. Valenzuela and Hervé, 1972; Smellie, 1980, Pankhurst and Smellie, 1983) and minor columnar basalts and dacites (Pankhurst et al., 1979). The geochronological control of this intrusive unit is limited to the age obtained by Pankhurst and Smellie (1983), who obtained a K-Ar age of 46 Ma ± 2 from a dacite of President Head Peninsula. Nevertheless, most of these rocks have experienced low-temperature alteration (Smellie et al., 1980) and likely the daughter isotope is disturbed, as has been shown on other islands of the archipelago (Bastias et al., 2016; Bastias et al., 2019).

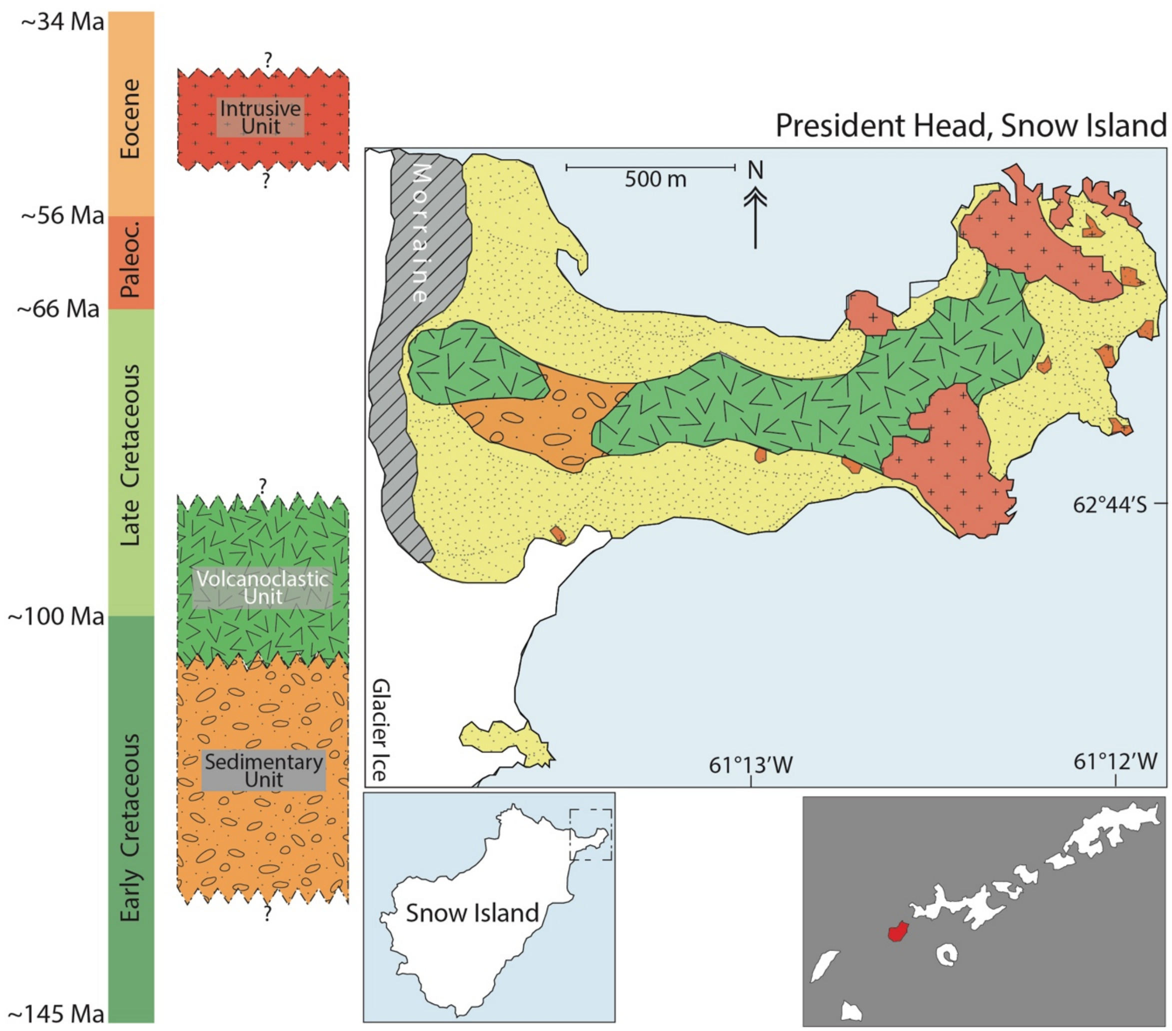

Figure 5. Revised geological map of the President Head Peninsula, Snow Island, modified from Israel (2015).

4.4. Deception Island: Quaternary volcanism

This island is amongst the most active volcanoes in Antarctica with a record of over 20 explosive eruptions in the last two centuries (e.g. Smellie, 2001). It consists of a composite volcano with a basal diameter of 30 km and rising over 500 m above sea level (e.g. Luzón et al, 2011). The island has a horseshoe-shape, whose submerged central portion represents a collapsed caldera (Fig. 6). The normal magnetic polarity of the rocks exposed on Deception Island indicates that the rocks exposed there are younger than 0.75 Ma (Valencio et al., 1979) while K-Ar data suggest that most of the subaerial part of the island was built in the last 0.2 Myr (Keller et al., 1992). This active volcanic island has attracted several volcanologists, which have greatly contributed to a better understanding on its age and evolution (e.g. González-Ferrán et al., 1971; Orheim, 1972; Roobol, 1982; Ibáñez et al., 2003; Almendros et al., 2018; Geller et al., 2017, 2019) along with a detailed geological map (Smellie et al., 2002). Nevertheless, with regards this study, Deception Island comprises part of the Pleistocene units of the South Shetland Islands. Although a considerable amount of detailed information has been obtained on the Quaternary geology of this island, it is not useful for the purposes of regional lithostratigraphical correlations across the South Shetland archipelago.

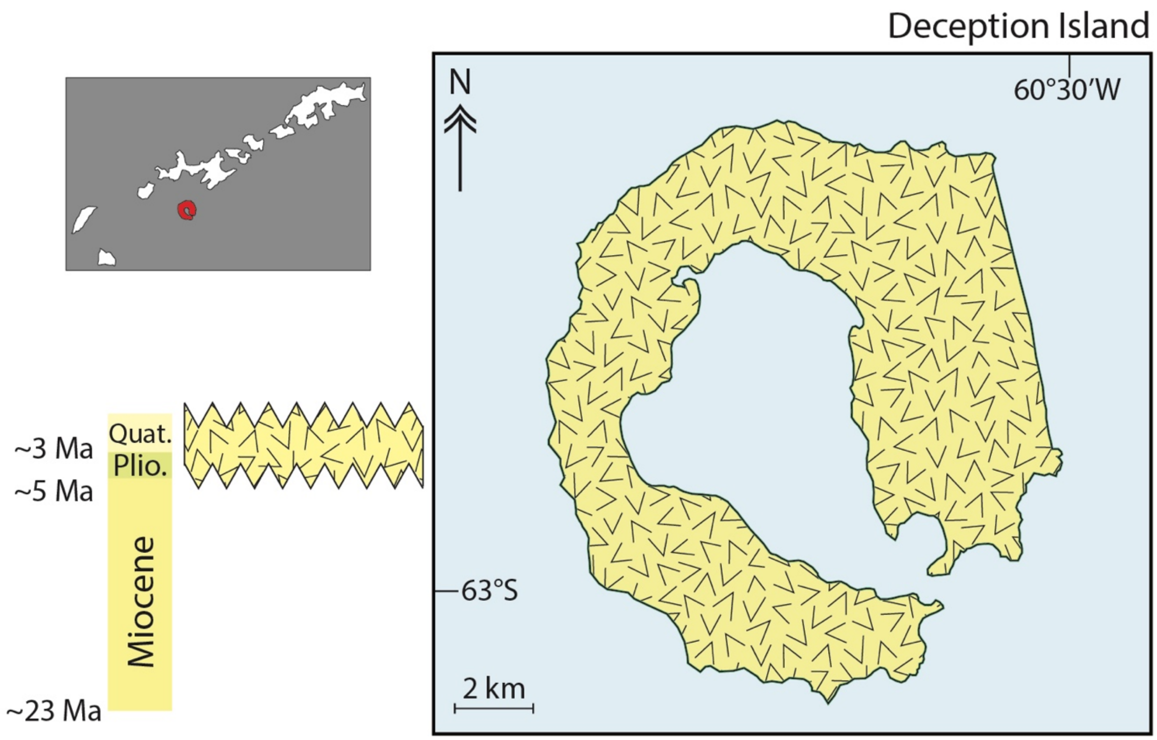

Figure 6. Geological map of Deception Island.

4.5. Livingston Island

Most of the rock exposures on Livingston Island are restricted to the Byers Peninsula (Fig. 7), where the Byers Group crops out (Hathway and Lomas, 1998). The base and the top of the group are not exposed, but the overall thickness is estimated to be at least 2.7 km. Contact relations between the formations are locally obscured by high-angle faults, folding and of younger hypabyssal and plutonic intrusions (e.g. Hathway and Lomas, 1998); others are covered by the ice cap. Hathway (1997) and Hathway and Lomas (1998) reviewed the Byers Group lithostratigraphy originally proposed by Smellie et al. (1980) and Crame et al. (1993), and is presented here in stratigraphic order.

4.5.1. Byers Peninsula

4.5.1.1. Anchorage Formation: Late Jurassic deep marine sedimentation

First defined by Crame et al. (1993), the outcrops of the Anchorage Formation are confined to an area of 0.5 km^2 in the northwest sector of Byers Peninsula (Fig. 7), where they form a homoclinal succession dipping to the north-west and are intruded by tabular microdioritic bodies. Hathway and Lomas (1998) subdivide the Anchorage Formation into the New Plymouth and Punta Ocoa members, which reach a combined thickness of about 120 m. Although the members are lithologically distinguishable and differ markedly in the degree of bioturbation, these two members are not differentiated on the map presented here. Likewise in Deception Island, this detailed information is not useful for the purposes of regional lithostratigraphical correlations across the South Shetland archipelago. The New Plymouth Member consists mainly of mudstones that exhibit notable bioturbation and numerous silicic tuff layers, sediments are normally graded and locally bioturbated. The Punta Ocoa Member is mainly composed of radiolarian-rich shale with volcanic ash tuff layers and fine layers of massive sandstone with mudstone intraclasts (Pirrie and Crame, 1995; Hathway and Lomas, 1998). The fossil record of the Anchorage Formation suggests a Late Jurassic age (Kiessling et al., 1999). The mudstones can be mostly interpreted as deep hemipelagic deposits but some may have been deposited by small turbidity currents. The sandstones and some tuffs are interpreted as turbidites (Pirrie and Crame, 1995) and show a limited volcaniclastic input from the Late Jurassic magmatic arc. Bastias et al. (2019) presented a zircon U-Pb age of 153 ± 2 Ma from an ash-layer of the Anchorage Formation, complementing the age control suggested by the fossil record of this unit.

4.5.1.2. President Beaches Formation: Berriasian (~145–140 Ma) marine fan sedimentation

The President Beaches Formation was defined by Crame et al. (1993) and redefined by Hathway and Lomas (1998), where it delimits an area extending 0.5 to 1.5 km inland from Byers Peninsula and estimate a thickness of at least 600 m (Fig. 7). It overlies the Anchorage Formation in apparent unconformity. The President Beaches Formation consists mainly of dark grey shales with pyrite, radiolarians and isolated trace fossils. Minor presence of yellowish clay-rich strata 0.5–4 cm thick (interpreted as altered tuffs), layers of grey-green sandstones and centimetric carbonate concretions. There are also lenticular bodies of sandstone of greater thickness (8–45 m) with small-scale syn-sedimentary deformation (load structures, flame structures, convolute lamination, clastic dykes) likely caused by the submarine relief and high sedimentation rates (Lomas, 1999). Hathway and Lomas (1998) proposed that the President Beach Formation was probably deposited on a submarine slope fan below the base level of storm waves, where mud deposition was sporadic and locally interrupted by sandy turbidite flows confined to small channels. The source region may have been a volcanic arc with vegetation located to the southeast (Lomas, 1999). The rocks contain macrofossils of Berriasian affinity, including ammonites (*Spiticeras, Blanfordiceras, Himalayites, Bochianites;* Smellie et al., 1980), bivalves (*Manticula* Waterhouse, *Praeaucellina, Inoceramus*; Crame et al., 1993) and belemnites (Smellie et al., 1980; Crame et al., 1993, Pirrie and Crame, 1995). Duane (1994, 1996) further described the presence of dinoflagellate assemblages that would indicate a mid-to-late Berriasian age (~145 Ma). This formation does not have a reported geochronological control.

4.5.1.3. Start Hill Formation: Valanginian (~140–133 Ma) shallow marine fan sedimentation and basaltic subaerial volcanism

The Start Hill Formation was defined by Hathway and Lomas (1998), and has an estimated thickness of at least 265 m. The top is unknown, while its base is well defined and is in apparent discordant contact with the mudstones of the President Beaches Formation (Fig. 7; Valenzuela and Hervé, 1972; Smellie et al., 1980). The formation is mainly composed of massive, poorly sorted volcanic breccias and both clast- and matrix-supported conglomerates that form very thick strata (5–30 m). The clasts, mostly smaller than 30 cm but up to 3 m in diameter and generally angular, are of basalt and basaltic andesite with porphyritic and amygdaloidal textures. The contacts between stratal packages are mainly gradational with pronounced lateral variation. There are also geographically restricted volcanic agglomerates of grey and brown, finely-laminated, lapilli tuffs and lavas. Combined with the presence of a molluscan assemblage dominated by oyster shell fragments (Smellie et al., 1980), this formation has been interpreted as part of a fan of submarine debris that surrounded one or more basaltic subaerial volcanoes. While Hathway and Lomas (1998) assigned a Berriasian age (~145–140 Ma), Haase et al. (2012) suggested a Valanginian age based on $^{40}Ar/^{39}Ar$ dating of aphyric lavas that yielded an age of 135 ± 3 Ma.

4.5.1.4. Chester Cone Formation: Valanginian (~140–133 Ma) marine-continental transitional sedimentation and volcanism

The Chester Cone Formation was defined by Crame et al. (1993) and redefined by Hathway and Lomas (1998), who divided it into two members: the basal Devils Point Member, previously included at the top of the Anchorage Formation by Crame et al. (1993), and an upper Sealer Hill member. We nevertheless do not separate this formation into two units on the geological map (Fig. 7), with the aim of maintaining consistency on the level of detail on the maps presented. The Devils Point Member is wedge-shaped with an estimated thickness of 300 m and has been affected by numerous faults. Its base is composed of matrix- to clast-supported pebble conglomerates, of mottled appearance and grey-orange colour, with well-rounded clasts of basaltic to intermediate volcanic rocks and beige-grey mudstone intraclasts. The upper part consists of conglomerates of gravel and thick greyish sandstones with fossil remains. The fossil content is dominated by thick-shelled bivalves that appear to be reworked (Crame et al., 1993), probably by turbidity currents that remobilised coastal gravels (Duane, 1996). The Devils Point member is intruded by the Chester Cone andesite plug, for which Pankhurst et al. (1979) obtained a Valanginian K-Ar amphibole age of 132 ± 4 Ma. The contact of the Sealer Hill member with the Devils Point Member is gradational (Crame et al., 1993; Hathway, 1997; Hathway and Lomas, 1988). The Sealer Hill member consists of dark, finely laminated, fairly bioturbated shales, millimetric to centimetric layers of fine-grained graded sandstone that sometimes reach greater thicknesses and exhibit parallel to wavy low-angle stratification, and rare one-metre thick strata of massive and poorly sorted conglomerates, with pebble-size basaltic clasts in a gravel-to-sand matrix. The shales might be low-energy suspension deposits while the sandstones and conglomerates might be the product of storm deposits (Hathway and Lomas, 1998). An ammonite assemblage with species of *Bochianites, Uhligites and Neocomites,* indicates a Valanginian age (Covacevich, 1976), while belemnites (*Belemnopsis (Belemnopsis) alexandri* Willey and *B.(B.) gladiatoris* Willey (Crame et al., 1993), plants and bivalves are also present. Duane (1996), by means of palynology, defined the maximum age as Valanginian.

4.5.1.5. Cerro Negro Formation: late Early Cretaceous volcanism

The Cerro Negro Formation (Hathway, 1997) encompasses all the continental volcaniclastic strata exposed on the eastern side of Byers Peninsula (Fig. 7). It has an approximate thickness of 1.4 km, measured in a composite type section, and dips gently to the ENE. The top is hidden under the present ice cap. Hathway (1997) informally divided the Cerro Negro Formation into two members separated by an abrupt and striking colour change, which may represent a diagenetic front rather than a primary stratigraphic colour change. The first member consists of 200 to 240 m of silicic volcaniclastic rocks of pale green to grey colour (a weathering effect), and silicic welded or non-welded ignimbrites intercalated with reworked and silicic pyroclastic material. The second member is composed of lapilli tuffs and tuff breccias, mostly syn-eruptive and of dark red–purple colour (due to the predominance of basaltic rocks). The lapilli tuffs and tuff breccias are interpreted as debris flows and flood flows, respectively. Changes in thickness and facies support a syn-sedimentary displacement along normal and south-vergent ENE-trending faults (Hathway, 1997). These rocks contain abundant macro- and micro-fossil vegetation of various plant groups of late Aptian age (Torres et al., 1997; Duane, 1996; Cantrill, 2000). $^{40}Ar/^{39}Ar$ analysis of plagioclase from a basal tuff yielded an age of 120.3 ± 2.2 Ma (Hathway, 1997; Hathway et al., 1999). Ignimbritic clasts in a conglomerate unit 140 m below the top of the succession yielded $^{40}Ar/^{39}Ar$ ages of 119.4 ± 0.6 Ma and 119.1 ± 0.8 Ma on biotite and plagioclase, respectively (Hathway et al., 1999), suggesting rapid deposition during the latest Aptian. Many younger K-Ar ages have been obtained from the upper part of the Cerro Negro Formation including the intrusive rocks at Negro Hill, ranging from ~110 Ma to 80 Ma (Pankhurst et al., 1979; Smellie et al., 1984). Some authors consider the possibility of placing the Cerro Negro Formation in a stratigraphic framework outside the Byers Group, and thus denoting its current status within the Byers Group as provisional (Hathway, 1997; Hathway and Lomas, 1998). Considering the current international stratigraphic conventions, we retain the unit in the Byers Group.

4.5.2. Late Cretaceous volcanism in Hannah Point

Hannah Point is located on the south-central coast of Livingston Island (Fig. 7) and forms part of the widespread occurrence of volcanic rocks on the central-east side of the island. With a thickness of ~500 m of layered rocks, Pallàs et al. (1999) recognised five members within this succession, from base to top: (i) 120 m of polymictic volcaniclastic breccias, (ii) 70 m of volcaniclastic breccias, (iii) 65 m of basaltic lavas, (iv) 65 m of volcaniclastic breccias and (v) 150 m of andesitic lavas, suggesting that this succession were emplaced as pyroclastic flows associated with explosive volcanic activity in a subaerial environment. The age has been determined by a combination of palaeontology and geochronology. On the base of leaf imprints and fossil trunks, Leppe et al. (2007) suggested a Late Cretaceous age. This was confirmed by Haase et al. (2012) based on $^{40}Ar/^{39}Ar$ whole rock dating which yielded an age of 98 ± 1 Ma. Later, Bastias et al. (2016) studied the ubiquitous presence of low-temperature mineral associations in Hannah Point and attributed these assemblages to a combination of burial and hydrothermal alteration.

4.5.3. Hurd Peninsula: Jurassic marine sedimentation, Cretaceous acidic plutonism and Palaeogene hypabyssal mafic plutonism

Located to the southeast of Livingston Island, Hurd Peninsula hosts the Miers Bluff Formation, which is considered a multiply deformed and metamorphosed turbidite sequence (Hobbs, 1968; Smellie et al., 1984; Willan et al., 1994) unconformably overlain by Cretaceous volcanic strata (Fig. 7; Caminos et al., 1973; Smellie et al., 1984, 1995). These units are cut by widespread mafic dyke swarm (Zheng et al., 2003). Hervé et al. (2006) conducted geochronological studies in the peninsula that indicated a Middle to Late Jurassic age for Miers Bluff Formation based on zircon U-Pb provenance studies. Additionally, they also reported a 138 ± 1 Ma zircon U-Pb age for a granodiorite, constraining the felsic plutonism to the Valanginian. This was later used by Bastias et al. (2019) to suggest the presence of an Early Cretaceous arc-related belt of plutons on the South Shetland Islands. Finally, Zheng et al. (2003) based on K-Ar and $^{40}Ar/^{39}Ar$ ages, suggested that the mafic dike swarm was emplaced from ~79 to 31 Ma.

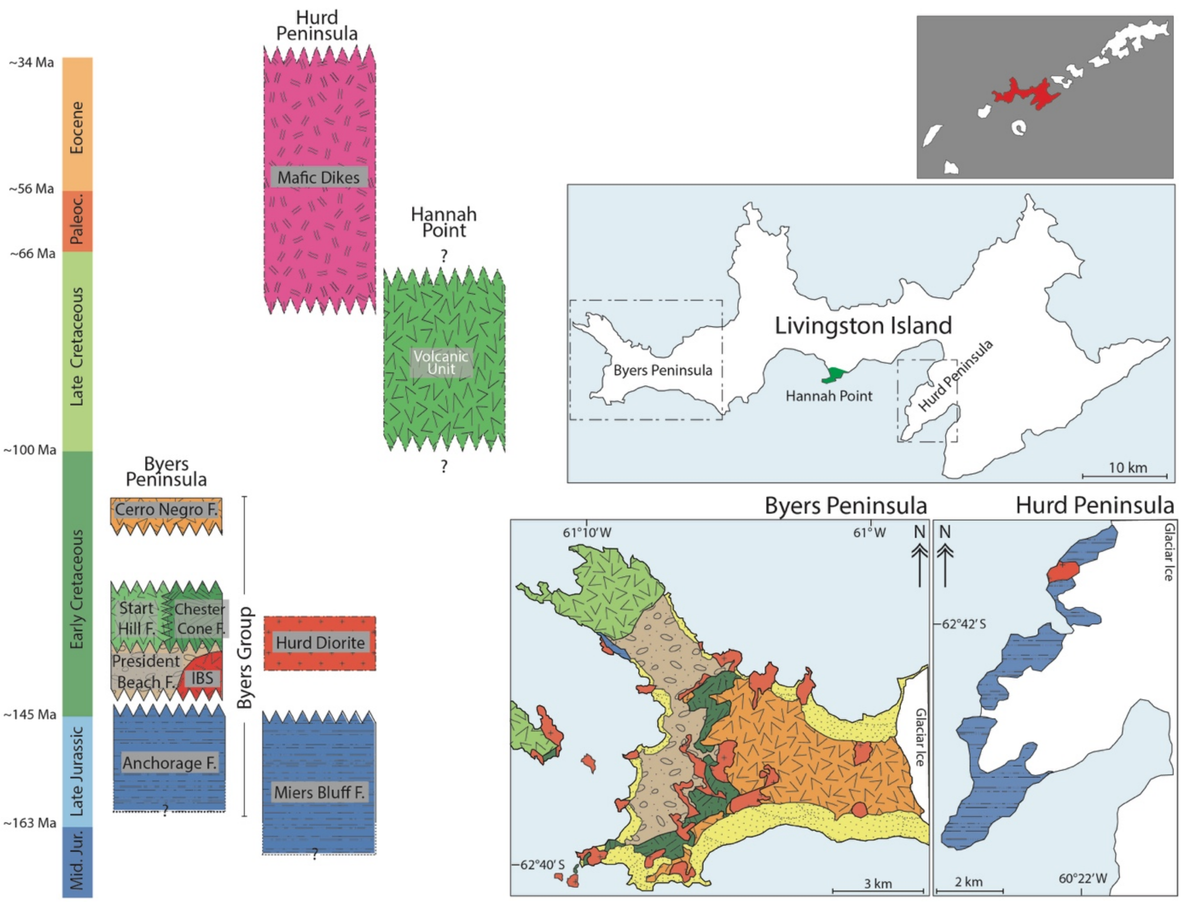

Figure 7. Revised geological map of Byers and Hurd peninsulas, Livingston Island, modified from Hathway and Lomas (1998) and Hervé et al. (2006).

4.6. Greenwich Island, Fort Point: Late Cretaceous volcanism and plutonism

The Fort Point exposures crop out on the east coast of Greenwich Island (Fig. 8), and is comprised of volcanic rocks (basalts, basaltic andesites, and andesites) and plutonic rocks (granites, tonalites, diorites and gabbros) (e.g. Machado et al., 2005). Grikurov et al. (1970) reported a ~105 Ma K-Ar age on a tonalite from the central part of the island (uncertainty was not reported), while Smellie et al. (1984) documented a K-Ar age of 80 ± 2 Ma for a basalt sill.

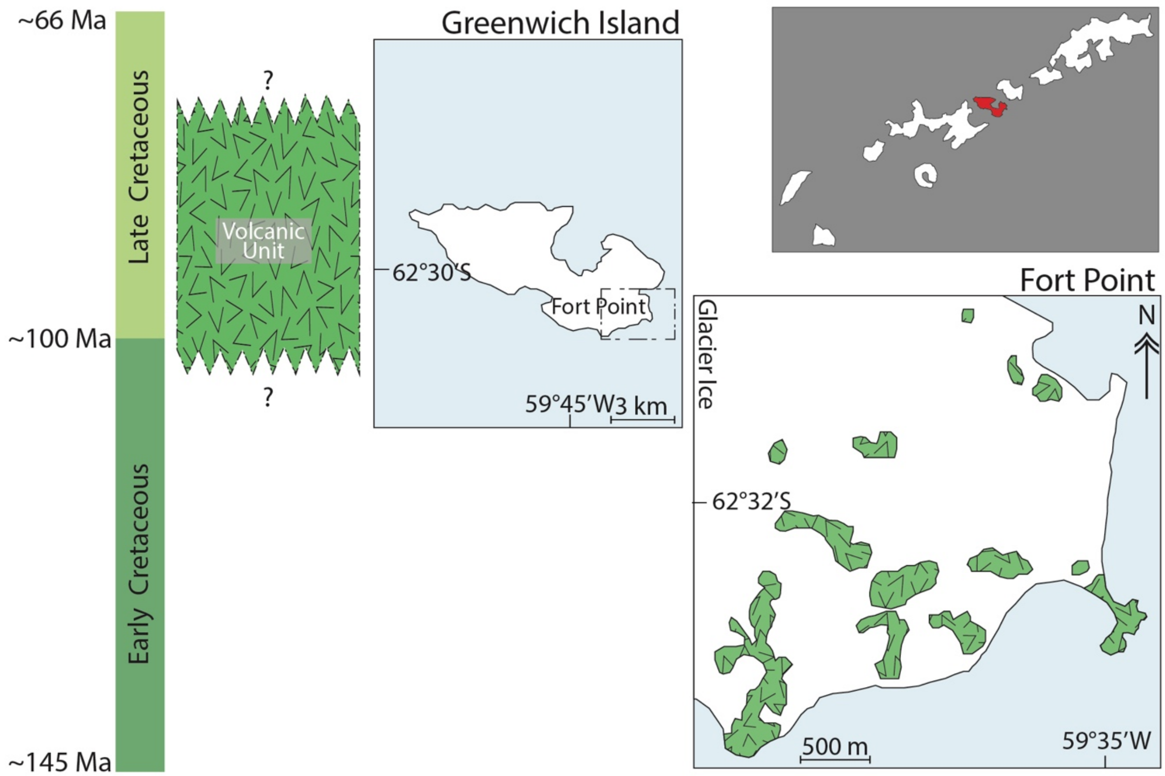

Figure 8. Geological map of Fort Point, Greenwich Island, from Machado et al. (2005).

4.7. Robert Island, Coppermine Peninsula: Late Cretaceous volcanism

Located within the central part of the archipelago, Robert Island is 18 km long and 18 km wide, and is mostly covered by an ice-cap (Fig. 9). While some regions along the coastline have rock exposures, its largest outcrop is found on the Coppermine Peninsula on the southwest coast (Fig. 9). These rocks form part of the Coppermine Formation, which is dominated by basalts and andesitic-basaltic agglomerates (e.g. Caballero and Fourcarde, 1959; Hervé and Araya, 1965; González-Ferran and Karsui, 1970). Smellie et al. (1984) reported K-Ar ages of ~83 to 78 Ma from the Coppermine Formation. Recently, Villanueva (2021) undertook a compilation study of the geology of Coppermine Peninsula and reports $^{40}Ar/^{39}Ar$ plateau ages between~81 and 62 Ma.

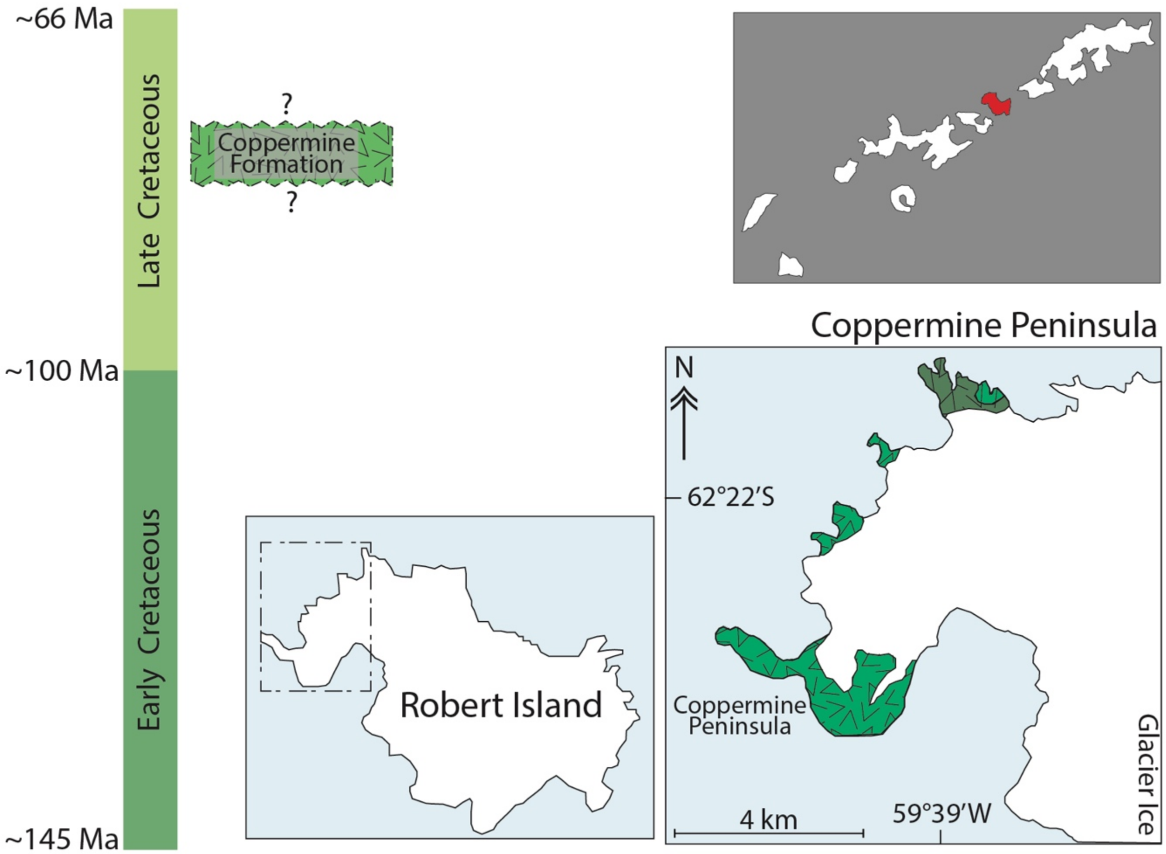

Figure 9. Geological map of Coppermine Peninsula, Robert Island, from Machado et al. (2005).

4.8. King George and Nelson islands

Nelson and King George islands are only separated by a narrow channel of ~500 m wide and share a common geological evolution. King George Island is the largest of the South Shetland Islands and given its relative ease access, several geological studies have been subjected in its rock exposures. This has resulted in a complex stratigraphy that has been supported mainly by K-Ar geochronology (e.g. Leat and Riley, 2021). The exposures in King George Island are dominated by Cenozoic volcanic and sedimentary strata (e.g. Birkenmajer and Zastawniak, 1989; Dutra and Batten, 2000; Poole et al. 2001; Fontes and Dutra, 2010; Warny et al., 2019; Smellie et al., 2021a, 2021b). These exposures are relevant because they record the setting and evolution of the icecap of Antarctica, where a few late Cenozoic glacial episodes have been postulated on the rocks cropping out at the northeast of the King George Island (e.g. Warny et al., 2019; Smellie et al., 2021a, 2021b).

4.8.1. Nelson Island and Fildes Peninsula: Late Cretaceous to Palaeocene volcanism and sedimentation

Situated on southwestern King George Island, the bedrock geology of the Fildes Peninsula (Fig. 10) consists of a suite of subalkaline volcanic, plutonic and volcaniclastic rocks (Zheng et al., 1991). Early work indicated the presence of an abundance of volcanic rocks with Cenozoic and Mesozoic fauna (Hawkes, 1961; Barton, 1965). Smellie et al. (1984) defined the rocks of the peninsula into the Fildes Peninsula Formation, comprised of three members, (i) a lower member of coarse porphyritic basalt and basaltic andesite interbedded with laterally discontinuous volcanic rocks, (ii) a middle member of volcaniclastic rocks with minor basalt and basaltic andesite lavas and (iii) and an upper member of fine-grained aphyric and microporphyritic andesite and dacite lavas. Furthermore, Smellie et al. (1984) reported K-Ar whole-rock ages between ~58–42 Ma. Later, Machado (1997) separated these rocks into four formations from bottom to top: (i) the Clement Hill Formation, composed of basalt and andesites interbedded with polymictic volcanic breccias, (ii) the Fildes Strait Formation comprised of trachybasalts and porphyritic basalts associated with volcanic breccias, (iii) the Schneider Bay Formation composed of porphyric basalts, andesites and dacites interbedded with breccias and (iv) the Winkel Point Formation, which is comprised of basalts and basaltic andesites interbedded with volcanic breccias, agglomerates, conglomerates and tuffs.

Alternatively, a group of contributions employing a different stratigraphy for Fildes Peninsula has been proposed by Li et al. (1992, 1996), Li and Liu (1987), Liu and Zheng (1988), Zheng et al (1991) and Gao et al. (2017). These studies have divided the rocks of the peninsula into five formations which are Jasper Hill, Agathe Beach, Fossil Hill, Block Hill and Long Hill and this is the stratigraphic nomenclature adopted herein. At the base, the Jasper Formation is mostly exposed on the southwestern tip of the Fildes Peninsula and is composed of basaltic and basalt-andesitic lavas and breccias. The overlying Agathe Beach Formation consists of amygdaloidal basalt, basalt-andesitic lavas and volcanic breccias. The Fossil Hill Formation overlies the Agathe Beach Formation and consists of volcanic breccia, tuff and pyroclastic-sedimentary rocks intercalated with several fossiliferous layers. Fossils of middle Eocene ages found in Fossil Hill Formation include pollen spores, plant leaves and stems and bird footprints (Li, 1994; Li and Shen, 1990; Song, 1998). The overlying Block Hill Formation is exposed on the east of the peninsula and consists of basaltic andesite, breccia and agglomerate. Finally, the Long Hill Formation is restricted to the eastern part of the peninsula, and consists mainly of basaltic and basalt-andesitic subvolcanic rocks. The reliability of the geochronological results presented in the archipelago is affected by the ubiquitous mineral alteration of the volcanic rocks on the archipelago (e.g. Bastias et al., 2016). However, Gao et al. (2017) compiled and updated the geochronological information of Fildes Peninsula to slightly modify the stratigraphy presented by Li et al. (1996), which is presented herein and can be used to construct a succession of subaerial volcanic and volcaniclastic strata of Late Cretaceous to Pliocene age.

The geological history of Nelson Island is closely shared with King George Island, particularly with Fildes Peninsula, which is the north continuation of Nelson Island. Furthermore, Li et al. (1996) presented a common lithostratigraphy, indicating that the units exposed in Nelson Island correspond to Jasper Hill and Fossil Hill formations. However, it is not clear why Agathe Beach formation is not present, which should be stratigraphically between these two units. Regardless, the geology is presented in the Fig. 11.

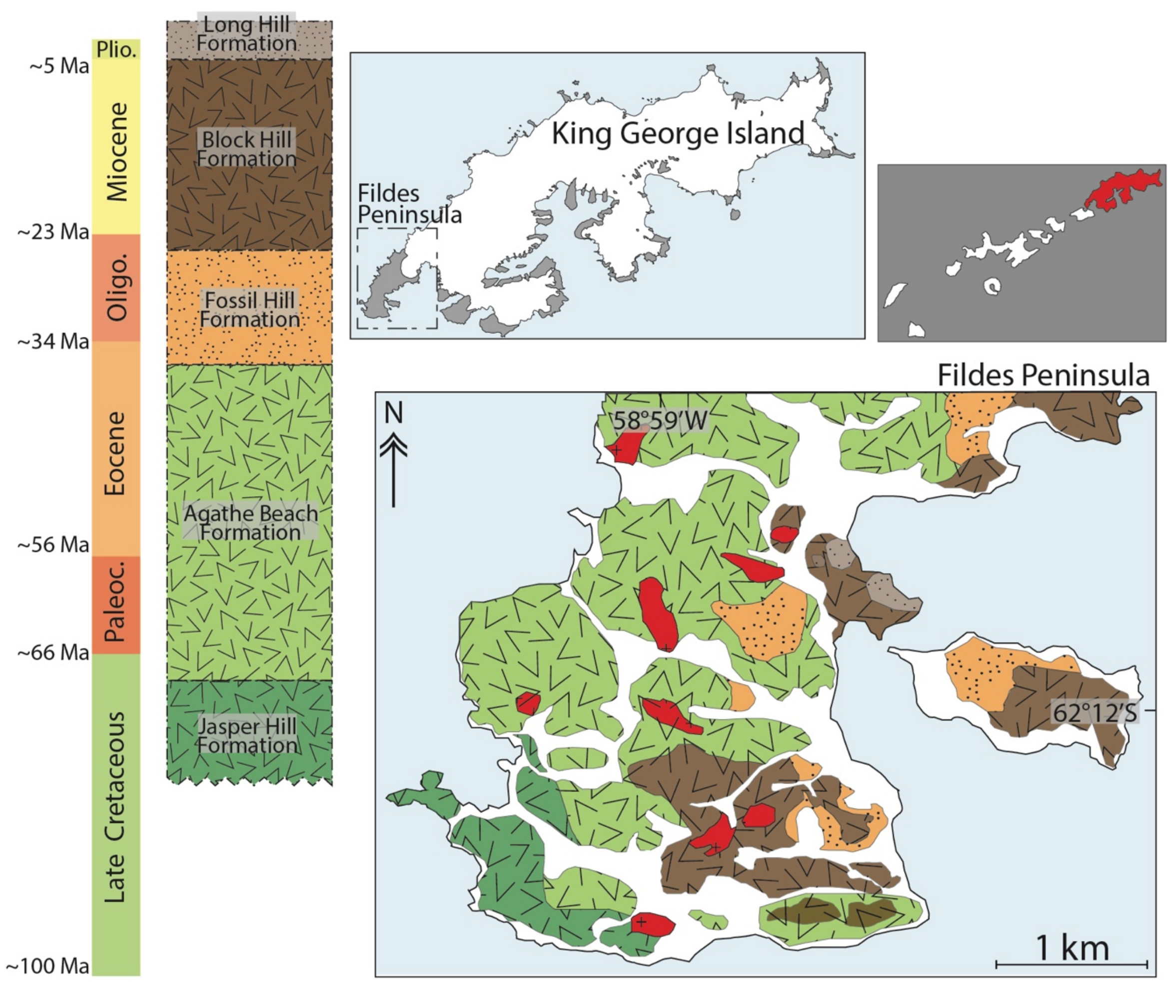

Figure 10. Geological map of Fildes Peninsula, King George Island, modified from Li et al. (1996).

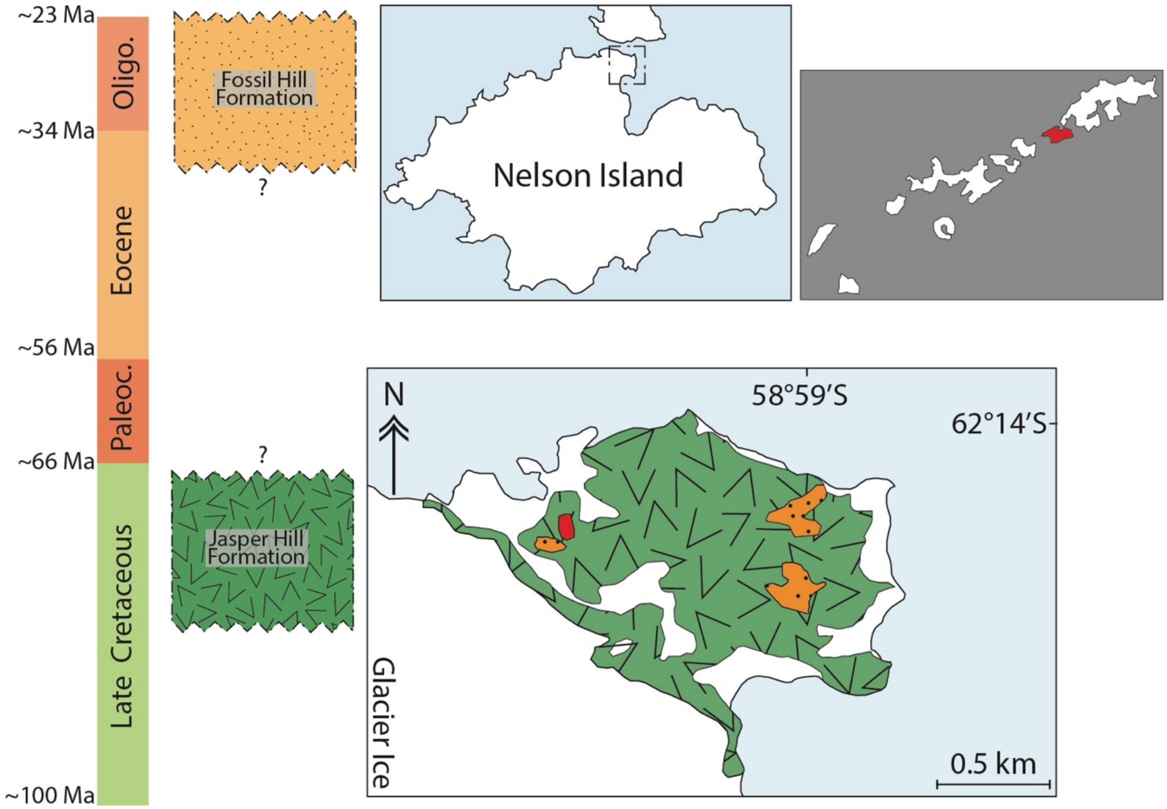

Figure 11. Geological map of Nelson Island, modified from Li et al. (1996).

4.8.2. Barton Peninsula: Palaeocene to Eocene volcanism and Eocene plutonism

The Barton Peninsula is dominated by volcanic and plutonic rocks (Fig. 12). The exposures of the volcanic rocks cover most of the peninsula and range in composition from basalt to andesite (e.g. Hwang et al., 2011). The Sejong Formation outcrops along the southern coast of the peninsula (Fig. 12; e.g. Yoo et al., 2001). This formation is largely composed of lapilli tuffs and volcanic breccias with a maximum thickness of 100 m and gently dip to the south/southwest. Overlying the Sejong Formation are mafic to intermediate volcanic lavas, which are widespread on the peninsula and range from basalt to andesite in composition (e.g. Hwang et al., 2011). Although the presence of alteration minerals prevents constraining the precise age of the lavas (Davies, 1982; Armstrong, 1995; So et al., 1995; Hur et al., 2001; Willan and Armstrong, 2002; Hwang et al., 2011), most of the lavas seem to have erupted during the Palaeocene to Eocene (Park, 1989; Kim et al., 2000; Willan and Armstrong, 2002; Yeo et al., 2004). Calc-alkaline granodiorite and diorite plutons intrude the volcanic rocks in the northern Barton Peninsula, which have been interpreted to be mid-Eocene in age (Park, 1989; Lee et al., 1996; Kim et al., 2000). Furthermore, Kim et al. (2000) reported a $^{40}Ar/^{39}Ar$ age of ~49 ± 2 Ma from a fine-grained diorite.

4.8.3. Potter Peninsula: Palaeocene to Eocene volcanism and early Eocene plutonism

The peninsula is dominated by a volcanic sequence with an estimated thickness of ~90 m (Fig. 12). Kraus (2005) reported that the volcanic sequence on the Potter Peninsula is comprised of volcanic centres composed of lava flows, pyroclastic rocks (ash-fallout, pyroclastic flow deposits, volcanic breccia and agglomerates) and hypabyssal intrusions (dykes, sills and small subvolcanic intrusive bodies). These rocks range from basaltic to andesitic in composition and often exhibit columnar jointing. Only K-Ar ages have been reported from this location and range between ~58 and 42 Ma (Watts, 1982; Smellie et al., 1984), which agree with the stratigraphic interpretation of Kraus (2005). Furthermore, the age of the intrusive rocks is further constrained by the presence of mafic dykes on the Potter Peninsula, which yield $^{40}Ar/^{39}Ar$ ages between ~48 and 45 Ma (Kraus, 2005).

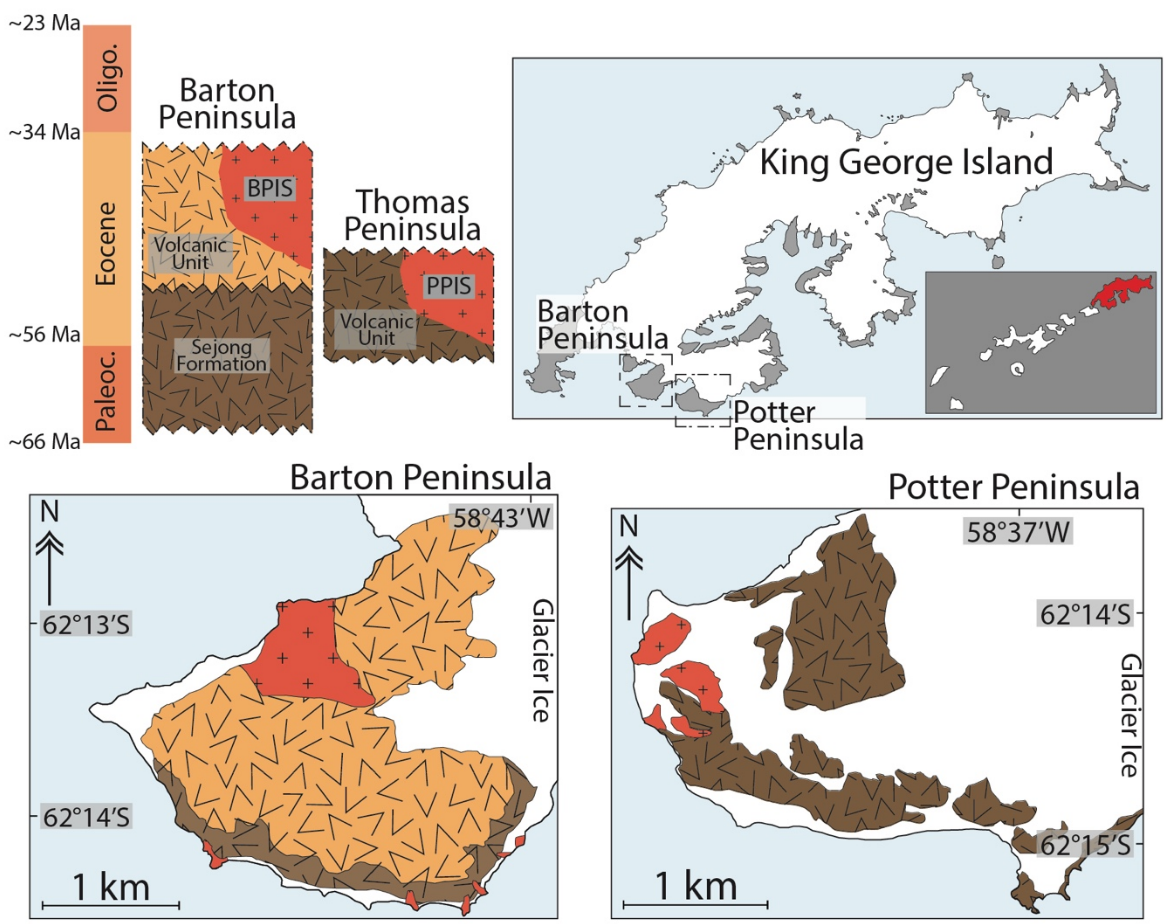

Figure 12. Geological map of Barton and Potter peninsulas, King George Island, modified from Kraus and del Valle (2008) and Lee et al. (2002).

4.8.4. Polonez Cove and surrounding area

Located in centre-west King George Island, the exposures in Polonez Cave and surrounding areas are dominated by the Chopin Ridge Group (Birkenmajer, 1980). This group of rocks include porphyritic lava flows and pyroclastics, marine tillites, associated glaciogenic sediments and sandstones. The rocks cropping out near Polonez Cove along with other minor exposures have been extensively reviewed by Smellie et al. (2021), and only a summary is presented here to facilitate comparison with the other units in the archipelago.

In the Polonez Cove region, three formations are recognised - the Mazurek Point, Polonez Cove and Boy Point formations (Fig. 13; Smellie et al., 2021). The Mazurek Point Formation is a subaerial volcanic sequence of Eocene age. While previous authors considered this unit as part of a larger marine-related group (Chopin Ridge Group; e.g. Birkenmajer, 1982), Smellie et al. (2021) suggested to not include this unit within that group considering its different age, features and origin. The Polonez Cove Formation is perhaps the most intensively investigated unit in the archipelago due to its association with an important glacial episode (Polonez Glaciation; e.g. Barton, 1965; Birkenmajer, 1980, 1982, 1987, 1995; Smellie et al., 1998, 2021; Troedson and Smellie, 2002; Quaglio et al., 2008, 2014; Warny et al., 2019; Nawrocki et al., 2021). The formation is exposed in a 2km-long cliff along with other exposures in the area. It is dominated by sandstones, conglomerates with abundant fossils and clear glacial features such as tillites, striations and roche moutonée landforms (Birkenmajer, 1995). These rocks are intercalated with volcanic units consisting of lava-fed deltas and breccias formed during submarine extrusion (Smellie et al., 1998). The depositional environment ranged from deep to shallow water conditions with a dominant mafic volcanic detrital component (Smellie et al., 2021). Although fossil material is abundant, much of it has been reworked from older strata (Smellie et al., 2021), and suggests an imprecise Oligocene palaeontological age (Gazdzicki and Pugaczewska 1984; Gaździcka and Gaździcki 1985; Birkenmajer and Gaździcki 1986). Ages on shell material derived from the global $^{87}Sr/^{86}Sr$ marine curve suggest an age of ~29 - 28 Ma (Dingle et al., 1997; Dingle and Lavelle, 1998), however alteration is widespread and these correlations with the global $^{87}Sr/^{86}Sr$ marine curve should be treated with caution. Recently, Smellie et al. (2021), presented new $^{40}Ar/^{39}Ar$ data on the Polonez Cove Formation, which yield plateau ages between ~28 and 26 Ma also suggesting an Oligocene age. Overlying the Polonez Cove Formation is the Boy Point Formation, composed of ~150 to 100 m of felsic-intermediate lavas and agglomerates, conglomerates and sandstones (Birkenmajer, 1982, 2001; Smellie et al., 2021). This unit formerly comprised two different formations (Birkenmajer, 1982, 2001) and has been redefined into one by Smellie et al. (2021). While the formation is dominantly composed of lavas, it is interbedded with volcanic-sourced sediments which are mostly comprised of sandstones and conglomerates (Smellie et al., 2021). Recently, Smellie et al. (2021) presented new $^{40}Ar/^{39}Ar$ data on the Boy Point Formation, which yield an age between ~27 and 25 Ma. This suggests that there is no significant time gap between the Polonez Cove and the Boy Point formations and deposition may have been essentially continuous (Smellie et al., 2021).

4.8.5. Melville Peninsula

Melville Peninsula is located in northeast King George Island. The rock exposures in the peninsula are comprised by the Sherratt Bay, Destruction Bay and Cape Melville formations along with the recent volcanism associated with the Melville Peak volcano (Fig. 13; e.g. Warny et al., 2015). The Sherratt Bay Formation represents an andesitic-basaltic succession, which crops out on the eastern edge of the Melville Peninsula. A stratigraphic hiatus separates the Sherratt Bay Formation and the overlying fossiliferous Oligocene Destruction Bay Formation (Birkenmajer, 1985). The Destruction Bay Formation consist of a ~100-40 m thick succession of volcaniclastic material with horizons rich in marine invertebrates (Birkenmajer, 1985). Overlying this formation is the Cape Melville Formation, which is composed of glacio-marine sediments including sandstones, conglomerates, clay-shales and silty shales with occasional ice-rafted dropstones (Birkenmajer, 1985). The dropstones often show glacial striae and glacially polished facets, thus giving primary evidence for a continental ice-sheet presence in Antarctica (e.g. Birkenmajer et al., 1985; Warny et al., 2015). Similar to the Polonez Cave Formation, the Cape Melvile Formation records a glacial event and thus it is one of the most studied geological units on King George Island. Furthermore, this glacial event has been correlated with the Mi-1 glaciation (e.g. Warny et al., 2015), an event that followed the shrinking of the icecap established during the Eocene-Oligocene by brief periods of interglaciation (e.g. Liebrand et al., 2011). Recently, Smellie et al. (2021), presented a $^{40}Ar/^{39}Ar$ age from a tuff within Cape Melville Formation and suggested an age of ~22 – 21 Ma for this formation. The formations on the Melville Peninsula are cut by a swarm of Miocene basaltic dykes (~20 Ma; Birkenmajer et al., 1985) and overlain by deposits associated with the Quaternary Melville Peak volcano (Birkenmajer and Keller, 1990).

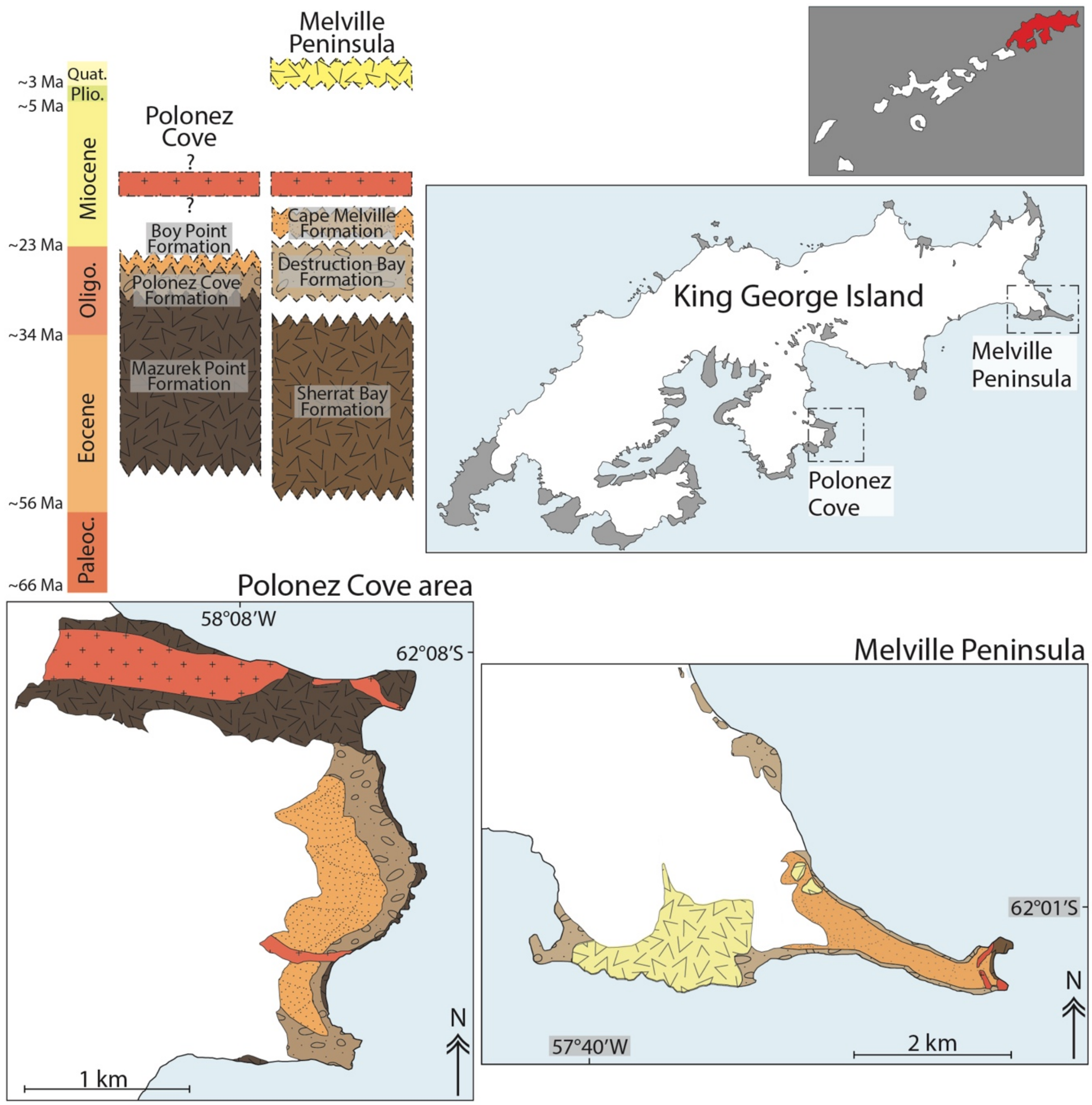

Figure 13. Geological map of Polonez Cove area and Melville Peninsula, King George Island, modified from Smellie et al. (2021) and Warny et al. (2015).

5. Lithostratigraphic correlations

5.1. Deep marine sedimentation: ~164 (?) – 140 Ma

The oldest geological units in the archipelago are Middle to Late Jurassic sedimentary rocks, which crop out on Low and Livingston islands and are represented by the Pencil Beach Member (Low Island) and the Anchorage and Miers Bluff formations (Livingston Island) (Fig. 14). These rocks are dominated by mudstones and fine-grained sandstones interbedded with minor ash layers (e.g. Smellie, 1979; Hathway and Lomas, 1998; Bastias et al., 2019), and with fossil assemblages implying a deep-marine setting (Thomson, 1982; Kiessling et al., 1999). While the specific depositional environment is not clear, a turbiditic setting has been proposed (e.g. Hervé et al., 2006). Provenance studies have indicated derivation from a Permian magmatic belt, similar to the Triassic turbidite deposits of the Trinity Peninsula Group on the Antarctic Peninsula (Castillo et al., 2016). While these rocks crop out in the southern and central South Shetland Islands (Fig. 14), they may form part of the basement present throughout the archipelago (e.g. Haase et al., 2012) and this basement. Therefore, its presence suggests that probably most of the South Shetland Islands were under a marine sedimentation environment during the Jurassic.

5.2. Subaerial volcanism and sedimentation: ~140 – 35 Ma

The marine Jurassic succesions of Low and Livingston islands are overlain by volcanic and volcaniclastic rocks of continental affinity with abundant palaeobotanical fauna (e.g. Bastias et al., 2019). These two units are separated by an erosional hiatus and are the evidence of a clear change in depositional environment in the South Shetland archipelago. The earliest volcanism is dated at ~140-137 Ma (Low Island; $^{40}Ar/^{39}Ar$ in whole-rock; Bastias et al., 2019) and is accompanied by ~137 Ma acidic plutonism (Low and Livingston islands; Hervé et al., 2006; Bastias et al., 2019). Furthermore, ~140 Ma volcanism on the South Shetland archipelago links in with the resumption of arc magmatism on the Antarctic Peninsula, which was continuously active throughout the Cretaceous and most of the Cenozoic (e.g. Leat et al., 1995; Riley et al., 2018). Bastias et al. (in review) based on thermochronological studies suggested that the Antarctic Peninsula experienced rock uplift and erosion during the Early Cretaceous due to the westward migration of South America induced by the opening of the South Atlantic (e.g. Mpodozis and Ramos, 1989). This is in agreement with similar exhumational and erosional processes occurring further north in the Andes during the Early Cretaceous in Patagonia (Gianni et al., 2018, 2020), Colombian and Ecuadorian (Sarmiento and Rangel, 2004; Martin-Gombajov and Winkler, 2008; Villagomez and Spikings, 2013; Spikings et al., 2015). Therefore, the first deposits of subaerial volcanism in the South Shetland archipelago may be the record of the early development of the so-called ‘Andean cycle’ in the Chilean-Argentinian Andes, which is responsible for most of the onset of the development of the current Andean range.

The Cretaceous period on Livingston Island is marked by shallow water sedimentation, volcanism, slope-apron sandstones and conglomerates of the Byers Group (Hathway et al., 1999), and on Low

Island it is dominated by volcanic and volcaniclastic rocks (Smellie, 1979; Bastias et al., 2019). Cretaceous volcanic and volcaniclastic rocks are also present on Snow, Greenwich, Robert and King George islands (Fig. 14). However, it is noteworthy that several of these successions have not been dated by modern geochronological techniques. Volcanism, minor plutonism and associated deposition of volcaniclastic rocks continued throughout the Palaeocene and Eocene, which is represented by exposures on Livingston and King George islands (Fig. 14; e.g. Smellie et al., 1984; Zheng et al., 1991). While the presence of early Cenozoic deposits has not been reported on Low, Snow, Greenwich, Livingston and Robert islands, they may be buried beneath the icecap cover.The period between ~140 – 35 Ma is key for the evolution of life on Antarctica and particularly on the Antarctic Peninsula. The mountainous landscape allowed the development of a complex and diverse palaeobotanical fauna that is still providing new paleontological discoveries (e.g. Birkenmajer, 1980; Cao, 1994; Poole et al., 2001; Hunt and Poole, 2003; Warny et al., 2015; Trevisan et al., 2022).

5.3. Glacial and interglacial volcanism and sedimentation: ~35 Ma – recent

The sudden and widespread Cenozoic glaciation of Antarctica is one of the most fundamental reorganisations of global climate known in the Phanerozoic geological record (e.g. Kennett, 1977; Zachos et al., 2001; De Conto & Pollard, 2003). While there is reasonable consensus on the timing of this rapid and widespread glaciation of Antarctica, there is ongoing debate regarding the causative mechanism responsible for such a change. The two main hypotheses are (i) the Cenozoic opening of sea gateways between Antarctica-Patagonia and Antarctica-Australia, which correspond to the Drake Passage (Barker & Burrell, 1977) and the Tasmanian Passage (Exon et al., 2000), which reduced southward heat transport, cooling the Southern Ocean and the land masses around it; and (ii) a global decline in atmospheric CO_2 (e.g. De Conto & Pollard, 2003). Regardless, the rocks that document the onset of this glacial event are remarkably exposed in northeast King George Island (Fig. 13; e.g. Warny et al., 2015); specifically in Polonez Cove (Fig. 13; e.g. Birkenmajer, 1980; Warny et al., 2019) and on the Melville Peninsula (Fig. 13, e.g. Bikenmajer, 2001; Warny et al., 2015). These rocks are mostly comprised of two main sequences, (i) Eocene to Oligocene lava flows and pyroclastics, marine tillites, associated glaciogenic sediments and sandstones and (ii) Eocene to Miocene volcano-sedimentary sequences. Research is being currently conducted by different research groupings to further understanding of the timing and duration of these glacial events.

5.4. Smith Island: age and tectonic implications

The mid ocean ridge basalts of Smith Island that were metamorphosed under blueschist/greenschist-facies conditions rocks yield an age of ~58–47 Ma for metamorphism ($^{40}Ar/^{39}Ar$ in white mica; Grunow et al., 1992). While these rocks have been interpreted as part of a subduction complex composed of ocean floor material mixed with arc-derived sediment (Grunow et al., 1992; Truow et al., 1998a), its tectonic implication and protholith age is not clear. Furthermore, Bastias et al. (in review), pointed out that Smith Island resides in the same structural position to the west of the Mesozoic intrusions, as elsewhere in the Andes (northern Andes; e.g. Spikings et al., 2015). Therefore, it is feasible that these high pressure-low temperature rocks were exhumed during the Early Cretaceous in the west margin of the Antarctic Peninsula by compression associated with the westward migration of South America induced by the opening of the South Atlantic (e.g. Mpodozis and Ramos, 1989), although more work is required (and is ongoing) to support this hypothesis.

5.5. Migration of volcanism

A broad geographical trend of decreasing ages has been proposed for the volcanism from southwest to northeast (Pankhurst and Smellie, 1983). The cause of the age migration is unknown, while a potential explanation has been linked to the rotation of the Antarctic plate (Birkenmajer, 1994), this has been challenged with the presence of Cenozoic dykes in Livingston Island (Willan and Kelley, 1999). Bastias et al. (2019) suggested that the migration of the volcanism may have occurred from the Late Cretaceous, based on the presence of Early Cretaceous volcanism in Low and Livingston islands. Alternatively, Haase et al. (2012) suggested that the volcanism proceeded in major steps, rather than a progressive north-eastern migration. Nevertheless, the presence of Eocene magmatism in Livingston Island, located in the centre of the archipelago (Willan and Kelley, 1999), challenge any model of progressive north-easterly migration. The Figure 14 depicts this problem, where it is possible to observe a younging age trend on the volcanism towards the islands located in the north, with the exception of a dyke unit in Livingston Island of early Cenozoic age.

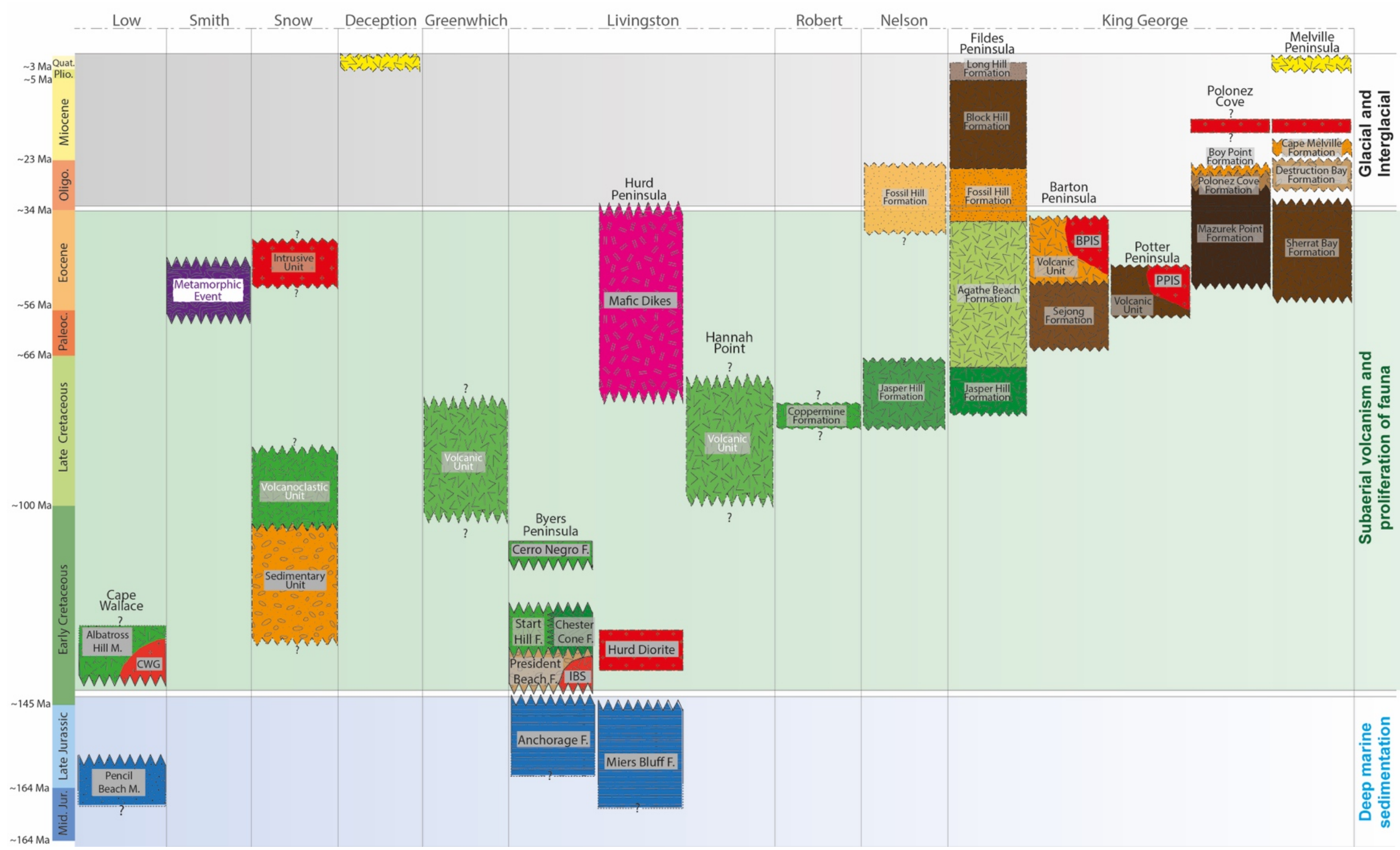

Figure 14. Major lithostratigraphic units of the South Shetland Islands. Three main tectonostratigraphic episodes are recognised: (i) deep marine sedimentation from ~164 to 139 Ma, (ii) subaerial volcanism and proliferation of plant fauna from ~139 to 34 Ma and (iii) glacial and interglacial sedimentation and volcanism from ~34 Ma.

6. Conclusions

The South Shetland Islands host a Mesozoic to Cenozoic sedimentary, volcanic and volcaniclastic succession that record three main tectonostratigraphic successions:

1. Deep marine sedimentation from the Middle to Late Jurassic (~164 to 140 Ma), present in the centre and south of the archipelago.
2. Subaerial volcanism and deposition with proliferation of plant fauna from the Early Cretaceous to the Eocene/Oligocene boundary (~140 to 35 Ma), which is present throughout the archipelago.
3. Glacial and interglacial deposits from the Oligocene until modern time (from ~35 Ma), exclusively present on King George Island.

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