Cretaceous stratigraphy of Antarctica and its global significance

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Abstract: The Cretaceous Period is particularly well represented by a thick sequence of clastic sedimentary rocks exposed in the Antarctic Peninsula region of western Antarctica. This was an active margin throughout the Late Mesozoic, and in total some 7 km+ of Cretaceous sedimentary rocks accumulated in a series of fore-, intra- and back-arc basins. The Fossil Bluff Group of eastern Alexander Island can be traced from the Jurassic–Cretaceous boundary into the Upper Albian and represents a broad-scale shallowing-upward sequence from deep marine to a prominent Upper Albian fluvial interval in which high-density forests developed at a palaeolatitude of 75° S. The Cretaceous sequence exposed in the James Ross Island group continues right through the Upper Cretaceous to the Cretaceous–Paleogene boundary. The Campanian–Maastrichtian succession in particular is over 2 km in total thickness and richly fossiliferous. The improved Cretaceous stratigraphy of Antarctica is an invaluable terrestrial record of climatic change at a high palaeolatitude. This includes a gradual increase in temperature to the Cretaceous Thermal Maximum, and then a decline to the Cretaceous–Paleogene boundary. There may be no simple link between these palaeotemperature changes and Cretaceous patterns of biotic radiation and extinction.

The Cretaceous Period has always figured strongly in the geological investigation of West Antarctica. The Antarctic Peninsula represents the southernmost continuation of the Andean magmatic arc, and within this region a series of fore-, intra- and back-arc basins have yielded an extensive sedimentary record that covers almost the entire period (Storey and Garrett 1985; Hathway 2000; Bastias *et al.* 2020; see also below). Cretaceous fossils from the Antarctic Peninsula were in fact some of the very first palaeontological specimens to be collected from the entire continent and triggered a period of intensive exploration at the beginning of the twentieth century (Zinsmeister 1988). The fact that some of these fossils were of plant material generated considerable speculation that Antarctica may not always have been an ice-covered continent (Nordenskjöld and Andersson 1905).

Otto Nordenskjöld, the leader of the Swedish South Polar Expedition (1901–03), made the first detailed investigations of the geology of Seymour Island within the James Ross Island (JRI) group (Fig. 1) and in so doing recorded numerous ammonites that he thought 'belong to the middle or upper cretaceous system' (Nordenskjöld and Andersson 1905, p. 250). The strata containing them were quite distinct from the overlying 'Tertiary beds' exposed in a more central position on the island. In making this distinction, Nordenskjöld was in effect, for the very first time, delimiting the upper limit of the Cretaceous Period in Antarctica and thus the general position of the Cretaceous–Paleogene (K–Pg) boundary. The Late Cretaceous age of Nordenskjöld's ammonites was confirmed using comparisons with both southern India and Madagascan faunas, although there was a general consensus that the age was that of the Campanian Stage rather than the terminal Maastrichtian (Weller 1903; Nordenskjöld and Andersson 1905; Kilian and Reboul 1909; Spath 1953). Even as late as 1966, the entire Cretaceous succession within the JRI group was regarded as essentially Campanian in age (Howarth 1966).

Nevertheless, building on extensive fieldwork in the 1958–59 field season, J.S. Bibby (Falkland Islands Dependencies Survey/British Antarctic Survey) was able to establish a general 10° E/SE dip across the whole JRI group and, even accounting for the presence of one or more large, shallow folds, estimated a total thickness in excess of 5 km for the Cretaceous succession (Bibby 1966). It was highly unlikely that such a thickness could be attributed to the Campanian Stage alone, and subsequent research over the last 50+ years has established both that this is, in fact, the true thickness and that it spans much of the Early–Late Cretaceous Period (to be discussed in further detail below).

On the Pacific, or forearc, side of the Antarctic Peninsula, early exploration in the 1930s and 1940s established the presence of a similar huge
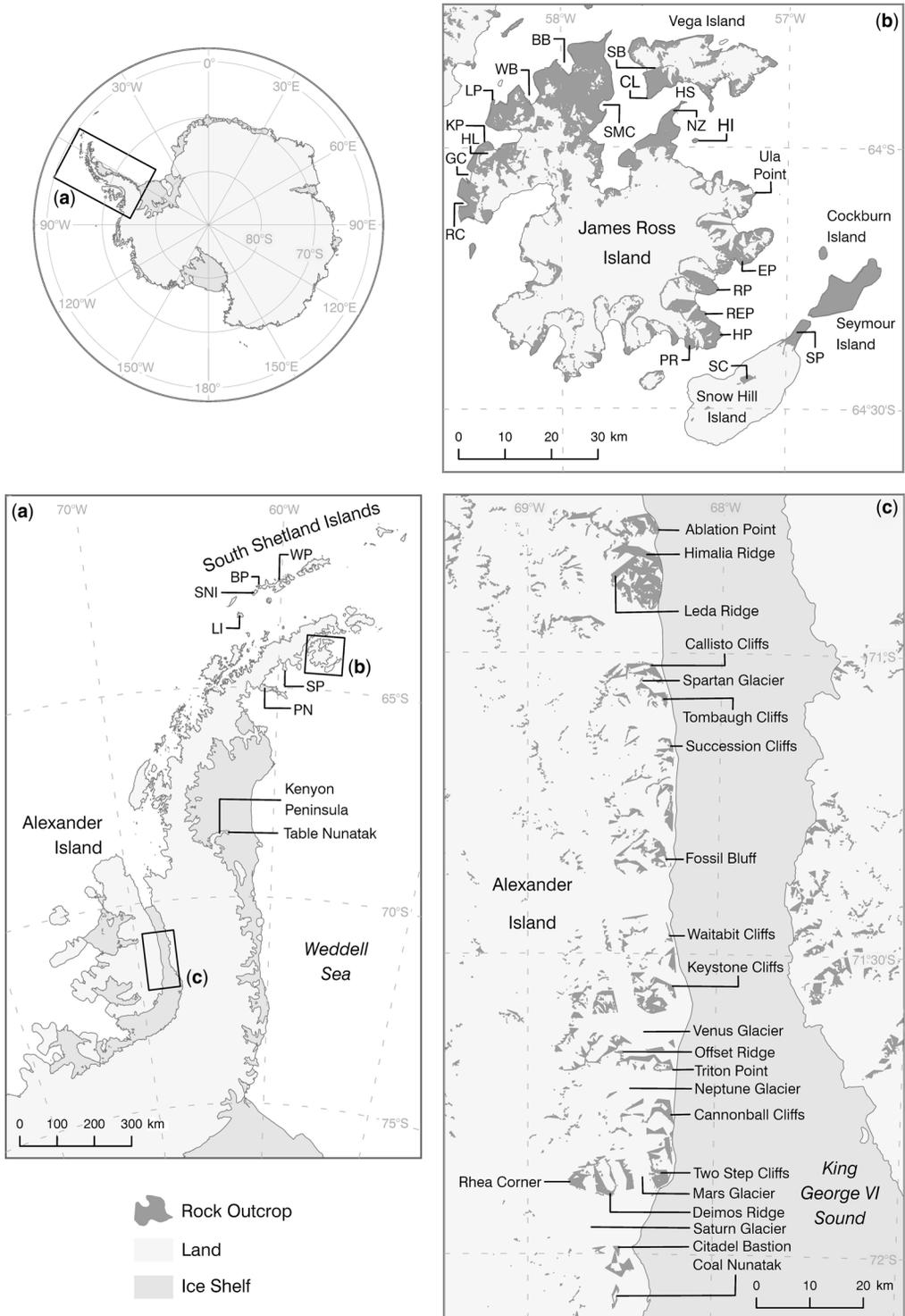
Fig. 1. Locality map for the Antarctic Peninsula region. (**a**) Antarctic Peninsula showing the location of the James Ross Island group (**b**) and eastern Alexander Island (**c**). BP, Byers Peninsula, Livingston Island, South Shetland

thickness of clastic and volcaniclastic sedimentary rocks along the eastern margin of Alexander Island (Fig. 1). Initial fossil collections of ammonites, belemnites and bivalves indicated a Late Jurassic age for the more northerly exposures and Early Cretaceous for those to the south (Cox 1953; Spath 1953; Howarth 1958). Thus, the Alexander Island stratigraphic sequence contains the Jurassic–Cretaceous (J–K) boundary, and a number of tie points with the lower JRI sedimentary succession (Crame 1985; Crame and Howlett 1988; Bastias et al. 2020). Taken together, these two sequences represent an almost complete record through the Cretaceous and as such offer a unique opportunity to make high- to low-latitude palaeoenvironmental and palaeobiological comparisons for the period. It is the intention of this study to review our current knowledge of Antarctic Cretaceous stratigraphy and show how it can be used to promote studies of Cretaceous climatic and biotic evolution in the high latitudes.

The forearc basin of Alexander Island

The Fossil Bluff Group (FBG) of eastern Alexander Island comprises a remarkably complete record of the evolution of a major forearc basin across the J–K boundary (Butterworth et al. 1988; Moncrieff and Kelly 1993; Nichols and Cantrill 2002). It is exposed in a zone c. 200 km long by 25–30 km wide that is bounded to the east by King George VI Sound and to the west by a faulted, unconformable contact with the Le May Group accretionary complex (Doubleday et al. 1993) (Fig. 1). It is characterized by a wide range of clastic and volcaniclastic lithologies that range in age from Late Jurassic (Kimmeridgian–Tithonian) at Ablation Point to Late Albian in the vicinity of Coal Nunatak (Fig. 1). The Lower Cretaceous succession reaches a total thickness in excess of 5 km, and there are no obvious major stratigraphic breaks; as such, it serves as an important Lower Cretaceous reference section for regional correlations.

The basal Ablation Point Formation is characterized by a huge synsedimentary mélange that is interpreted as a massive slope-collapse slump unit (Butterworth and Macdonald 1991). Siltstone–sandstone lithologies predominate in the rotated blocks, and there is marked lateral variation throughout the unit (Butterworth et al. 1988). The Ablation Point Formation has yielded a distinctive perisphinctid ammonite–belemnopseid belemnite–buchiid bivalve fauna with Kimmeridgian age affinities (Crame and Howlett 1988; Howlett 1989).

The Kimmeridgian–Tithonian boundary is placed at the base of the Virgatosphinctes and Hibolithes belligerundi biozones and junction with the overlying Himalia Ridge Formation (Fig. 2) (Howlett 1989). The latter is 2200 m thick and comprises a range of lithologies: coarse-grained conglomerates with subordinate pale sandstones, channelled conglomerate complexes and interbedded sandstones and siltstones; slump zones and synsedimentary faults are common (Butterworth et al. 1988). A rich Berriasian marine fauna in the upper levels of the formation includes the ammonites Raimondiceras, Spiticeras, Bochianites and Haplophylloceras, together with a variety of belemnites and bivalves (Crame and Howlett 1988); on the basis of these occurrences, the J–K boundary is placed at the 2255–2384 m level in the Himalia Ridge section, Ablation Point region and 503 m level in the Spartan Glacier section (Fig. 2) within both the Haplophylloceras and Hibolithes belligerundi biozones (Crame and Howlett 1988; Howlett 1989). The Himalia Ridge Formation is interpreted as a predominantly submarine fan channel complex that includes some shallower intervals at mid- to outer shelf depth (Nichols and Cantrill 2002).

The Berriasian–Valanginian boundary is placed within the topmost levels of the Himalia Ridge section and mid-levels of the Spartan Glacier section at the base of the Belemnopsis lanceloti Biozone and Belemnopsis alexandri Sub-biozone (Fig. 2) (Howlett 1989). This in turn is very close to the transition between the Himalia Ridge Formation and overlying Spartan Glacier Formation (Fig. 2), with the latter being represented by a thick, monotonous sequence of dark mudstones and siltstones and interbedded thin, fine-grained sandstones (Butterworth et al. 1988). Bioturbation is common throughout

Fig. 1. *Continued.* Islands; LI, Low Island, South Shetland Islands; PN, Pedersen Nunatak; SNI, Snow Island, South Shetland Islands; SP, Sobral Peninsula; TN, Table Nunatak; WP, Williams Point, Livingston Island, South Shetland Islands. (b) James Ross Island group. BB, Brandy Bay; CL, Cape Lamb, Vega Island; EP, Ekelöf Point; GC, Gin Cove; HI, Humps Island; HL, Hidden Lake; HP, Hamilton Point; HS, Herbert Sound; KP, Kotick Point; LP, Lagrelius Point; NZ, The Naze; PR, Punta Redonda; RC, Rum Cove; REP, Redshaw Point; RP, Rabot Point; SB, Sandwich Bluff, Vega Island; SC, Sanctuary Cliffs, Snow Hill Island; SMC, Santa Marta Cove; SP, Spath Peninsula, Snow Hill Island; WB, Whisky Bay. Tumbledown Cliffs are located between Gin Cove (GC) and Rum Cove (RC). (c) Eastern Alexander Island. Phobos Ridge forms the western boundary of Mars Glacier. Source: map produced by the Mapping and Geospatial Information Centre (MAGIC), © British Antarctic Survey, UK Research and Innovation, 2023. Antarctic coastline and polygon taken from the SCAR Antarctic Digital Database, accessed 2023. Rock outcrops taken from Gerrish (2020).
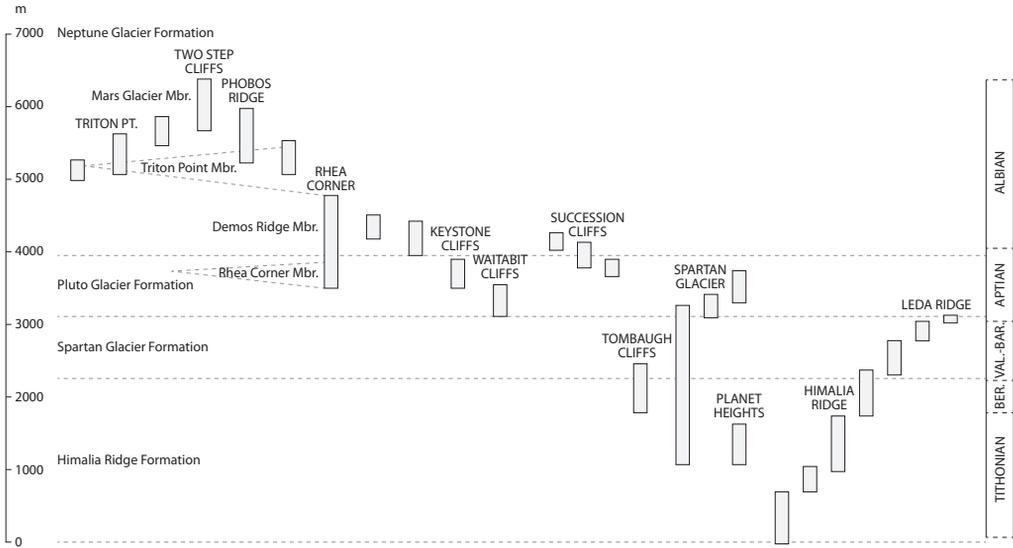
Fig. 2. Simplified stratigraphic correlations for the principal Cretaceous exposures of the Fossil Bluff Group, Alexander Island. Based on correlations presented in Butterworth *et al.* (1988), Crame and Howlett (1988) and Moncrieff and Kelly (1993).

the formation, with *Chondrites*, *Planolites* and *Zoophycos* all prominent, and ammonites, belemnites and bivalves being locally abundant. In the upper Spartan Glacier section, a 22 m thick siltstone sequence has yielded a prolific bivalve assemblage, which includes *Retroceramus everesti*, *Grammatodon*, *Pinna*, *Entolium*, *Myophorella*, astartids and anomalodesmatans (Crame and Howlett 1988). The Spartan Glacier Formation represents a series of outer shelf to slope palaeoenvironments (Nichols and Cantrill 2002).

Although there are no obvious stratigraphic discontinuities within the Spartan Glacier Formation, definition of the top of the Valanginian Stage by macrofossils remains elusive, as does any formal recognition of the succeeding Hauterivian and Barremian stages. A distinctive, recessive-weathering dark mudstone unit that can be traced from the upper Tombaugh Cliffs northwards around Spartan Glacier to the base of Leda Ridge (Fig. 2) has yielded specimens of the inoceramid bivalve *Anopaea trapezoidalis* with Hauterivian–Barremian affinities (Crame and Kelly 1995). This occurrence passes up in turn to a more prolific fauna that includes lytoceratid ammonites, the first representatives of an *Aucellina* bivalve referred to an *Aucellina caucasica–Aucellina aptiensis* group, and the first occurrences of members of the *Inoceramus neocomiensis* group (Crame and Howlett 1988). Even though there are no completely diagnostic taxa in this fauna, it may well prove to be essentially

Barremian in age. It is succeeded in the topmost levels of the Spartan Glacier Formation by a basin-wide influx of aconeceratid ammonites together with the ammonite *Eulytoceras*, belemnite *Tetrabelus willeyi* and bivalves *Inoceramus deltoides* and *Aucellina andina–radiatostriata* group. This distinctive fauna has been taken as a marker for the base of the Aptian Stage within the FBG (Butterworth *et al.* 1988; Crame and Howlett 1988) (Fig. 2).

The succeeding Pluto Glacier Formation is *c.* 800 m thick and can be traced southwards from Spartan Glacier through Succession Cliffs and Fossil Bluff to the region of Keystone Cliffs and Offset Ridge (Figs 1 & 2). It is characterized by a higher proportion of fine- to medium-grained sandstones that are locally cross-bedded and intensely bioturbated (Butterworth *et al.* 1988; Moncrieff and Kelly 1993). In its higher levels there are extensive sequences of thickly bedded sandstones that have weathered into a series of distinctive pyramidal peaks (Butterworth *et al.* 1988; Moncrieff and Kelly 1993) (Fig. 3). The Pluto Glacier Formation is generally interpreted to represent a significant shallowing to tidal shelf palaeoenvironments (Nichols and Cantrill 2002).

The lower Pluto Glacier Formation has yielded a fauna that includes aconeceratid ammonites such as *Sanmartinoceras*, large heteromorphs assigned to *Australiceras* and *Tropaeum*, the belemnite *Tetrabelus willeyi* and the bivalves *Inoceramus deltoides* and *Aucellina andina–radiatostriata* group, all of

Fig. 3. Upper Pluto Glacier Formation, Spartan Glacier, Alexander Island. There is a steep regional dip SW, and *c.* 500 m of Aptian mudstones and interbedded sandstones are exposed immediately beneath a localized synsedimentary thrust zone. Source: photo from BAS Photographic Collections.

which indicate an Aptian age (Butterworth *et al.* 1988; Crame and Howlett 1988; Moncrieff and Kelly 1993) (Fig. 2). However, in the upper levels there is a suite of ammonites and bivalves with Albian affinities, and this is backed up by palynological data (Dettmann and Thomson 1987). A precise level for the Aptian–Albian boundary has yet to be established.

Further extensive fieldwork in the late 1980s–early 1990s at the southern end of the area of outcrop in the Venus Glacier–Neptune Glacier–Saturn Glacier region (Figs 1 & 2) demonstrated that the FBG is in fact considerably more extensive than originally thought. An additional thickness of *c.* 2.8 km of strata in this region was added on top of the *c.* 4 km already established to the north, and this was assigned to both the Pluto Glacier Formation and a newly defined Neptune Glacier Formation, the uppermost formation of the FBG (Moncrieff and Kelly 1993). The latter unit, which is *c.* 2.2 km thick, is readily distinguished by a suite of coarser-grained lithologies that represent a further marked decrease in water depth: a nearshore/paralic facies, followed by an emergent, fluvial interval and finally a return to nearshore/paralic conditions (Nichols and Cantrill 2002). These three contrasting facies were in turn used as the basis for establishing three component members: Deimos Ridge, Triton Point and

Mars Glacier (Moncrieff and Kelly 1993) (Fig. 2). The base of the *c.* 700 m thick Deimos Ridge Member is defined by a massive sandstone bed and its top by the first *in situ* terrestrial vegetation in the Cretaceous succession. Further massive sandstones with faint internal lamination and normal grading are interpreted as high-density turbidites, but at higher levels in the unit, cross-bedding represents extremely shallow-water, probably tidal, conditions (Moncrieff and Kelly 1993). Macrofossils are uncommon, but the ammonite *Lechites*, together with the belemnite *Dimitobelus*, is indicative of a Late Albian age (Moncrieff and Kelly 1993).

Although the succeeding Triton Point Member is only 200 m thick at the type locality, it thickens rapidly southwards to perhaps as much as 1200 m in the Citadel Bastion–Coal Nunatak region and exhibits a much greater range of sedimentary facies. For this reason, Nichols and Cantrill (2002) elevated it to formation status and demonstrated that the component sandstone, mudstone and occasional conglomerate lithologies represented a series of fluvial environments. They subdivided the new formation into two main categories: the 'Citadel Bastion Member', comprising coarse-grained overbank facies and channel fill sandstones deposited in a braided river system; and the stratigraphically higher 'Coal Nunatak Member', demonstrating a higher proportion of

mudrocks related to a fine overbank facies (Nichols and Cantrill 2002). As the precise stratigraphic relationships of these two new members are still unclear, we have retained the use of Triton Point Member in this study pending further investigations.

Within the lower 'Citadel Bastion Member', overbank areas were sufficiently stable to form soils, and floodplains were dominated by diverse forest communities comprising conifers (mainly podocarps) with Bennettitales, pteridophytes, liverworts and some angiosperms (Falcon-Lang *et al.* 2001; Nichols and Cantrill 2002; Cantrill and Poole 2012). The fluvial palaeoenvironment matured from a braided river system with frequent floods and unstable channel banks to one of a meandering river system with more stable floodplains as represented by the 'Coal Nunatak Member' (Cantrill and Poole 2012). This unit records the spectacular development of high-density climax forests of podocarpaceous conifers with an understory of ginkgos, cycadophytes, pteridophytes and liverworts (Nichols and Cantrill 2002). Araucarian conifers become increasingly important in this member where they formed open stands with pteridophytes and herbaceous angiosperms (Nichols and Cantrill 2002) (Figs 4 & 5).

It is likely that the Triton Point Member at least partly underlies the Mars Glacier Member, with the latter comprising in the region of 1000 m of normally graded massive sandstones interbedded with thinner mudstones (Moncrieff and Kelly 1993) that have yielded an abundant marine fauna. The common occurrence of thick-shelled bivalves such as mytilids, cucullaeids, trigoniids and various heterodonts attests to a shallow, high-energy environment, and the belemnite *Dimitobelus diptychus*, ammonite *Puzosia* and members of both the *Inoceramus carsoni* and *Inoceramus sutherlandi* groups indicate a Late Albian age (Moncrieff and Kelly 1993). The youngest marine fossil assemblage in the FBG occurs below the summit of Two Steps Cliff and includes the Late Albian ammonite *Hamites virgulatus* (Moncrieff and Kelly 1993) (Figs 2 & 5).

The importance of the Fossil Bluff Group sedimentary record

The FBG represents a major shallowing-upward sequence of Kimmeridgian–Albian age in the region of 7 km in thickness (Fig. 2); as such, it comprises a remarkably complete record of the evolution of a forearc basin (Butterworth and Macdonald 1991; Kelly and Moncrieff 1992). The component sediments are very largely derived from the volcanic arc to the east and show a compositional trend reflecting the progressive un-roofing of the arc (Butterworth *et al.* 1988; Hathway 2000). Major slump
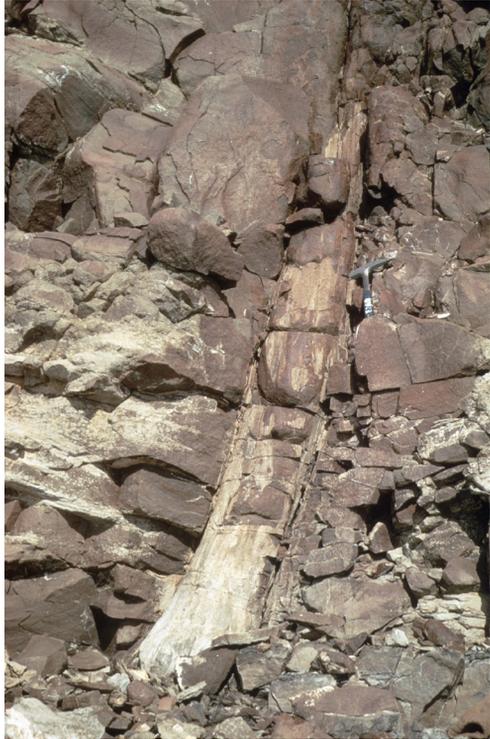

Fig. 4. Isolated standing fossil conifer tree exposed in massive fluvial sandstones of the Triton Point Member, Neptune Glacier Formation, Coal Nunatak, Alexander Island. Hammer on the upper right of the trunk for scale. Source: photo from BAS Photographic Collections.

zones throughout the sequence suggest active tectonic control on the fill of the basin (Butterworth *et al.* 1988), but despite this all the stratigraphic breaks observed are very probably of only local significance (Crame and Howlett 1988).

The Lower Cretaceous component of the FBG comprises a total thickness of *c.* 5.3 km and can be traced from the J–K boundary to the Upper Albian, a stratigraphic record of *c.* 45 Myr. Although the two basal Cretaceous stages, Berriasian and Valanginian, are clearly defined by macrofossils, the two succeeding ones, Hauterivian and Barremian, are not. In all probability they are represented by a 700 m thickness of predominantly homogeneous siltstone lithologies within the mid- to upper levels of the Spartan Glacier Formation. It is hoped that future macro- and micro-palaeontological studies may provide a clearer definition of both the Hauterivian and Barremian boundaries. The base of the Aptian Stage is defined by the basin-wide influx of a fauna that includes aconeceratid ammonites such as *Sanmartinoceras*, large heteromorph ammonites

Fig. 5. Normally graded massive sandstones interbedded with thinner mudstones of the Mars Glacier Member, Neptune Glacier Formation, Two Step Cliffs, Alexander Island. Approximately 350 m thick section exposed. Source: photo from BAS Photographic Collections.

assigned to *Australiceras* and *Tropaeum*, the belemnite *Tetrabelus willeyi* and bivalves *Inoceramus deltoides* and *Aucellina andina–radiatostriata* group (Crame and Howlett 1988). The base of the Albian probably occurs in the uppermost levels of the Pluto Glacier Formation and is tentatively defined by both macrofossils and palynology (Dettmann and Thomson 1987; Crame and Howlett 1988; Moncrieff and Kelly 1993).

As far as can be determined, the stratigraphically highest levels of the FBG, that is, within the uppermost Neptune Glacier Formation, are no younger than Late Albian in age (Fig. 2); no trace of Late Cretaceous fossils has been found in the southernmost areas of outcrop (Kelly and Moncrieff 1992; Nichols and Cantrill 2002). The age of these highest FBG beds is very close to that of a peninsula-wide mid-Cretaceous tectonic event that has been inferred to be the product of a superplume, and the most likely cause of abrupt shallowing within the highest stratigraphic levels of the FBG (Vaughan 1995; Vaughan and Storey 2000). The plume may have effectively curtailed any further marine sedimentation on the forearc side of the Antarctic Peninsula.

Despite its remote location, the FBG of eastern Alexander Island has huge potential to serve as a major reference section for Early Cretaceous regional stratigraphic correlations. Preliminary links with Patagonia and New Zealand in particular (e.g. Riccardi 1988; Cooper 2005) can be refined to assist with more detailed correlation to the standard Northern Hemisphere reference sections.

The fossil forests of southern Alexander Island

The Cretaceous fossil forests of southern Alexander Island represent one of the most spectacular components of the entire Antarctic fossil record (Figs 4 & 6). These high-density climax forests, comprising araucarian and podocarp conifers, ginkgos, cycadophytes and ferns, are testament to a relatively humid climate punctuated by high mean annual rainfall at an estimated palaeolatitude of 75° S, and thus have implications for global palaeoclimate studies (Jefferson 1982; Falcon-Lang *et al.* 2001; Cantrill and Poole 2012). They represent a key component for defining a greenhouse to icehouse transition which represents one of the most striking features of recent Earth history (Francis *et al.* 2008).

When both wood and foliage is considered, the Late Albian flora of Alexander Island is found to be surprisingly diverse. Approximately 42 genera

Fig. 6. Reconstruction of the high-density climax forests of Alexander Island, comprising araucarian and podocarp conifers, ginkgos, cycadophytes, ferns and other taxa. Based on the work of J. Howe and others (further details on the palaeocommunity analysis are contained in Francis *et al.* 2008). Source: © Bob Nicholls Art.

and 70 fossil species have been recorded so far, and these figures are high in comparison with many mid-latitude Early Cretaceous floras (Cantrill 1997; Falcon-Lang *et al.* 2001). Conifers constitute the most abundant taxa, especially large trees of fossil types similar to modern Araucariacea, Podocarpacea and Taxodiaceae (Figs 4 & 6), but it is interesting to note that they constitute only 25% of the total floral diversity (Cantrill and Poole 2012). Much of the latter is in fact made up of an unusually high component of liverworts and ferns, with the latter representing in excess of 40% of the total diversity (Cantrill and Poole 2012). Liverworts amount to *c.* 15%, and a further important component is the presence of both large arborescent and small herbaceous angiosperms (12% of the total floral diversity) (Cantrill and Nichols 1996). However, as with all fossil plant assemblages, the sedimentary setting may have influenced the preservation (or not) of some components of the flora.

These Late Albian fossil forests of Alexander Island (Fig. 6) are similar to other temperate, broad-leafed, araucarian–podocarp rainforests that are thought to have covered much of Gondwana above 60° S at this time (Spicer and Chapman 1990; Cantrill and Poole 2012). They bear a strong resemblance to extant rainforests in New Zealand that are currently growing under warm temperate, humid climates (Falcon-Lang *et al.* 2001; Cantrill and Poole 2012).

The intra-arc basin of the South Shetland Islands

Upper Jurassic–Lower Cretaceous sedimentary sequences also occur at a series of more isolated localities in the South Shetland Islands: Byers Peninsula, Livingston Island; President Head, Snow Island; and Cape Wallace, Low Island (Fig. 1). Collectively, they also record a regressive megasequence, in this particular instance from deep-marine to continental intra-arc facies (Hathway 1997; Hathway and Lomas 1998; Bastias *et al.* 2020).

The most-extensive and best-exposed succession is that of the Byers Group, Byers Peninsula, which has an estimated total thickness of *c.* 2.7 km (Hathway and Lomas 1998; Bastias *et al.* 2020). The lowermost stratigraphic component, the 120 m thick Anchorage Formation, comprises bioturbated, hemipelagic mudstones with interbedded pale-weathering tuffs. It is divided into a lower New Plymouth Member and an upper Ocoa Point Member (Hathway and Lomas 1998) and has yielded a mixed ammonite–belemnite–bivalve–radiolarian

assemblage with Tithonian–Berriasian age affinities (Crame *et al.* 1993). However, an essentially Late Jurassic age for the Anchorage Formation (Pirrie and Crame 1995) has been reinforced by a new U–Pb zircon age of 153.1 ± 1.9 Ma (i.e. Kimmeridgian) from an ash bed in the Punta Ocoa Member (and thus contemporaneous Zr, not detrital grains; Bastias *et al.* 2020).

The succeeding President Beaches Formation (PBF) overlies the Anchorage Formation with a slight angular unconformity, so the exact position of the J–K boundary cannot be determined (Hathway and Lomas 1998; Bastias *et al.* 2020). The PBF has a maximum thickness of *c.* 600 m, but exposures are largely restricted to foreshore platform localities and a series of east–west-trending gullies. Basal dark mudstones interbedded with fine- to medium-grained sandstones pass up into olive-grey-weathering mudstones with fine planar laminations; these in turn pass up into coarser-grained sandstones and matrix-supported conglomerates. This formation is characterized by an ammonite assemblage that includes '*Spiticeras*', *Blanfordiceras* and *Himalayites*, together with a range of pteriomorph bivalves (Smellie *et al.* 1980; Crame *et al.* 1993). The mid to Late Berriasian age affinities of this macrofossil assemblage has been supported by the accompanying palynomorph assemblage (Duane 1994, 1996).

The PBF is in turn unconformably overlain by the Chester Cone Formation, which comprises a thick basal unit of coarse volcaniclastic sandstones, pebbly sandstones and conglomerates that passes up into a thinner finely laminated mudstone unit with occasional thin sandstones and massive conglomerates (Hathway and Lomas 1998; Bastias *et al.* 2020). It is at least 350 m thick, and a macrofauna from the lower levels includes the belemnites *Belemnopsis* (*Belemnopsis*) *alexandri* and *Belemnopsis* (*Belemnopsis*) *gladiatoris*, and the ammonites *Bochianites*, *Uhligites* and *Neocomites*, which collectively indicate a Valanginian age (Covacevich 1976; Crame *et al.* 1993). This age determination is again supported by palynology (Duane 1994, 1996).

Thereafter the Cretaceous sedimentary record within the South Shetland Islands is incomplete. Continued marine regression is subsequently replaced by subaerial basaltic volcanism on both Livingston Island and Snow Island that is attributed to the Valanginian–Hauterivian interval by radiometric dating (Hathway 2000; Bastias *et al.* 2020). No further marine faunas have been recorded from the South Shetland Islands.

The back-arc Larsen Basin

The extensive Larsen Basin occupies the continental shelf on the eastern flank of the Antarctic Peninsula from almost its northern tip to a latitude of *c.* 69° S. It is best exposed in the northern sector within the JRI group, but small, isolated outcrops can still be traced as far south as the Kenyon Peninsula area (Fig. 1) (Macdonald *et al.* 1988; Hathway 2000). It comprises a minimum total stratigraphic thickness of 5 km, and this figure undoubtedly increases substantially offshore (Macdonald *et al.* 1988).

NE Antarctic Peninsula

The mudstone-dominated Nordenskjöld Formation is exposed intermittently along the northeastern flank of the Antarctic Peninsula, and although neither the base nor the top of this unit is revealed, it is estimated to be in the region of 800 m thick (Farquharson 1983; Macdonald *et al.* 1988). It was initially subdivided into a Kimmeridgian (?)–Tithonian Longing Member and Berriasian Ameghino Member, with the latter possessing thicker and more common tuff beds and bioturbated mudstones (Farquharson 1983; Whitham and Doyle 1989). A third Larsen Member, consisting of unbioturbated mudstones, abundant tuffs and occasional thin sandstones, is of probable Berriasian age, but its relationship to the Ameghino Member is unknown (Whitham and Doyle 1989). The first two of these members record a transition from largely anoxic basin deposition to dysoxic slope environments, but the precise position of the J–K boundary within the Nordenskjöld Formation is uncertain (Doyle and Whitham 1991; Whitham 1993).

The incomplete Late Berriasian–Barremian stratigraphic record noted on both Alexander Island and the South Shetland Islands is very probably also present in the Larsen Basin. More specifically, in the northern region the Aptian Pedersen Formation rests in tectonic contact with the Ameghino Member of the Nordenskjöld Formation on the Sobral Peninsula (Fig. 1) (Hathway 2000). To the south, in the immediate vicinity of Kenyon Peninsula, small, isolated outcrops of intensely bioturbated, cleaved black mudstones do have possible Hauterivian–Barremian age affinities (Thomson 1967). However, much depends on the correct identification of some poorly preserved ammonites and inoceramid bivalves, and tentative links to the Spartan Glacier Formation of Alexander Island (Crame 1985). The Late Berriasian–Barremian is the least well understood interval of stratigraphic evolution within the whole Antarctic Peninsula region, and this could well be because it coincides with a major phase of Palmer Land (i.e. southern Antarctic Peninsula) deformation (Hathway 2000). The unconformity between the Anchorage and President Beaches formations on Byers Peninsula, South

Shetland Islands may represent a time-equivalent surface (Hathway 2000).

The Pedersen Formation is only exposed on southern Sobral Peninsula, with a possible further occurrence on the nearby Pedersen Nunatak (Fig. 1). The base of this unit is not seen, but it comprises 750–1000 m of conglomerate-dominated strata interpreted as a combination of submarine fan deposits and mudstone/sandstone slope apron deposits (Hathway 2000). A combination of $^{40}Ar/^{39}A$ radiometric dates and palynological analyses indicate an Early Aptian age; the presence of a possible Hauterivian ammonite from Pedersen Nunatak needs further confirmation (Thomson and Farquharson 1984; Hathway and Riding 2001).

Northern James Ross Island

Gustav Group

The Pedersen Formation is believed to be laterally equivalent to the Lagrelius Point Formation, the basal unit of the extensive Gustav Group exposed on the NW coast of James Ross Island (Fig. 1) (Hathway and Riding 2001). Difficult to access, the Lagrelius Point Formation comprises a sequence of massive-bedded conglomerates with thin, interbedded sandstones that is at least 500 m thick (Ineson et al. 1986). It has yielded a dinoflagellate cyst assemblage of earliest Aptian age (Riding et al. 1998).

The succeeding Kotick Point Formation is a mudstone-dominated unit with interbedded pale-weathering breccias and conglomerates that has a

maximum thickness of c. 1000 m (Fig. 7). It is also characterized by the presence of very large glide blocks of Nordenskjöld Formation throughout and interpreted as the product of a deep-water submarine fan and slope apron environment adjacent to the faulted western margin of the basin (Ineson 1989; Buatois and Medina 1993).

A characteristic fauna from the mid–upper levels of the Kotick Point Formation includes both ancyloceratid and aconeceratid ammonites, belemnites referable to Dimitobelus (Dimitobelus) stimulus and the bivalves Aucellina sp. and Inoceramus stoneleyi, and has strong Aptian–Albian affinities (Ineson et al. 1986). The uppermost levels of the formation have included the ammonites Silesites aff. antarcticus and 'Pseudothurmannia' cf. mortilleti, both of which are known from the Albian of Alexander Island (Thomson 1984). Associated faunal elements include the belemnite Tetrabelus seclusus, and bivalves Inoceramus cf. sutherlandi, Maccoyella sp. and a second species of Aucellina, all of which have strong Albian age affinities (Ineson et al. 1986). The age range of the Kotick Point Formation is taken as Aptian–Albian (Fig. 7).

The following Whisky Bay Formation is a complex lithostratigraphic unit that shows marked lateral facies variation. It is up to 950 m thick and dominated by coarse-grained lithologies that form pale-weathering scarps and ridges throughout the area of outcrop (Ineson et al. 1986). Three component members are defined in the Brandy Bay–Whisky Bay region, and a further three some 30 km to the SW in the Hidden Lake–Gin Cove–Rum Cove region (Fig. 1) (Ineson et al. 1986; Crame et al.
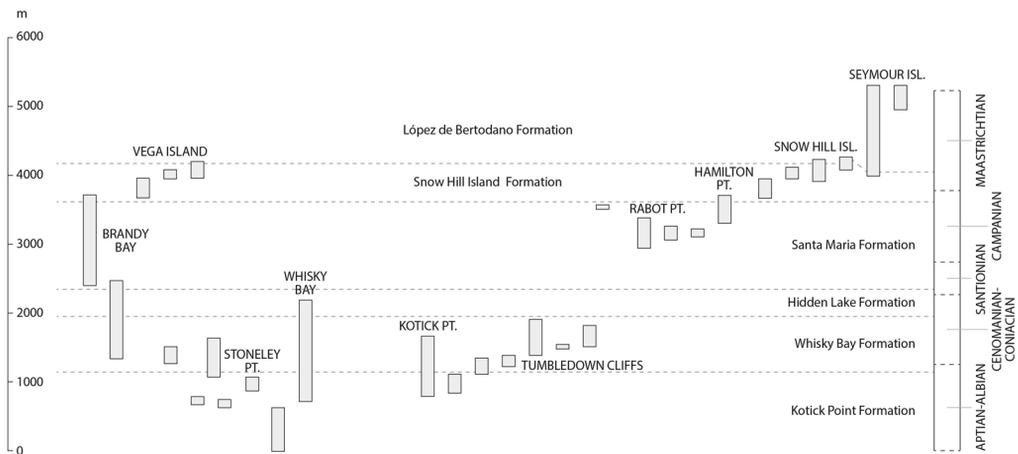

Fig. 7. Simplified stratigraphic correlations for the principal Cretaceous exposures within the James Ross Island group, northeastern Antarctic Peninsula. The lower four formations compose the Gustav Group, and the upper three the Marambio Group. Basal Lagrelius Point Formation not shown. Individual members within the formations are defined in the following references: Ineson et al. (1986); Pirrie et al. (1991); Pirrie et al. (1997); Crame et al. (2006); Crame (2019); Montes et al. (2019).

2006). The basal, *c.* 80 m thick Bibby Point Member in the former of these regions comprises predominantly dark-green-weathering silty sandstones with a fauna of *Silesites*, *Phyllopachyceras*, *Dimitobelus* and *Aucellina* that indicates approximate equivalence with the upper levels of the Kotick Point Formation (Ineson *et al.* 1986). It passes up in turn into the 400–500 m thick Lewis Hill Member, which is characterized by graded conglomerates and pebbly sandstones that contain the distinctive Middle–Upper Albian inoceramid bivalve *Actinoceramus concentricus* (Crampton 1996*a*), together with *Aucellina* sp., gaudryceratid ammonites and a distinctive smooth terebratulid brachiopod (Ineson *et al.* 1986). In addition, the associated palynoflora has unequivocal Late Albian Australasian affinities (Riding and Crame 2002). The three stratigraphically highest samples investigated have yielded the first records of *Ascodinium serratum*, and the incoming of this taxon is a candidate for establishing the Albian–Cenomanian boundary in Antarctica (Crame *et al.* 2006).

The boundary between the Lewis Hill Member and the overlying Brandy Bay Member exhibits a distinct stratigraphic hiatus, with the base of the latter marked by the abrupt incoming of the inoceramid *Tethyoceramus madagascariensis*, which has traditionally been regarded as Late Turonian–Early Coniacian (Ineson *et al.* 1986; Crampton 1996*b*). It comprises 200–400 m of weakly lithified sandstones with intercalated beds of breccia, graded conglomerates and pebbly sandstones (Ineson *et al.* 1986). Ammonites are rare in this unit, but a gaudryceratid was obtained near the middle of the member and a kossmaticeratid, probably *Marshallites*, occurs at the top (Thomson 1984). Dinoflagellate cyst taxa from the Brandy Bay Member strongly suggest an Early Turonian age and a correlation with the *Palaeohystrichophora infusorioides* Interval Zone of Helby *et al.* (1987) (Crame *et al.* 2006). Nevertheless, $^{87}Sr/^{86}Sr$ isotope ages from oyster samples suggest a Late Turonian age, at least for the uppermost levels (McArthur *et al.* 2000). It has been suggested that the Turonian–Coniacian boundary should be placed at the junction between the Brandy Bay Member and the overlying Hidden Lake Formation (Crame *et al.* 2006) (Fig. 7).

An interesting fauna from the mid-levels of the Brandy Bay Member comprises small dimitobelid belemnites, a colonial coral, encrusting bryozoans and bivalves referable to a large, thick-shelled astartid, a bakevelliid and a distinctive flat oyster (Crame *et al.* 2006). This is one of the first obviously very shallow-water assemblages to be encountered within the Gustav Group and coincides with a notable increase in the presence of fossil wood, and a palynological assemblage that indicates the diversification of various plant taxa, including angiosperms (Hayes *et al.* 2006; Poole and Cantrill 2006).

In the region to the SW of Brandy Bay–Whisky Bay, the 315 m thick Gin Cove Member, comprising a range of siltstone and sandstone lithologies with subordinate conglomerates, has yielded an ammonite–inoceramid bivalve assemblage with Late Albian age affinities (Ineson *et al.* 1986). However, there are some indications that the top of this unit may be Cenomanian in age, and it is clear that it may also be laterally equivalent to the base of the succeeding Tumbledown Cliffs Member (Ineson *et al.* 1986). The latter member is slightly thicker and comprises a series of graded sandstones, pebbly sandstones and conglomerates, with occasional slumped intervals and at least two very large slide blocks of Nordenskjöld Formation. Its mid to upper levels have yielded a very distinctive macrofossil assemblage that includes the ammonites *Newboldiceras*, *Sciponoceras* and *Desmoceras*, together with *Inoceramus pictus*. Such a fauna is unequivocally Mid to Late Cenomanian in age (Olivero and Palamarczuk 1987; Kennedy *et al.* 2007). The *c.* 100 m thick Rum Cove Member is mudstone-dominated and has yielded abundant *Tethyoceramus madagascariensis* bivalves (Ineson *et al.* 1986; Crampton 1996*b*).

The Hidden Lake Formation, the topmost lithostratigraphic unit of the Gustav Group (Fig. 7), is lithologically much more homogeneous across the whole of the NW James Ross Island region. With an estimated maximum thickness of 400 m, it has the characteristic appearance of rusty-brown to pale green-brown-weathering sandstones and bioturbated mudstones that represent a partial basin inversion event separating predominantly deeper-water submarine and slope apron environments below from overlying shallower marine shelf facies that characterize the succeeding Marambio Group (Whitham *et al.* 2006). The Hidden Lake Formation has yielded a suite of inoceramid bivalves with strong Coniacian age affinities (Crampton 1996*b*; Crame *et al.* 2006), and this is strongly endorsed by both palynological and strontium isotope studies (McArthur *et al.* 2000; Riding and Crame 2002). A sparse vertebrate assemblage includes reptile bones and shark teeth and vertebrae (Crame *et al.* 2006), and a distinctive bed in the mid-levels of the formation has yielded both charcoalified wood fragments and small angiosperm leaves (Hayes *et al.* 2006). For practical purposes, the Turonian–Coniacian boundary has traditionally been placed at the Whisky Bay Formation–Hidden Lake Formation boundary (Crame *et al.* 2006), but this may now need to be re-evaluated (see below).

Marambio Group

The conformable transition between the Gustav and Marambio groups is best exposed in the hinterland of

the Brandy Bay region. Here, the basal Santa Marta Formation comprises *c.* 1200 m of predominantly shallow shelf clastic/volcaniclastic sediments that are in places intensely fossiliferous (Olivero *et al.* 1986; Pirrie 1991; Scasso *et al.* 1991). It was originally subdivided into three component members, Alpha, Beta and Gamma, that can be traced successively from the Brandy Bay region southeastwards into St Martha Cove (Olivero *et al.* 1986) (Fig. 1). The Alpha Member constitutes 215 m of tuffaceous silty sandstones and mudstones with occasional intraformational conglomerates. Benthos is somewhat restricted in this member, but a distinctive large inoceramid bivalve referred to an informal *Inoceramus expansus* group and the heteromorph ammonite *Baculites* cf. *kirki* have strong Santonian affinities (Scasso *et al.* 1991; Olivero 2012; Crame 2019).

The succeeding 429 m thick Beta Member is characterized by a variety of coarser-grained lithologies, including grey/rusty-brown-weathering conglomerates, pebbly and tuffaceous sandstones, and a series of coquinas in the upper levels that comprise a distinctive trigoniid biofacies (Crame *et al.* 1991; Scasso *et al.* 1991). Besides species of both *Pterotrigonia* and *Eselaevitrigonia*, these shell beds have yielded other thick-shelled bivalves such as *Cucullaea* and *Eriphyla*, nuculanids, a distinctive aporrhaid gastropod and fragmentary hermatypic corals; at least 36 different benthic taxa have been recorded (Scasso *et al.* 1991). The upper levels of the Beta Member are also defined by the first appearance of the giant inoceramid bivalve *Antarcticeramus rabotensis*, which regularly reaches more than 1 m in length (Crame and Luther 1997).

In their review of the Campanian–Maastrichtian stratigraphy of the James Ross Island region, Crame *et al.* (1991) concluded that the Alpha and Beta members together comprised a distinctive lithostratigraphic unit that they formally renamed the Lachman Crags Member. This was in turn conformably overlain by a 250 m thick interval that was distinguished by the presence of a series of massive sandstone units interbedded with siltstones and minor coquinas. This was Olivero *et al.*'s (1986) Gamma Member, which was renamed the Herbert Sound Member and shown to pass conformably upwards into the Cape Lamb Member of the Snow Hill Island Formation on Cape Lamb, Vega Island (Crame *et al.* 1991; Pirrie *et al.* 1991) (Fig. 1). Nevertheless, this stratigraphic scheme has not been universally accepted and both Milanese *et al.* (2019, 2020) and Roberts *et al.* (2023) have suggested that the base of the Gamma/Herbert Sound Member is marked by a sequence boundary and included it in the Snow Hill Island Formation rather than the Santa Marta Formation. The Santonian–Campanian boundary has traditionally been placed in northern James Ross Island at approximately the 250 m level in the Alpha/Lachman Crags Member where it is marked by the first occurrence of the kossmaticeratid ammonite *Natalites rossensis* (Olivero *et al.* 1986; Crame *et al.* 1991; Olivero 2012) (Fig. 7). The position of this stage boundary will be considered further below when equivalent strata in southeastern James Ross Island are described.

SE James Ross Island and Snow Hill Island

Marambio Group

The giant inoceramid *Antarcticeramus rabotensis* can also be traced in southeastern James Ross Island from Rabot Point southwards through Redshaw Point and Punta Redonda (Fig. 1). Its occurrence here is such as to indicate that the Beta/Lachman Crags Member of the Santa Marta Formation is repeated in this region by a major ENE–WSW-trending fault (or faults) across central James Ross Island (Crame *et al.* 1991; Crame and Luther 1997). The 396 m thick sequence of bioturbated silty mudstones and muddy sandstones interbedded with occasional ash layers and conglomerates exposed at Rabot Point were assigned originally by Lirio *et al.* (1989) to the Rabot Formation, but then subsequently reduced to the Rabot Member of the Santa Marta Formation by Pirrie *et al.* (1997). These authors identified three informal mapping units, very similar to those of Lirio *et al.* (1989), with the lowermost one containing reworked fossils such as the heteromorph ammonite *Baculites bailyi* and the belemnite *Dimitobelus* (*Dimitocamax*) *cramensis*. Unit b is characterized by the giant inoceramids, a diverse ammonite assemblage and other bivalves, brachiopods and trace fossils (Lirio *et al.* 1989; Marenssi *et al.* 1992; Buatois *et al.* 1993). A palynomorph assemblage from the base of the section indicates an Early Campanian age, while those from units b and c are more indicative of an Early–Mid Campanian age (Pirrie *et al.* 1997). In addition, ammonites such as *Metaplacenticeras* and *Hoplitoplacenticeras* might suggest a Late Campanian age for the Rabot Member, but baculitid taxa indicate an older age in the Late Santonian–Early Campanian range. Palaeontological data in their entirety suggest that the Rabot Member ranges from Early–Late Campanian in age and correlates with both the upper Lachman Crags and lower Herbert Sound members of the Santa Marta Formation in northern James Ross Island (Lirio *et al.* 1989; Crame *et al.* 1991; Marenssi *et al.* 1992; Pirrie *et al.* 1997; Olivero 2012) (Fig. 7).

Neither the upper nor the lower contacts of the succeeding Hamilton Point Member are exposed, and it can only be estimated to be in the region of

400–560 m thick. It is characterized by monotonous, bioturbated, silty mudstones containing thin, inter-bedded ash layers and abundant oval carbonate concretions (Pirrie *et al.* 1997). Macrofossils are relatively abundant throughout the unit, and baculitid ammonites from the base of the section at Ekelöf Point have been referred to *Baculites rectus*, a species known to occur in the Lower–Middle Campanian Lachman Crags Member (Olivero *et al.* 1986; Crame *et al.* 1991). Ammonites from higher levels in the Hamilton Point Member generally have Late Campanian affinities (Olivero 2012), and this agrees well with palynological data from Ekelöf Point (Sumner 1992). Palynological dating throughout this member signifies a mid–Late Campanian age (Pirrie *et al.* 1997).

Largely on the basis of macrofossil content, the Hamilton Point Member is thought to overlie the Rabot Member but pass beneath the Sanctuary Cliffs Member on Snow Hill Island (Fig. 1). It was included in the Santa Marta Formation as a partial lateral equivalent of the Herbert Sound Member by Pirrie *et al.* (1997), but an alternative interpretation by both Milanese *et al.* (2020) and Roberts *et al.* (2023) shows the Hamilton Point Member to be the lowermost component of the overlying Snow Hill Island Formation, with its base represented by a sequence boundary. To some extent, assignment of this unit to either the Santa Marta or Snow Hill Island formations would seem to be arbitrary.

The Snow Hill Island Formation is a direct descendant of stratigraphic intervals previously identified as the 'Snow Hill beds', 'Snow Hill Island Series' and, in part, López de Bertodano Formation by various early workers (Crame *et al.* 1991; Pirrie *et al.* 1991, 1997 and references therein). The type sections are now identified to be on Sanctuary Cliffs and Spath Peninsula, Snow Hill Island, but it is also exposed on southwestern Seymour Island, Cockburn Island, Ula Point, Humps Island, Vega Island and The Naze and St Martha Cove regions on northern James Ross Island (Figs 1 & 7). This very wide distribution within the JRI group makes it a key interval for correlating Antarctic Late Cretaceous stratigraphy.

The basal Sanctuary Cliffs Member comprises a minimum 244 m thickness of grey, sandy mudstones with sparse fine sandstone interbeds (Pirrie *et al.* 1997). It is characterized throughout by the distinctive kossmaticeratid ammonite *Neograhamites* (Olivero 2012), plus a mixed benthic fauna of the abundant serpulid *Rotularia (Austrorotularia) fallax* and sporadic bivalves, echinoids, gastropods and decapod crustaceans (Pirrie *et al.* 1997). The first appearance of *Gunnarites*, one of the common ammonites of the Snow Hill Island Formation (Olivero 2012), occurs in the uppermost levels of the member and is associated with dinoflagellate cyst

taxa such as *Cerodinium diebelii*, *Isabelidinium cretaceum*, *Isabelidinium korojonense* and *Isabelidinium pellucidum* with strong Late Campanian–Early Maastrichtian age affinities (Marshall 1984; Helby *et al.* 1987; Pirrie *et al.* 1997). Such an assemblage also characterizes the upper Herbert Sound and lower Cape Lamb members of Vega Island (Pirrie *et al.* 1991).

Although the base of the succeeding Karlsen Cliffs Member is not seen, it forms a prominent 170 m thick unit along the southwestern flank of Spath Peninsula, Snow Hill Island (Fig. 1) where it is unconformably overlain by the Haslum Crag Member (Pirrie *et al.* 1997). The Karlsen Cliffs Member is a lateral equivalent of the Cape Lamb Member exposed on Vega Island and the geographically most widespread element of the Snow Hill Island Formation. It is essentially mudstone dominated, with the proportion of sandstones increasing up-section. These mudstones are dark, sulfurous and heavily bioturbated, with a distinctive rusty-orange banding due to weathered pyritic clay-rich horizons (Pirrie *et al.* 1997). A relatively abundant macrofauna comprises bivalves, gastropods, serpulids, decapod crustaceans and nautiloids, but particularly distinctive is an ammonite fauna containing taxa such as *Gunnarites antarcticus*, *Diplomoceras lambi*, *Jacobites crofti*, *Kitchinites darwini* and *Maorites* spp. (Olivero 2012 and references therein). This is very similar to the Cape Lamb Member ammonite fauna, but the Karlsen Cliffs Member is distinguished by the very high abundance of the chemosymbiotic bivalve *Conchocele townsendi*. These bivalves are associated with a series of shallow-water hydrocarbon seeps that occur again in the overlying López de Bertodano Formation on Seymour Island (Little *et al.* 2015). The dinoflagellate cyst flora from the Karlsen Cliffs Member is virtually identical to that of the Herbert Sound Member with very strong Late Campanian–Early Maastrichtian affinities (Pirrie *et al.* 1991, 1997 and references therein).

The base of the overlying Haslum Crag Member is an omission surface marked by framework-supported conglomerate, encrusting corals and bryozoans, and the presence of both glauconite and phosphate (Pirrie *et al.* 1997). This 195 m thick member, which is well exposed on Spath Peninsula and the southwestern tip of Seymour Island (Fig. 1), comprises grey-green muddy sandstones and yellow, clay-rich tuffaceous beds passing upwards into more mud-rich levels that retain the characteristic pale-green-weathering hue (Pirrie *et al.* 1997). A deep-red-weathering concretionary horizon 110 m above the base preserves a prolific macrofauna, including the ammonites *Gunnarites*, *Grossouvrites*, *Kitchinites* and *Maorites*, bivalves including 'Trigonia', *Lahillia* and *Solemya*, various gastropods, decapods and serpulids (Pirrie *et al.* 1997); 45 m

from the top of the unit there is a prolific single-species assemblage of *Conchocele townsendi*. The dinoflagellate cyst association of *Isabelidinium cretaceum* and *I. pellucidum* is indicative of a Late Campanian–Early Maastrichtian age (Marshall 1984; Helby *et al.* 1987; Askin 1988). On the south-eastern flank of Spath Peninsula and the southwestern tip of Seymour Island, the Haslum Crag Member is unconformably overlain by the basal pale-weathering mudstone unit of the López de Bertodano Formation (Montes *et al.* 2019).

Vega Island

Marambio Group

The bottom 52 m of the stratigraphic section exposed on the southeastern flank of Cape Lamb, Vega Island (Fig. 1) comprises a sequence of silty mudstones with minor sandstones passing up into more massive sandstones that can be closely matched with the Gamma/Herbert Sound Member of the Santa Marta Formation of northern James Ross Island (Pirrie *et al.* 1991). Macrofossils are sparse in this unit, but distinctive dinoflagellate cyst florules have strong Late Campanian–Early Maastrichtian affinities (Pirrie *et al.* 1991). The Herbert Sound Member

in turn passes conformably upwards into a 319 m thick sequence of highly bioturbated silty mudstones to silty sandstones containing abundant early diagenetic concretions. This is the Cape Lamb Member, which contains the characteristic *Gunnarites antarcticus* fauna and is a direct lateral equivalent of the Karlsen Cliffs Member on Snow Hill Island (Pirrie *et al.* 1991, 1997; Olivero *et al.* 1992; Olivero 2012; Roberts *et al.* 2014, 2023) (Fig. 8). A single inverse-graded granule to cobble conglomerate in the upper levels of the Cape Lamb Member is hard to interpret. Its lower surface is erosional, can be traced laterally for at least 4 km and marks a change in both macrofaunas and palynofloras (Pirrie *et al.* 1991). It has been used to subdivide the Cape Lamb Member into upper and lower components (Roberts *et al.* 2023), but whether it correlates with more distal sequence boundaries, such as at either the base or top of the Haslum Crag Member, is currently unknown. Both macro- and micro-palaeontological dating indicate that the Cape Lamb Member has a Late Campanian–Early Maastrichtian age range. Recalibration of strontium isotope dates from a level between 81.5 and 96.5 m above the base strongly suggests that the Campanian–Maastrichtian boundary occurs in the lower to mid-levels of the Cape Lamb Member (Crame *et al.* 1999; Roberts *et al.* 2023) (Fig. 7; see also below).

Fig. 8. A 200 m section of silty mudstones and fine- to medium-grained sandstones of the Cape Lamb Member, Snow Hill Island Formation; Humps Island, James Ross Island group. The beds show a gentle dip SE and are characterized by abundant early diagenetic concretions. Source: photo from BAS Photographic Collections.

The 111 m thick Sandwich Bluff Member exhibits a range of lithologies, including thin conglomerates, pebbly sandstones, sandstones and mudstones, and contains slump sheets, hummocky cross-stratification and rare rootlet horizons (Pirrie *et al.* 1991). The various lithologies have a strong volcaniclastic component and were the product of both coeval arc volcanism and tectonic activity during a significant phase of marine regression. In their original description of the Sandwich Bluff Member, Pirrie *et al.* (1991) believed the base of the unit to be discordant, but this view has been questioned by later investigators (Roberts *et al.* 2014, 2023). Ammonites referable to *Maorites densicostatus* from the lower levels of the member and a dinoflagellate cyst assemblage characterized by *Manumiella druggii* are strongly indicative of the Late Maastrichtian stage (Pirrie *et al.* 1991; Olivero *et al.* 1992; Bowman *et al.* 2012; Olivero 2012). The Sandwich Bluff Member has been assigned traditionally to the López de Bertodano Formation (Fig. 7).

In the lower levels of the Sandwich Bluff Member, there is not a particularly good correlation between palynostratigraphy and magnetostratigraphy (Roberts *et al.* 2023), but between the informal stratigraphic intervals 4–16 of these authors there is a distinct reversed polarity that most likely correlates with Global Polarity Timescale (GPTS) Chron C29R, and the topmost level 16 with C29N. Such an interpretation supports the finding of Roberts *et al.* (2014) of a K–Pg boundary at the very top of the member, and this is further backed up by a robust detrital zircon age of 66.3 $\pm$ 1.1 Ma for the top of interval 16 (Roberts *et al.* 2023). A sequence boundary defined by both sedimentology and palynology close to the top of the member may correlate with an irregular base to the K–Pg boundary section on Seymour Island, but this needs further investigation. In any event, there may be no more than 2–3 m of basal Paleocene strata exposed at Sandwich Bluff, and this most likely represents some form of extremely shallow-water, reduced salinity palaeoenvironment (Roberts *et al.* 2023).

Seymour Island

Marambio Group

The discordant base of the López de Bertodano Formation can be traced from eastern Spath Peninsula onto the southwestern tip of Seymour Island (Fig. 1). Here, there is a 125 m thick basal unit of poorly lithified, pale-weathering mudstones that are sparsely fossiliferous and linked to a shallow-water, tidally influenced setting (Olivero *et al.* 2008). While it is possible that this basal unit could be distinguished as being of member status, it has proved impossible to provide a meaningful lithostratigraphic subdivision of the succeeding 715 m. This comprises a remarkably uniform sequence of rusty-brown to tan and grey muddy siltstones that are intensely bioturbated and often appear structureless (Feldmann and Woodburne 1988; Macellari 1988; Crame *et al.* 2004). Dense levels of small- to medium-sized early diagenetic concretions become more common up-section, as do pale to dark-green-weathering glauconitic sandstones. Sadler's (1988) original informal mapping units Klb 1–9 are still in widespread use, with Klb 2–6 coinciding approximately with Macellari's 'Rotularia Units' and Klb 7–9 with the more fossiliferous 'Molluscan Units' (Feldmann and Woodburne 1988). There is a slight coarsening-upward trend throughout the López de Bertodano Formation (Crame *et al.* 2004).

Rotularia (Austrorotularia) fallax is the dominant faunal element in the Rotularia Units, and there are also records of the ammonites *Diplomoceras lambi* and *Maorites* (at least two species), numerous echinoid spines, as well as occasional bivalves and decapod crustaceans (Macellari 1988). There is also a single horizon in which the belemnite *Dimitobelus* (*Dimitobelus*) *seymouriensis* is common, but this stratigraphically highest record of belemnites in Antarctica is still some 640 m beneath the K–Pg boundary (Doyle and Zinsmeister 1988; Crame *et al.* 2004). The more prolific fauna found from approximately the mid-levels of unit Klb 6 contains a range of ammonites, together with three gastropods that are then present right up to the K–Pg boundary: *Cryptorhytis philippiana*, *Taioma charcotiana* and 'Cassidaria' *mirabilis*. However, the benthos is dominated by bivalves, with common genera including: *Solemya*, *Austrocucullaea*, *Pinna*, *Entolium*, *Acesta*, *Pycnodonte* (*Phygraea*), *Eselaevitrigonia*, *Oistotrigonia*, *Thyasira*, *Lahilla*, *Surobula*, *Goniomya* and *Thracia* (Zinsmeister and Macellari 1988). The mid-levels of Klb 7 are marked by the sudden incoming of the desmoceratid ammonite *Kitchinites*, the kossmaticeratid *Grossouvrites gemmatus* and at least two species of *Maorites* (Olivero 2012; Witts *et al.* 2015). Kossmaticeratid ammonites predominate in the topmost levels of the formation, with *Maorites densicostatus* and *Grossouvrites gemmatus* being particularly common (Macellari 1986). *Diplomoceras*, *Anagaudryceras*, *Pseudophyllites* and *Kitchinites* are also present, and in the uppermost 200 m of the section there is the sudden appearance of large pachydiscids referred to an informal *Pachydiscus riccardi*–*Pachydiscus ultimus* Biozone (Macellari 1986; Crame *et al.* 2004; Olivero 2012; Witts *et al.* 2015). Other distinctive taxa present in the very latest Maastrichtian of Antarctica include

the bivalves *Lahillia larseni*, *Pycnodonte* sp., *Phelopteria feldmanni*, *Pulvinites antarctica* and *Acesta shackletoni* (Macellari 1988; Zinsmeister and Macellari 1988; Crame *et al.* 2004).

In the lower levels of the López de Bertodano Formation there is an important change in the dominance of marine dinoflagellate cyst florules from *Isabelidinium cretaceum* to *Manumiella seymourensis* (Askin 1999). The unchanged nature of the latter throughout the mid- to upper levels of the formation is an important indicator of relatively stable shelf conditions in the James Ross Basin throughout the greater part of the Maastrichtian (Askin 1999). A pronounced *Manumiella* spike occurs immediately prior to the K–Pg boundary, and throughout the López de Bertodano Formation there are three distinct acmes of the small chorate dinoflagellate cyst *Impletosphaeridium clavus* (Bowman *et al.* 2013*a*; Amenábar *et al.* 2019). These may represent seasonal blooms that in turn have distinct palaeoclimatic implications (Bowman *et al.* 2013*b*; see below).

Placement of Upper Cretaceous stage boundaries in Antarctica

The precise placement of Upper Cretaceous stage boundaries in Antarctica is still under investigation (Fig. 7). There has been a general acceptance that the Turonian–Coniacian boundary can be placed at approximately the base of the Hidden Lake Formation, but the Coniacian–Santonian boundary is less certain. Whereas there is some palaeontological evidence to place it at the base of the Santa Marta Formation, $^{87}Sr/^{86}Sr$ isotope stratigraphy suggests a position some 150 m higher (McArthur *et al.* 2000), and magnetostratigraphy one as low as the mid-levels of the Hidden Lake Formation (Milanese *et al.* 2020). Similarly, whereas magnetostratigraphy suggests that the Santonian–Campanian boundary could be as low as the mid-Alpha Member of the Santa Marta Formation (Milanese *et al.* 2020), ammonite stratigraphy would place it well within the Beta Member, and this strengthens correlation with the Rabot Member of southeastern James Ross Island (Pirrie *et al.* 1997; Olivero 2012). Magnetostratigraphy is more definitive in affirming that the Campanian–Maastrichtian boundary occurs in the lower Cape Lamb Member, Vega Island and mid- to upper levels of the Sanctuary Cliffs Member, Snow Hill Island, in both cases in very close proximity to the first occurrence of the ammonite *Gunnarites* (Crame *et al.* 2004; Milanese *et al.* 2019). These correlations indicate that the Campanian–Maastrichtian interval in Antarctica is represented by a total thickness well in excess of 2 km of strata (Fig. 7).

The K–Pg boundary in Antarctica

The K–Pg boundary on Seymour Island is marked by a prominent 5–6 m thick glauconitic interval that caps a small NNE–SSW-trending scarp (Fig. 9). At the base of the interval is a 1 m thick concretionary glauconitic sandstone that can be traced laterally for some distance and shown to be slightly discordant. It is in turn 50–100 cm above the last known ammonites and approximately coincident with a small iridium spike (Zinsmeister *et al.* 1989; Elliot *et al.* 1994); the base of this sandstone marks the position of the K–Pg boundary in Antarctica. This Lower Glauconite passes up into a 2–3 m interval of siltstones and fine sandstones that exhibits an intermittent 'fish kill horizon' in its lower levels (Zinsmeister 1998). This unit is in turn capped by an Upper Glauconite, followed by a significant lithological change to recessive, grey-brown siltstones/fine sandstones bearing numerous small early diagenetic concretions (the Lower Danian Unit 10 of Montes *et al.* 2019).

Recent recognition of a second K–Pg boundary section on Vega Island at the very top of the Sandwich Bluff Member of the López de Bertodano Formation (Roberts *et al.* 2023) has heightened the need for a more precise correlation of terminal Cretaceous strata across the JRI group basin (Figs 1 & 7).

The extended and well-exposed K–Pg boundary on Seymour Island has been the subject of intense recent field and laboratory investigations. The view that, at least in the marine realm, the extinction was the product of a single catastrophic event of equivalent intensity to that seen in lower latitudes (Witts *et al.* 2016; Whittle *et al.* 2019) has not gone unchallenged. There has been a persistent school of thought that the Seymour Island K–Pg section in fact displays evidence of a two-fold extinction event, with one due to a bolide event and the other to the effects of Deccan Traps volcanism (Tobin *et al.* 2012; Petersen *et al.* 2015; Tobin 2017; Mohr *et al.* 2020). There may be no immediate resolution of this important debate.

Discussion

Less than 50 years ago, our knowledge of the Cretaceous stratigraphy of Antarctica was very incomplete. On the western (forearc) flank of the Antarctic Peninsula we were aware of a thick succession of J–K boundary strata and, by analogy with the then better-known sequences in Patagonia, there appeared to be an Aptian–Albian interval too. But the total thickness of beds was unknown, and it was difficult to place them into a meaningful regional context. The sequence on the eastern (back-arc) flank of the peninsula was even less well known, and for a

Fig. 9. The K–Pg boundary exposed in central Seymour Island. There is a shallow SE regional dip, and the position of the boundary is indicated in the centre-foreground. Approximately 1100 m of Paleogene strata are exposed above the boundary, with the topmost beds forming the plateau (or meseta) on the horizon. Source: photo by J.A. Crame.

long time the entire Cretaceous sedimentary succession was assigned to the mid–Late Campanian, very largely on the misidentification of a single ammonite. Today, we are in a much better position, and the legacy of a whole series of recent stratigraphic, sedimentological and palaeontological studies is a 7km+ thick detailed Cretaceous stratigraphy in a unique polar position. Undoubtedly, some gaps in our knowledge still remain, particularly in the latter part of the Lower Cretaceous sequence, but the framework established is still sufficient to promote a wide range of regional studies using Antarctica as a stable reference point through time. We will investigate the potential of two of these here, using Antarctica to examine Cretaceous trends in both climate and biodiversity change.

Climatic trends through the Antarctic Cretaceous

We often view the greenhouse–icehouse transition leading to the modern world as an early Cenozoic phenomenon, with widespread tropical conditions in the Early–Middle Eocene gradually retracting in the Late Eocene before a pronounced temperature drop at the Eocene–Oligocene boundary, which marks the actual junction between the two phases and corresponds to the first appearance of glacial diamictites at the continental margin (Zachos *et al.* 2008). Nevertheless, it is now apparent that the

true global temperature peak over the last 150 Myr occurred at 92 Ma, and the interval 92–84 Ma has come to be known as the Cretaceous Thermal Maximum (KTM) (Huber *et al.* 2018; Tierney *et al.* 2020). At this time latitudinal temperature gradients were greatly reduced and global sea-level was up to 170 m higher than at present (O'Connor *et al.* 2019). From the KTM, which is almost exactly coincident with the Turonian–Santonian stages, global temperatures declined gradually to the K–Pg boundary before increasing again in the Middle-Danian and rising to a Cenozoic maximum in the Middle Eocene (Miller *et al.* 2020; Westerhold *et al.* 2020; Petrizzo *et al.* 2022).

Unfortunately, we have little firm palaeoclimate data for the earliest Cretaceous stages in Antarctica, apart from some preliminary oxygen stable isotope studies on belemnites, which indicate a wide range of temperature values (Ditchfield *et al.* 1994; Pirrie *et al.* 1995). However, in Australia, which was still attached to Antarctica at a high palaeolatitude, there is evidence from the northern Flinders Ranges of two significant phases of glaciation: one in the Middle Valanginian and the other in the Early Aptian (Alley *et al.* 2020; Alley and Hore 2022). To this can be added stable isotope palaeothermometry on both planktonic and benthic foraminifera from the Falkland Plateau, which shows a pattern of gradual warming from approximately the Early Albian up to the KTM (Bice *et al.* 2003; Huber *et al.* 2018). Such a trend of progressive global warming through

the Early Cretaceous would fit well with the occurrence of the Late Albian fossil forests on Alexander Island. Even though some account must be taken of the broad-scale shallowing-upward trend throughout the FBG, these conifer-dominated forests are mature and taxonomically diverse, suggesting warm climates during the middle parts of the Cretaceous at 75° S palaeolatitude (Cantrill and Poole 2012).

The effects of the KTM on Antarctic terrestrial biota have been further illustrated recently by the discovery of key elements of a temperate lowland rainforest at a palaeolatitude of 82° S (Klages et al. 2020). These were recovered from a shallow borehole in the Amundsen Sea, western Antarctica and are precisely dated to the 92–83 Ma interval. Some 62 spore and pollen taxa, including podocarp and araucarian conifers and tree ferns, indicate a mean annual temperature of 13°C only some 900 km from the palaeo-South Pole (Klages et al. 2020). Mean annual temperatures of 17–19°C were deduced from analysis of fossil floras of Coniacian to Early Campanian age from the James Ross Basin and attest to low latitudinal temperature gradients during the KTM (Hayes et al. 2006).

A fossil flora of Santonian age (c. 85 Ma) is preserved as charcoalified remains in sediments at Table Nunatak on the eastern side of the Antarctic Peninsula (Fig. 1). The fossils represent burnt vegetation washed into the marine back-arc basin and consist of temperate forest plants, including ferns and conifers (Eklund et al. 2004). Angiosperms are also present, including rare preservation of fossil flowers, similar to those in the families Siparunaceae, Winteraceae and Myrtaceae, which live in Southern Hemisphere temperate forests today (Eklund 2003).

There are indications too, of warm-water, possibly subtropical, temperatures from the unusual marine macrofauna characterizing both the mid-levels of the Coniacian Brandy Bay Member of the Whisky Bay Formation and the upper levels of the Santonian Beta/Lachman Crags Member of the Santa Marta Formation. The latter contains a hermatypic coral, a further bakevelliid bivalve and gastropods assigned to 'Nerinea' and 'Terebra' (Scasso et al. 1991). A full taxonomic revision of these warm-water faunas is urgently required.

General indications of cooler Campanian–Maastrichtian climates in the Antarctic Peninsula region have come from both palaeobotanical and geochemical investigations (Dingle and Lavelle 2000; Francis and Poole 2002). In addition, a detailed palynological analysis of the Maastrichtian López de Bertodano Formation detected three conspicuous abundance peaks of the dinoflagellate cyst Impletosphaeridium conspicuous that were attributed to the accumulation of resting cysts from seasonal blooms over extended periods of time (Bowman et al.

2013b). These blooms were in turn linked to winter sea ice decay and thus the implication that the climate through the greater part of the Maastrichtian was cold enough to initiate short-term glaciations (Bowman et al. 2013b). Both geochemical and modelling studies have supported the presence of winter sea ice in Antarctica (De Conto et al. 2007; Robinson et al. 2010), but so far there is no evidence of either Maastrichtian tillites or ice-rafted debris (Huber et al. 2018). Cretaceous outcrops with such evidence may well be preserved under the present ice sheets. A phase of very latest Maastrichtian warming may be attributable to the onset of Deccan Traps volcanism (Tobin et al. 2012; Bowman et al. 2014; Petersen et al. 2015).

Biodiversity trends through the Antarctic Cretaceous

Recent advances in the Cretaceous stratigraphy of Antarctica provide a new impetus for re-examining taxonomic diversity patterns through the period. There is evidence to suggest that, on a global scale, most major clades of plants and animals expanded throughout the Cretaceous (Crane et al. 1995; Close et al. 2020), but less is known about regional patterns, and especially so at high latitudes.

Initial studies in the marine realm have indicated that belemnites and inoceramid bivalves, two groups that became extinct globally at the K–Pg boundary, are drastically reduced in numbers in Antarctica after the Early Campanian (Crame et al. 1996). The same is true of numbers of ammonite species, which reached their acme in the Lower Campanian Beta/Lachman Crags Member of the Santa Marta Formation (Zinsmeister and Feldmann 1996); ammonites are abundant in both the Snow Hill Island and López de Bertodano formations but not taxonomically diverse. Added to this is the fact that early studies of the Beta/Lachman Crags Member benthic fauna indicate that it too is more diverse than those in either of the two succeeding formations. Whereas it is tempting to link this trend of Campanian–Maastrichtian declining diversity in Antarctica to falling temperatures, some caution must be exercised in taking this step. The link between ambient temperature and diversity trends through time is a complex one (Jablonski et al. 2017) and other key factors may also be involved. For example, hydrocarbon seepage was common throughout the Antarctic Maastrichtian and almost certainly exerted a strong influence on the composition of marine faunas (Little et al. 2015; Mohr et al. 2020).

In the terrestrial realm, podocarp–araucarian conifer forests with an understory of bennettites, abundant ferns and bryophytes had spread from

southern South America to the Antarctic Peninsula by the Albian–Cenomanian (Cantrill and Poole 2012). Palynological analysis has demonstrated the presence of early angiosperms assigned to the Proteaceae and *Ilex* in the Turonian, and they were joined by diverse eudicots in the Coniacian (Dettmann and Thomson 1987; Dettmann 1989; Askin 1992). The Turonian–Coniacian was a time of rapid modernization of the Antarctic flora, as revealed by both the Whisky Bay and Hidden Lake formations on James Ross Island, the Late Santonian Table Nunatak charcoalified flora with fossil flowers (Eklund 2003; Eklund *et al.* 2004) and a Cenomanian–Lower Campanian sequence of fluvial sediments from Williams Point, Livingston Island (Fig. 1) (Chapman and Smellie 1992). Macrofloral analysis has revealed that 70% of the Hidden Lake Formation flora and 71% of the Santa Marta Formation flora is composed of angiosperms (Hayes 1999). This dominance of angiosperms in the James Ross Basin continues unabated into the Campanian–Maastrichtian (Bowman *et al.* 2014). Taxa that proliferated at this time include Nothofagaceae, Gunneraceae, Aquifoliaceae and Loranthaceae (Dettmann and Thomson 1987; Askin 1989, 1992; Dettmann 1989; Cantrill and Poole 2012; Bowman *et al.* 2014). It is clear that this major diversification of angiosperms was not affected by global cooling.

It should be emphasized that we are still some way from establishing a comprehensive Cretaceous biostratigraphy of Antarctica. The full ranges of key taxonomic groups, such as ammonites, belemnites and inoceramid bivalves, have yet to be replotted on the unified stratigraphic scheme (Figs 2 & 7), but perhaps the most urgent task is to develop a complete palynostratigraphy for the period. Many individual studies have been carried out to date but they have yet to be integrated into a standardized scheme, and this is especially so in the Late Cretaceous.

Refinement of key strategic ages also has considerable implications for future studies. For example, the revised chronostratigraphy for the Sandwich Bluff Member, Vega Island strongly suggests that the fossil water bird *Vegavis iaai* from that unit may be the oldest crown group bird yet identified (Roberts *et al.* 2023). The type and referred specimens of this taxon have now been shown to lie within Chron C31N, which equates to a numerical age of 69.2–68.4 Ma; as such, this strongly suggests that modern birds had an origin within the Late Cretaceous of southern Gondwana (Claramunt and Cracraft 2015; Roberts *et al.* 2023) (Fig. 10). Further dating of the Cape Lamb Member on Vega Island also indicates that one of the key Antarctic dinosaur faunas is Late

Fig. 10. Artist's impression of mixed temperate *Nothofagus*–podocarp–Proteaceae rainforest on the eastern flank of the Antarctic Peninsula during the latest Cretaceous (Maastrichtian). *Lagarostrobus* (Tasmanian Huon Pine) occupy the super-wet habitats along riverine margins; freshwater pools hosted aquatic ferns and green algae with bordering wetlands filled with diverse mosses and ferns. At higher altitudes, open heath-like vegetation may have grown beyond the tree line of montane araucarians. Both avian and non-avian dinosaurs shown. Further explanation given in Bowman *et al.* (2014). Source: © James McKay.

Campanian in age rather than Early Maastrichtian (Lamanna *et al.* 2019).

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Author contributions JAC: conceptualization (lead), investigation (equal), methodology (lead), writing – original draft (lead), writing – review & editing (lead); **JEF**: writing – review & editing (equal).

Data availability Data sharing is not applicable to this article as no datasets were generated or analysed during the current study.

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