Former extension of the East Antarctic Ice Sheet

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Knowledge about the Quarternary geology of the Australian sector of Antarctica is very incomplete. Scattered observations of glacial deposits in that area, made during the ANARE 6 expedition in 1987, indicate that the inland ice had formerly a considerably wider extension than today. The ice was more than 200 m thicker, probably of the order 1,000 m. This maximum stage cannot yet be dated, but conditions seem to favour a late Wisconsin-Weichselian maximum age. However, a much higher age cannot be excluded.

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The Antarctic ice sheet is the world's largest ice body, containing roughly 24×10^6 km^3 (Denton & Hughes 1981). Since the early exploration of Antarctica it has been noted that the ice sheet has been more extensive. Records of observations are scattered throughout the literature. In some places, mainly in West Antarctica and around the Ross Sea, there are some datings of the maximum extension, which may have occurred simultaneously with the maximum of the late Pleistocene ice sheets of the northern hemisphere. However, this hypothesis is far from established.

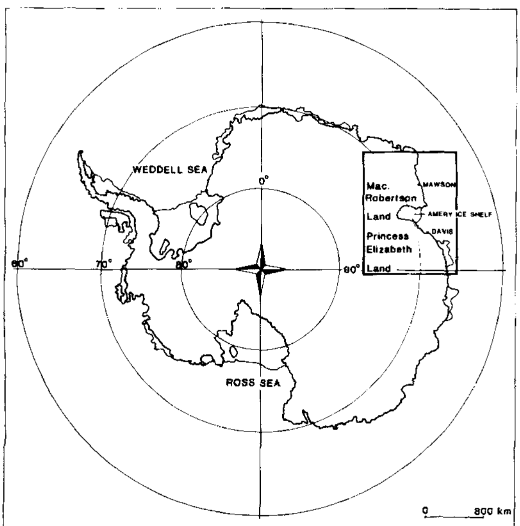

Fig. 1. Location of investigated areas. The square marks the position of map, Fig. 2.

A pre-Pleistocene maximum is recently inferred (e.g. Denton et al. 1984; Höfle 1987).

About 80% of the Antarctic ice sheet belongs to the East Antarctic part. The former extension of the ice is less well known in this sector than in West Antarctica (see Denton & Hughes 1981, e.g. p. 380). Because even a relatively small increase in volume and areal extension is important for the total water balance of the Earth (e.g. Nakada & Lambeck 1988), the following observations are significant, although incomplete and unsystematic. This applies especially to the areas discussed in this paper, which are virtually unknown from the point of view of Quaternary geology – with the exception of the studies referred to from Vestfold Hills.

The observations were made in the coastal region of Australian Antarctic Territory, between 60° and 80°E (Figs. 1, 2). They were undertaken during the 1987 Australian National Antarctic Research Expedition 6.

Vestfold Hills

The Vestfold Hills in Princess Elizabeth Land (the region east of Prydz Bay, Fig. 1) have an ice-free area of about 400 km^2 (Fig. 3). Their boundary to the east is the ice sheet and to the south the Sørsdal Glacier tongue, extending from the ice. To the northwest, there is the coastline of Prydz Bay, which is very irregular, with deep, narrow fjords and numerous islands. The land is flat, with a maximum altitude of only about 150 m, and consisting of Precambrian gneisses cut by numerous dolerite dikes (Adamson & Pickard 1986b).

The area is extremely dry and can be described as a cold desert. Evidence of weathering by frost and salt action as well as wind erosion is seen all over the area. Rocks and moraines are strongly affected by cavernous weathering, which seems to proceed rapidly. Therefore the observed moraines could be relatively young. It may be assumed that older moraines, if they existed, as well as glacially sculptured rocks have been removed.

In spite of the intense weathering, Derbyshire & Peterson (1978) and Adamson & Pickard (1983) have been able to identify traces of glaciation. By means of fossils contained in various deposits, Adamson & Pickard (1986a) have obtained radiocarbon dates for some former boundaries of the ice. It seems clear that the entire area was ice-covered during the maximum glaciation, at least before 8,000 B.P. A younger ice marginal limit was also identified by Adamson & Pickard (1986a) and dated between 3,000 and 1,500 B.P. In some areas, mainly along the margin of the Sørsdal Glacier, a recent retreat of the ice margin is evident. One of the most conspicuous features is the hummocky moraine area called the Flanders Moraine by Adamson & Pickard (1986a). The author did not make any attempt at further dating of these or other ice-margin positions, but could, in general, verify the features described by the above-mentioned authors. From two points of view these observations are important in the present context.

An extension of the ice sheet to the present coastline and across the islands implies considerably thicker ice than today. If we assume that the ice cap had the same gradient at its marginal zone as today, which is a reasonable but not proved assumption, the thickness must have been at least 1,000 m greater at the glacial maximum (Adamson & Pickard 1986a). Certainly, the profile of the ice marginal zone may have been different due to the morphology of the substratum. For instance, the ice flow pattern observed and the occurrence of shear-moraines at the present margin in the area east of Bisernoye Lake (Fig. 3) indicate a rather high surface beneath the ice, possibly causing a steeper ice-marginal slope than could be expected over the flat Vestfold Hills farther west. Also the proximity of the Amery Ice Shelf may cause anomalies in the gradients of the surface and base of the ice. Therefore, the figure 1,000 m should be considered a maximum value for the increase in thickness.

The second important aspect regarding former ice extension concerns rock weathering. As mentioned above, cavernous weathering is extensive in the Vestfold Hills (Fig. 4). No obvious difference in weathering was observed within the areas deglaciated before 8,000 B.P. and between 5,000 and 1,000 B.P. In contrast, the area being deglaciated today and a narrow zone along the ice margin showed no or insignificant signs of cavernous weathering. The northern margin of Sørsdal Glacier at Chelnok Lake (Fig. 3) offers a good example. The recent moraine, which can be estimated to have been formed only a few decades ago (Pickard 1983), shows no sign of weathering. The next moraine, very close to the youngest, shows incipient weathering, and already the third, only some 100 m outside the first, is strongly weathered.

There seems to be no other way to explain the differences in weathering than by differences in age between the moraines. Because we do not yet know the rate of weathering well enough, the lack of weathering does not offer a means of dating, but it has some significance for the following discussion.

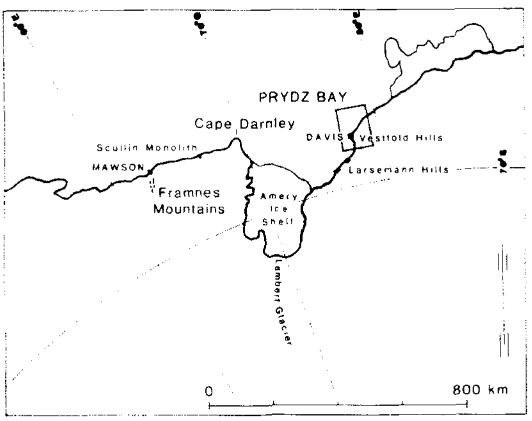

Fig. 2. Localities visited in the area around Prydz Bay. The square marks the position of map, Fig. 3.

Larsemann Hills

Larsemann Hills is a small ice-free area on the coast of Prydz Bay, 120 km southwest of Vestfold Hills (Fig. 2). No maps or air photos existed of

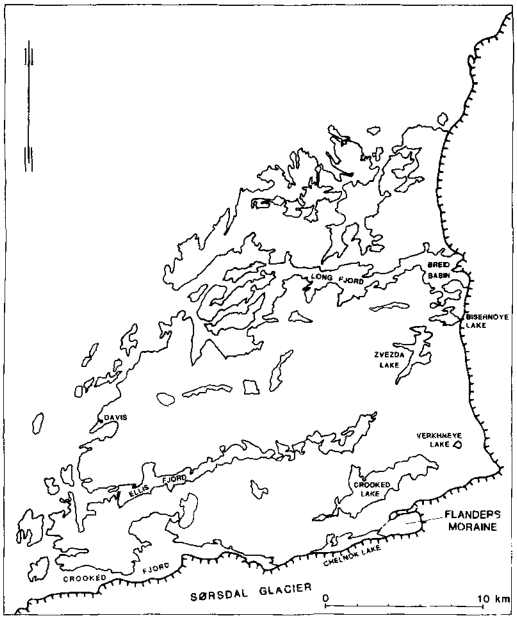

Fig. 3. Localities in the Vestfold Hills.

the area, and the altitude of the hills is unknown. However, the highest hills seem to reach an altitude of the order of only 50 m, and certainly less than 100 m. The distance between the coast and the inland-ice margin is about 3 km. A number of low islands fringe the coast. Shallow valleys extend from the ice sheet to the coast.

No climate record exists for this area, but in general it seems to be very similar to Vestfold Hills. The area is extremely dry with no vegetation except scattered lichens. Cavernous weathering is observed at the area visited, which did not include the ice-marginal zone. Salt patches were recorded in many places.

The intense weathering has released masses of debris, which move downslope to form slope deposits similar to moraines or beaches. A closer study will probably reveal the true nature of the debris flows. Their monolithic character and some granulometric differentiation along the flow paths are indicative.

In contrast to the debris flows, a few deposits were identified as moraines. Across a valley south of Law Base (established 1987) two ridges were observed. Their lithology is less indicative in the environment of Precambrian gneisses, but nevertheless there is a noticeable difference between the moraine ridges and the weathered residual. The latter shows clear boundaries between different lithologies, while the till in the ridges is more mixed and contains much better rounded rocks. The position and shape of the moraines show that they cannot have accumulated by mass movement from adjacent slopes. The only way to account for their deposition seems to be by glacier transport. A hypothetical alternative could be a thin surface layer of till upon bedrock sills. However, patterned ground with upwelling fine-grained material and tundra polygons indicate a certain depth to the bedrock. Typically, the moraines were the only positions where patterned ground was recorded. The weathered residual seems to be less susceptible to frost action and thus less suitable for formation of patterned ground.

The observations confirm that Larsemann Hills have been ice-covered. Considering the distance between the ice margin and the coast, the ice must have been at least 150 m thicker than today. A comparison between the weathering in this area and in Vestfold Hills suggests that the widest extension of the ice in Larsemann Hills at least occurred earlier than 1,500 B.P. No further dating has been possible, but a glacial maximum of more than 8,000 B.P. may be inferred.

Northeastern Mac.Robertson Land

On the coast of northeastern Mac.Robertson Land (the region west of Prydz Bay, Fig. 1), there are some outcrops of bedrock along the ice margin and as nunataks. One of the coastal outcrops is Scullin Monolith, about 280 km east of Mawson Station (Fig. 2). The Monolith consists of a single outcrop, rising about 400 m above sea-level, about 2 km in length and 200 to 300 m wide. The inland ice flows around the rock to form low ice cliffs on either side. Thus, the Monolith rises up to 350 m above the adjacent ice margin, but the ice surface is much higher on its upglacier side. The height is about 100 m below the Monolith's summit.

Scullin Monolith is strongly eroded and polished by katabatic winds. Cavernous weathering at its summit occurs only on the lee-side and in protected positions. This implies that time has been sufficient to allow strong weathering, even if the weathered residual is usually removed by wind. It also implies that weathering in combination with wind action may have removed at least an essential part of a till cover, if there has been any. Scattered erratics and thin till patches occur all over the Monolith, up to its highest part. Their distance of transport cannot be estimated, because of the variations within the local bedrock of gneisses. Nevertheless, the rocks are often different from their immediate substratum. They have a much fresher appearance than the bedrock because of lack of the brownish desert varnish characterizing the bedrock surface. The boulders are mostly subrounded, and in this respect different from frost-shattered debris (Fig. 5).

The erratics are good evidence that Scullin Monolith has been completely covered by the ice sheet. The ice must at some stage have been at least 100 m thicker than today. So far we have no possibility of dating this stage. Qualitatively, we may infer that this wider extension of the ice did not take place very recently. This is indicated by the desert varnish and by lichen growth. Lichens occur on the Monolith down to the present ice surface (Fig. 6). They are much more abundant than in the desert areas described above. Among the lichens, a black *Buellia* species is most abundant. Unidentified species of orange and white lichens are also common, as well as the green algae *Prasiola crispa*. These lichens and algae are fast-growing and in general less suitable for lichenometric dating. Together with the extensive and well-developed desert varnish, caused by the biological effect of algae and lichens (Scheffer et al. 1963), vegetative cover makes a recent deglaciation improbable. It is not possible to tell whether this means that deglaciation occurred 18,000–13,000 years ago or later. Even a much earlier retreat is possible, and more likely.

The geomorphology of the northeastern coast of Mac.Robertson Land is also informative. The continental margin reaches down to the depth of 500 m (Quilty 1985). Some morphological features are conspicuous on echo-sounder records. The shelf off Scullin Monolith is characterized by extremely broken surface forms. Numerous narrow steep-sided gorges are incised in the bedrock, which is in this way divided into closely spaced pinnacles. The morphology is similar to

Fig. 4. Cavernous weathering at North Portal, 7 km east of Davis Station.

Fig. 5. Erratic boulders (light coloured) near the summit of Scullin Monolith.

landforms created by weathering and unaffected by glacial erosion, seen on a smaller scale in adjacent land areas. The information recorded is insufficient to define an upper limit of this morphology. If it could be done, it would be possible to calculate a lower limit of glacial erosion and thus the thickness of the ice in relation to different sea-level positions. At this stage we can only conclude that the ice has not been thick enough to exert glacial erosion below a couple of hundred metres below the present sea-level. Considering the lower sea-level at the glacial maximum (e.g. Bloom et al. 1974), this means that the ice cap at its maximum extension in this region was not more than c. 200 m thicker than today.

Only in one area, on the western side of the Fram Bank close to the coast at Cape Darnley at the northwesternmost end of Prydz Bay, the shelf shows a feature which strongly resembles a U-shaped glacial valley. There the surface of the shelf is much more even and situated at a shallow depth, or 100 to 200 m below sea-level. The flat surface may be the result of glacial or other erosion, or of a cover of glacial and/or marine deposits. The U-shaped valley, reaching a depth of about 300 m, is difficult to explain other than by glacial erosion. Consequently, we obtain a picture of a glacier tongue at least 200 m thick (considering the low sea-level of the glacial maximum), extending from a grounded inland-ice margin. Whether this margin reached the surrounding shelf cannot be determined on the basis of available information. The general pattern, however, is similar to what can be seen today in many places along the East Antarctic coast.

The Cape Darnley-Scullin Monolith area provides evidence that the ice cap has been 200 to 300 m thicker than today. This corresponds to a glacial maximum of unknown age.

Mawson area – Framnes Mountains

The Framnes Mountains, Mac.Robertson Land (Fig. 2), consist of several mountain ranges, protruding through the ice sheet. The southernmost nunataks occur on 68°10'S, about 55 km inland from Mawson Station. These landforms are shaped mainly by weathering, creating sharp ridges and peaks, the lower parts of which are scree-covered. Frost shattering together with wind action are the major erosion forms. The scree slopes often extend on to the ice surface, making difficult the distinction between till and weathered residual (Lundqvist 1989).

In many places in these mountains, glacial deposits may be identified in spite of the screes. The glacial deposits are clearly recognized where the topographic position excludes accumulation by mass movements, for instance, where they form moraines some distance away from the mountain slopes. Also, close to the slopes or on the mountain sides, erratic boulders, striated and rounded rocks, are an indication of accumulation by glacier ice. These erratics may derive from higher-lying deposits, now obliterated by mass movements, and may not be diagnostic for the present-day deposits. However, for the identification of former high positions of the ice surface this is of minor importance.

On the northern end of Central Masson Range, 23 km south of Mawson Station, there is a large lee-side accumulation (Lundqvist 1989). It extends from the ice surface towards the northernmost peak in this range. The height of the peak is unknown, but it is estimated from helicopter inspection to be 200 to 300 m above the ice surface. Thus, the depositing ice must have been more than 200 to 300 m thicker than today. The glacial origin of the deposit seems to be beyond any doubt, even if it almost exclusively consists of the local bedrock. The shape (Fig. 7) excludes accumulation only by mass movement – it extends like a huge tail of a crag-and-tail deposit from the mountain. Morphologically it is separated from the scree slopes along the mountain sides.

The extension of the crag-and-tail indicates NNE ice movement. A lateral moraine along the western side of the mountain extends in the same

Fig. 6. Buellia lichens close to the ice margin at Scullin Monolith.

Fig. 7. Lee-side deposit extended towards the NNE at the northern end of Central Masson Range in the Framnes Mountains.

direction. The moraine contains some erratics and shows collapse forms typical of a moraine, but the bulk of the moraine material is derived by mass movement. However, the position of the ridge indicates a former ice surface at least 50 m higher than today.

A similar ridge on the western side of Rumdoodle Peak, 20 km south of Mawson Station, provides a useful comparison. In this case there is no conclusive evidence of a true moraine accumulation. The material seems to be accumulated only by slumping from the mountain side, where intense frost shattering occurs. However, even this deposit indicates by its position and shape that the ice surface must have been higher at an earlier stage, although not necessarily more than a few tens of metres higher.

Moraines reaching higher than the present ice surface were recorded also at other places. On the northern side of Mt. Burnett, 30 km south of Mawson Station, a moraine ridge reaches a level which corresponds to a higher ice surface. On the southern, upglacier side of the same mountain, erratics and till indicate an ice surface higher than the present. The same applies to Mt. Twintop in the southern nunatak area. In all these instances there is evidence of a higher ice surface, but the level need not have been much more than a few tens of metres higher.

The best evidence of a higher ice surface is found on the coast at Mawson Station. This Australian scientific station is situated in an ice-free area about 1 km^2 in extension at the ice margin. The rock reaches only a few tens of metres above the adjacent ice on the upglacier side, but this obstacle to the ice flow is sufficient to account for some up-transport of debris to the ice surface. This is in contrast to the ice on the sides of the ice-free area, where the more continuous downslope flow results in an almost complete absence of debris in the ice. By the compressive flow against the rock, debris has been transported all over the ice-free area. There are only insignificant patches of till, but numerous erratics, often with good striae, and striated bedrock clearly show that ice has overridden the whole area. Like in the abovementioned examples the ice was not necessarily more than 50 m thicker than today. However, a greater thickness seems likely, one evidence being the consistent direction of the striae.

Several islands occur just off the coast at Mawson. They are completely barren, with no visible erratics or till. This may be due to a total clearing of the surface by glacial erosion in a generally downslope position. However, the evidence is not conclusive. Weathering is strong, and intense katabatic winds prevailing in the region tend to remove all weathered residual. A possible till cover or erratics may have been removed in this way. Also wave abrasion during former, relatively higher sea-level positions may be responsible for this cleaning.

From the observations mentioned it may be inferred that the ice sheet was more than 200 m thicker than today in the Mawson area. Admittedly, the recorded glacial features may reflect different stages, but this does not make any difference with regard to the maximum ice extension. The dating of this extension is difficult. Among the observations made there is very little to give information in this respect, although lichens (*Buellia* sp.) occur all the way down to the ice margin (cf. above, Scullin Monolith).

One possibility of dating might be the numerous lakes occurring at the nunataks. The lakes are situated in deep wind-channels, probably also kept open by melting due to reflected radiation from the bedrock. They are now entirely frozen. According to Ahlmann (1944), they represent a thermal optimum, corresponding to the Holocene hypsithermal stage (about 8,000 to 6,000 B.P.). This idea may be seriously questioned. Small meltwater streams seem to feed some of the lakes. The lakes were possibly formed successively by annual freezing of meltwater. An indication of this is the occurrence of numerous meltwater lakes (frozen) upon glaciers draining the ice cap. This applies to Forbes Glacier in the Mawson area, to Sørsdal Glacier at Vestfold Hills and many others. On the other hand, pingo-like updomings on some lakes indicate that the lakes became frozen as large water bodies, not sheet by sheet. Isotope studies of the lakes may offer an answer.

Thus, if Ahlmann's (1944) view is accepted, the higher ice surface must predate the Holocene climatic optimum and possibly correspond to the glacial maximum. However, this idea must so far be considered speculative in more than one respect and the question about the age of the greater extension of the ice remains open. A much older maximum stage cannot be excluded.

Conclusions

The observations described above are much too fragmentary to provide detailed conclusions. However, they all suggest a wider extension of the ice sheet in the entire sector 60–80°E. In the coastal region this implies a thickness at least 200 m greater than today. Nothing contradicts an assumption of a greater thickness. On the contrary, if we assume that features like the moraines and glacial sculpture described above require a warm-based ice, an ice thickness of at least 1,000 m can be inferred. We cannot exclude the possibility of some of these features being very old, formed by an early wet-based ice and then preserved beneath cold-based ice.

The dating of the maximum extension is therefore difficult. It may not have been simultaneous in all areas, and thus reflects, mainly, changes of balance within the ice sheet. On the other hand, features of weathering etc. give no support of such an assumption. They are similar in all the visited areas, and give the impression that there is no considerable time difference. Especially if we consider the more precise datings in better known areas (cf. Denton & Hughes 1981) and the generally good agreement between them and the recorded global climatic changes, e.g. in the deep-sea sediments, it seems more reasonable to assume a simultaneous maximum extension all over the area. This may have taken place during the glacial maximum, 18–20,000 B.P., recorded in other parts of the world, although a much greater age is possible (cf. Denton & Armstrong 1968; Calkin & Nichols 1972; Denton et al. 1984; Höfle 1987).

The discussed ice thickness can probably not be extrapolated all across the Antarctic ice sheet. Its profile and marginal gradient may have changed, and the increase in thickness must have been greater at the coast than inland (Hollin 1962, Fig. 4). The fact that the discussed area flanks the Amery Ice Shelf may result in anomalous surface and basal gradients. Also the relations between the West and East Antarctic ice may have changed. An unchanged ice profile with the minimum increase discussed above would give an increase of ice volume by $2.8 \times 10^6\,km^3$ during the glacial maximum.

However, if we assume that the glacial features described above require wet-based ice, the figure 2.8 must be increased. Basal temperature above the pressure-melting point at the highest level of the lee-side deposit in Central Masson Range (Fig. 7) requires, under prevailing surface temperature conditions (cf. Budd et al. 1971), an ice thickness of at least 1,000 m. This is in agreement with Hollin's (1962) calculation of the effect of a sea-level lowering by 150 m. According to Hollin's (1962, Fig. 4) profile, this figure cannot be applied all over the ice sheet. The applied profile gives an increase of volume between 2.5 and $8.5 \times 10^6\,km^3$, most probably in the lower part of this range. This is less than the value obtained in Denton & Hughes' (1981) theoretical model, $11 \times 10^6\,km^3$. Altogether, the calculations support the assumption of the observed features as derived from an early stage with warm-based ice. This stage cannot yet be dated, but in general, conditions agree with what can be expected during the late Wisconsin-Weichselian glacial maximum 18–20,000 B.P., although similar conditions certainly prevailed repeatedly also earlier.

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