PHYSIOGRAPHY OF THE FLASK GLACIER–JOERG PENINSULA AREA, GRAHAM LAND

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ABSTRACT. The geomorphology and glacierization of part of the Oscar II, Foyn and Bowman Coasts of Graham Land are described, and the contrasts between the north and south of the area are emphasized. The differences in the size and distribution of cirque glaciers established in north- and south-facing landforms are discussed. Although glacial and periglacial deposits are rare, variations in the elevations of planed surfaces and cirque floors are used to outline tentatively the geomorphological history of this area.

THE Oscar II, Foyn and Bowman Coasts of Graham Land are bounded by the Larsen Ice Shelf (Mason, 1950*b*; Kennett, 1965, 1966; Renner, 1969), which is 60 miles (97 km.) wide east of Flask Glacier and is at its widest (about 150 miles (240 km.)) to the east of Joerg Peninsula (Fig. 1). Although access to this area by sea is generally difficult because of pack ice in the Weddell Sea, the northern part of the area was first sighted and described by Larsen (1894), who sailed along the edge of the Larsen Ice Shelf as far south as lat. 68°10′S. In 1902 members of the Swedish South Polar Expedition travelled over the Larsen Ice Shelf as far south as Borchgrevink Nunatak (Nordenskjöld and Andersson, 1905) and the southern part of the area was first seen from the air in 1928 by Wilkins whose observations suggested that Graham Land comprised "two large islands separated from lands further to the south by an ice-filled channel 40–50 miles [64–80 km.] broad" (Wordie, 1929). The continuity of the western coastline of the Antarctic Peninsula was established by the British Graham Land Expedition (Stephenson, 1940) and the area was first mapped during a Falkland Islands Dependencies Survey traverse southwards across the Larsen Ice Shelf in 1947–48 (Mason, 1950*a*).

Geological mapping in the field seasons 1963–64 and 1964–65 involved extensive travel in this area and during this time physiographic observations were made. Graham Land is dominated by the gently undulating, snow-covered central plateau which rises from about 3,500 ft. (1,066 m.) at its eastern margin to a maximum height of 7,640 ft. (2,330 m.) at Slessor Peak. The hinterland east of the plateau is 25 miles (40 km.) wide north of Leppard Glacier but it is generally less than 5 miles (8 km.) wide south of Cape Robinson (Fig. 1).

PHYSIOGRAPHY

In the Flask Glacier–Churchill Peninsula area, rock exposures, though small, are widely distributed. Rock cliffs form much of the margin of Cabinet Inlet, which appears to have been sculptured by an earlier enlarged south-flowing ice stream and its feeder glaciers. From Cape Robinson to Three Slice Nunatak the coast is dominated by the plateau edge which has been indented by the glaciers flowing into Mill, Whirlwind, Seligman and Trail Inlets. Of the major eastward extensions of the coast, Joerg Peninsula (about 4,000 ft. (1,220 m.) high) is an extension of the plateau, while Jason and Churchill Peninsulas are distinctly lower ice piedmonts only about 1,000 ft. (305 m.) in height.

Flask Glacier–Churchill Peninsula area

Between Flask and Leppard Glaciers, a static crevasse-free ice piedmont slopes gently eastwards from a level of about 1,500 ft. (457 m.). Approximately 13 miles (21 km.) inland the gradient of this ice piedmont increases in height to form a narrow 3,000 ft. (914 m.) "sub-plateau" which in turn rises abruptly to the 5,000 ft. (1,524 m.) level of the Bruce Plateau.

The lower ice piedmont is broken by long ridges trending eastwards. The wider ridges grade into mountainous areas with flat summits at about 3,000 ft. (914 m.), e.g. Mount Alibi (3,040 ft.; 926 m.) and Target Hill (3,310 ft.; 1,008 m.). Bildad Peak (3,250 ft.; 990 m.) also has a planed snow-covered summit but the nearby Peleg Peak (3,020 ft.; 920 m.) is a pyramid resulting from cirque erosion of the same surface (Fig. 2). Although small rock exposures

Fig. 1. Physiographical sketch map of the Flask Glacier–Joerg Peninsula area, Graham Land. The contour interval is 1,500 ft. (457 m.) and the exposed rock on the east coast is shaded black.

Fig. 2. Typical inland scenery in the coastal area between Flask and Leppard Glaciers viewed from the south-west. The ice piedmont in the foreground is at about 1,500 ft. (457 m.). Bildad Peak (A; 3,250 ft.; 990 m.) has a flat ice-capped summit while Peleg Peak (B; 3,020 ft; 920 m.) has a prominent horn shape. There is an extensive area of patterned ground on the hill left of centre. (Photograph by A. J. Schärer.)

surrounded by large windscoops are associated with the rise to the Bruce Plateau, most of the area farther inland is snow-covered, and Moider Peak (3,820 ft.; 1,164 m.) is an eastward extension of the Bruce Plateau.

Leppard Glacier descends steeply from the 5,000 ft. (1,524 m.) level of the Bruce Plateau to about 2,500 ft. (762 m.) but, except for the lower 500 ft. (152 m.) where the gradient increases towards its wide, heavily pressured snout, it has a steady gradient for 25 miles (40 km.) and a uniform width of over 4 miles (6·4 km.).

South-west of Target Hill, the 3,000 ft. (914 m.) planed surface, which is more extensive here than north of Leppard Glacier, appears to be static, but it was probably once part of the glacier that occupied Richthofen Pass. Between Leppard Glacier and Adie Inlet, the rock outcrops are widely scattered and the lower ice piedmont is relieved only by north-westerly trending nunataks ranging from 1,530 ft. (466 m.) to 2,260 ft. (689 m.) in height (Fig. 3). Typically, these nunataks have narrow apical ridges but nivation by cirques on the higher south-east faces has produced spurs.

Jason Peninsula, which is 40 miles (64 km.) long, is a low gently undulating ice piedmont 500–1,500 ft. (152–457 m.) high with its highest point at the eastern end. There are a few, low rock cliffs near its junction with the Larsen Ice Shelf but the boundary between the ice piedmont and the ice shelf is generally not clearly defined.

The "root" of the northern end of Churchill Peninsula is formed by a planed surface about 3,000 ft. (914 m.) high which extends around the head of Adie Inlet and descends to 1,000 ft. (305 m.) to the north-east. Nivation, on a smaller scale than that which has formed and is eroding the bold scarp faces south of Cape Robinson, is at present dissecting the rim of Adie Inlet into rock ridges, spurs and cliffs. North-west of Adie Inlet and east of Eden Glacier is a line of rock bluffs which is associated with the step-like rise in the snow surface to another planed level at 3,000 ft. (914 m.).

Churchill Peninsula–Cape Robinson area

The eastern margin of Cabinet Inlet is formed by Churchill Peninsula, which is largely snow-covered and is terminated by the 1,000 ft. (305 m.) rock cliffs of Cape Alexander. In the north of the peninsula, a series of narrow ridges with rock emerging at their apices forms the foothills of the 3,000 ft. (914 m.) "sub-plateau". Although these ridges are only beginning to emerge from beneath the snow cover, they already have a lenticular shape, suggesting that they were formed at a time when the ice level was lower than at present. The western side of Lyttelton Ridge, having once been the margin of a much larger glacier in Cabinet Inlet, is mainly snow-free. Eden Glacier, which flows from north to south, is atypical of the glaciers

Fig. 3. The coastal area north-west of Adie Inlet viewed from the south-east. From the 3,000 ft. (914 m.) plateau area, where McCarroll Peak (C; 3,620 ft.; 1,103 m.) is the highest point, there is a steep slope to an extensive ice piedmont at about 1,000 ft. (305 m.) which slopes gently to the Larsen Ice Shelf (foreground). The nunatak left of Gulliver Nunatak (D) has a characteristic narrow apical ridge. (Photograph by A. J. Schärer.)

in Cabinet Inlet. Its upper part is relatively steep but crevasse-free, while its lower 6 miles (9·7 km.) are almost level. The coastal cliffs west of Eden Glacier have been dissected by cirques, giving rise to spurs up to 3,000 ft. (914 m.) in height. Evidence for a formerly larger glacier in Cabinet Inlet is provided by the truncation of the extremities of these spurs which have steep cliffs about 1,000 ft. (305 m.) high.

The greater part of the ice in Cabinet Inlet is derived from four heavily crevassed major glaciers (Attlee, Bevin, Anderson and Sleipnir Glaciers; Figs. 4 and 5), all of which flow eastwards from the Bruce Plateau and are cutting back by headwall erosion. In western Cabinet Inlet, the easterly trending ridges, e.g. Balder Point (Fig. 4), which are snow-free on their north sides, represent the margins of glaciers which have retreated. Cape Casey has a narrow, 2,000 ft. (610 m.) central ridge which has been abruptly truncated at its eastern end by the higher ice that occupied Cabinet Inlet. The largely snow-covered minor ridges on the southern sides of this cape terminate at about 600 ft. (183 m.), probably the maximum level reached by Anderson Glacier. A small nunatak 2 miles (3·2 km.) south of Cape Casey has a crag-and-tail form with a steep western face and a flatter boulder-strewn eastern side, indicating sculpture by this formerly enlarged glacier.

Stanley Island (Fig. 4) has steep southern and eastern coasts which are evidently ice-sculptured, while the northern end slopes smoothly to the level of Cabinet Inlet. In north to south profile, Stanley Island is a stoss-and-lee feature (Flint, 1957), because there is no evidence

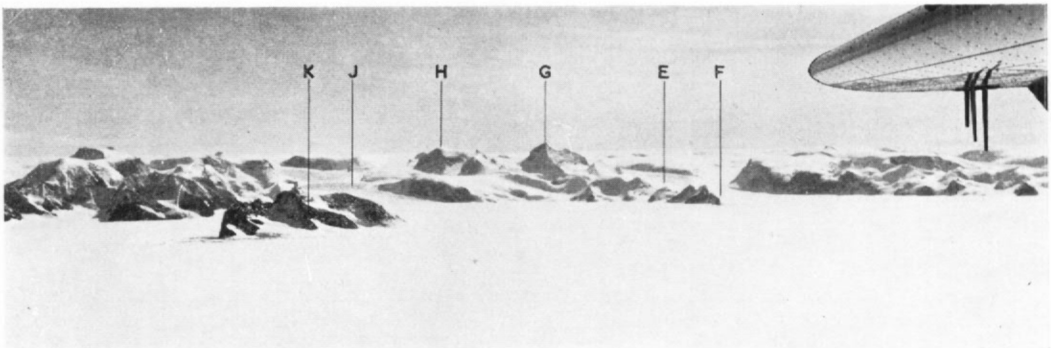


Fig. 4. Air photograph of part of western Cabinet Inlet from northern Anderson Glacier to the north of Spur Point viewed from the south-east. Between Anderson Glacier (E) and Sleipnir Glacier (J), which has a distinctive stepped profile, there is a narrow cirque-dissected ridge terminating at Balder Point (F). Frigga Peak (G; 5,150 ft.; 1,570 m.) has a prominent horn shape in contrast to the domed Mount Odin (H; 4,800 ft.; 1,463 m.). The profile of Stanley Island (K) emphasizes the steep southern bluff and the more gently sloping northern termination. (Photograph by E. Thornton.)

that there was ice movement from south to north in Cabinet Inlet and it is clear from the crevasse pattern and surface undulations that the present direction of ice transport is from north to south.

The plateau edge around the northern and eastern walls of Cabinet Inlet is formed by high mountains which are apparently the eroded remnants of the Bruce Plateau. Bastion Peak (5,290 ft.; 1,612 m.) and Mount Odin (4,800 ft.; 1,463 m.) have flat tops, whereas Frigga Peak (5,150 ft.; 1,570 m.) has been further eroded and is conical (Figs. 4 and 5). The contrast between Mount Odin and Frigga Peak is similar to the one between the ice-capped Bildad Peak and Peleg Peak (Fig. 2), suggesting that essentially similar forms of glacial erosion have occurred although the difference in elevation between these two groups of features is 2,000 ft. (610 m.).

In southern Cabinet Inlet, prominent ice-capped mountains, e.g. Mount Hulth (4,830 ft.; 1,472 m.) and Mount Denucé (5,030 ft.; 1,533 m.; Fig. 6), with steep, cirque-eroded eastern

Fig. 5. View from north-east of Anderson Glacier (E) of Frigga Peak (G; 5,150 ft.; 1,570 m.) and Mount Odin (H; 4,800 ft.; 1,463 m.), which display snow-covered benches of probable glacial origin at about 2,000 ft. (610 m.).

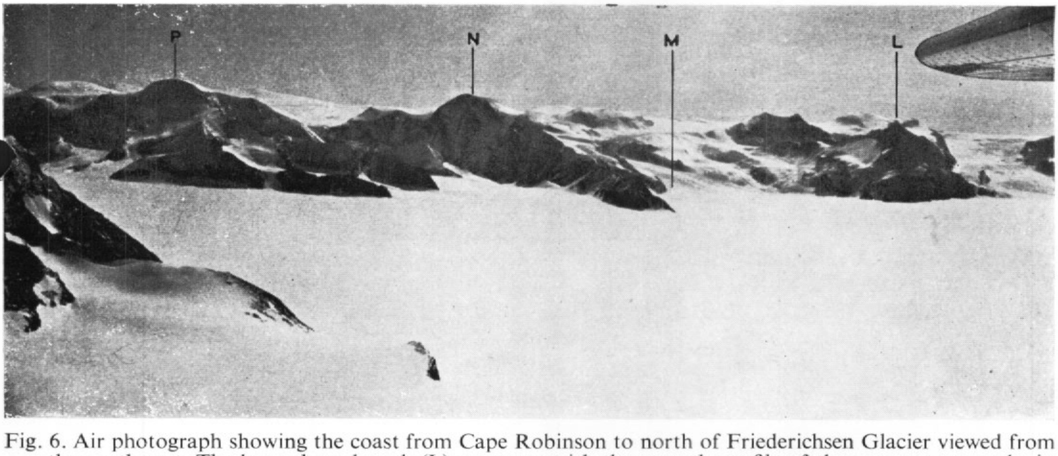


Fig. 6. Air photograph showing the coast from Cape Robinson to north of Friederichsen Glacier viewed from the south-east. The horn-shaped peak (L) contrasts with the smooth profile of the remnant nunataks in the middle of Friederichsen Glacier (M) which have been sculptured by valley glaciation rather than by cirque nivation. Long low ridges extend northward from Mount Hulth (N; 4,830 ft.; 1,472 m.) and Mount Denucé (P; 5,030 ft.; 1,533 m.). (Photograph by E. Thornton.)

faces descending to the Larsen Ice Shelf, form the eastern margin of the Avery Plateau. Spur Point, with a highest peak of 3,820 ft. (1,164 m.), and Mount Hayes (3,740 ft.; 1,140 m.; Fig. 7) have flat tops and are evidently eastward extensions of the Bruce and Avery Plateaux. Although it is unlikely that it was submerged by the former higher ice level in Cabinet Inlet, Spur Point has a crag-and-tail shape. West of Mount Hayes, a narrow cirque-dissected saddle rises without any sharp break to the level of the Avery Plateau and the ice-capped mountains. The contrast between the long narrow ridges developed on the north-facing side of Mount Hayes (Fig. 6) and the broad abrupt ridges to the south (Fig. 7) is striking (p. 67). The two isolated ridges east of Mount Hayes were probably once part of the same rock mass which extended from the Avery Plateau.

The ice piedmont of Cape Robinson, which rises to 1,250 ft. (381 m.) above the Larsen Ice Shelf, undoubtedly owes its preservation to its position between the two major ice masses in Cabinet and Mill Inlets. The present sub-radial ridge development on Mount Hayes is probably the product of a small local ice cap which exercised a protective role during maximum glaciation and has since been replaced by erosive cirque glaciers, which have reduced the massif forming Cape Robinson to the small relic of Mount Hayes.

Many of the glaciers flowing into Cabinet Inlet have a break in their profiles at about 2,000 ft. (610 m.), e.g. Sleipnir Glacier (Fig. 4) and Friederichsen Glacier (Fig. 6). This planed level is particularly marked in the profile of a relatively static glacier north-west of Cape Casey (Fig. 8), and it corresponds in height to the prominent snow-covered planed surfaces

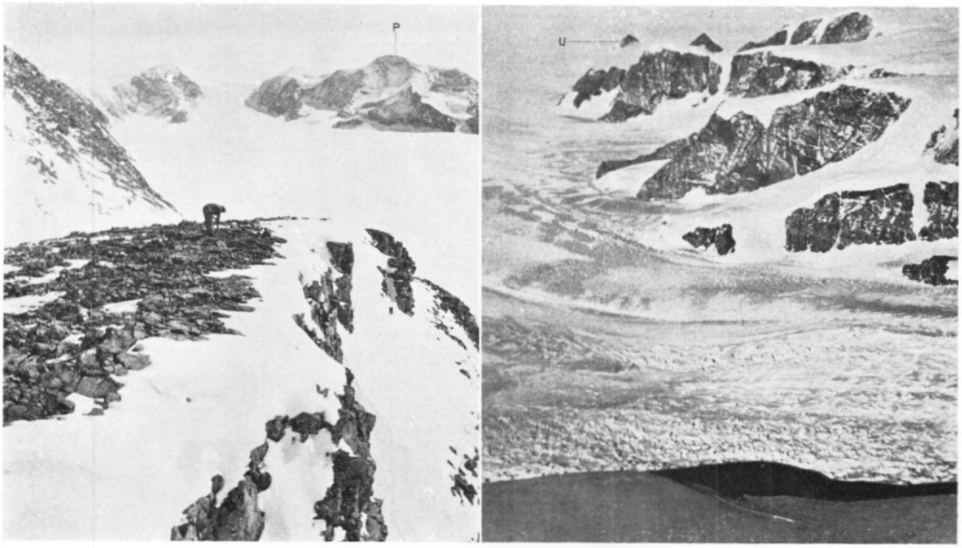
Fig. 7. Air photograph of Mount Hayes (R) from the south, showing its mesa-like pediment and massive south-facing ridges. (Photograph by E. Thornton.)

on the east face of Frigga Peak and Mount Odin (Fig. 5). On a small nunatak, 1,100 ft. (335 m.) high, forming the north-eastern rock exposure on Cape Robinson, there is a horizontal bench (Fig. 9a) which is level with the ice piedmont to the east but it has a vertical cliff 1,000 ft. (305 m.) high on the west. The upper surface of this bench is covered by a *felsenmeer* so that its origin is obscured. This surface was probably more extensive at one time, because south of Mount Hayes there is a small nunatak, 1,250 ft. (381 m.) high and similar in height to the bench described, near the boundary between the ice piedmont and Mill Inlet. The southern margin of this nunatak has been sharply truncated by the ice of an enlarged glacier that formerly occupied Mill Inlet.

Cape Robinson–Joerg Peninsula area

The coast between Mill and Trail Inlets is backed by a hinterland less than 5 miles (8 km.) in width. The major south-facing promontories, particularly Karpf Point (3,860 ft.; 1,176 m.; Fig. 10) and Cape Northrop (3,500 ft.; 1,066 m.), and the flat-topped extensions of the Avery Plateau form steep escarpments along the margin of the Larsen Ice Shelf. The narrow hinterland

Fig. 8. A glacier 7 miles (11·3 km.) west-north-west of Cape Casey viewed from the east. The 5,000 ft. (1,524 m.) planation of Bruce Plateau forms the horizon and there is a major break in the slope of the glacier at about 2,000 ft. (610 m.) while on the left of the photograph there is an extensive tract of level ground below 1,000 ft. (305 m.). The prominent peak on the right (2,980 ft.; 909 m.) has been eroded by an east-facing cirque.



a b

Fig. 9. a. The north-eastern extremity of Cape Robinson viewed from the south-east, showing the glacial bench at about 1,000 ft. (305 m.) which is covered by a *felsenmeer* to a depth of 14 in. (35 cm.). Mount Denucé (P) is in the background. (Photograph by E. Thornton.)

b. Air photograph of southern Whirlwind Inlet looking south-eastward from Chamberlin Glacier (foreground). The 3,500 ft. (1,066 m.) plateau escarpment has been strongly indented by cirque glaciation, and Tent Nunatak (U; 951 ft.; 290 m.) is a typical residual glacial pyramid. (Photograph by B. Hodges.)

is composed of low easterly trending ridges, which extend from the base of the plateau (Fig. 10), and rare, isolated, residual glacial pyramids, e.g. Tent Nunatak (951 ft.; 290 m.; Fig. 9b).

Steep, heavily crevassed valley glaciers have cut back the plateau escarpment to form the major inlets. Mill Inlet, which contains the most active glacier ice in this part of eastern Graham Land, is fed by at least five major glaciers (Aagaard, East Gould, East Balch, Breitfuss and Fricker Glaciers) draining either southward or eastward from the Avery Plateau. Because

Fig. 10. The coastline north of Karpf Point (S; 3,860 ft.; 1,176 m.) viewed from the east, showing the flat planation of the rectilinear edge of Avery Plateau at about 4,000 ft. (1,219 m.). A distinctly lower ridge (T) at about 1,500 ft. (457 m.) has been preserved between two major cirques.

these glaciers have cut back almost as far as the corresponding glaciers which flow westward into Darbel Bay, the Avery Plateau has been reduced to a narrow remnant which is locally only 5 miles (8 km.) wide. The boundary between the ice in Mill Inlet and the "static" ice piedmont of Cape Robinson is indicated by lenticular holes on the ice surface. At the junction between this same ice and the Larsen Ice Shelf to the south, there are similar crevasses, and rifting comparable with that described from south of Cape Disappointment (Fleet, 1965) occurs east of Monnier Point. Whirlwind Inlet, which is surrounded by the steepest cliffs in the whole of this area (Fig. 9b), is also fed by several radial glaciers. In general, the positions of the feeder glaciers in Mill and Whirlwind Inlets are comparable, i.e. Fricker and Chamberlin Glaciers (Fig. 9b) have similar orientations.

Francis Island lies between the major ice masses of Whirlwind and Seligman Inlets. There is a snow-covered col, about 500 ft. (152 m.) high, between the two main snow domes which are terminated at their southern ends by small conical nunataks. The higher western dome has subsidiary south-westerly ridges which descend from 2,320 ft. (708 m.) to about 900 ft. (274 m.) in the south, whereas the ridges on the east of Francis Island radiate from a snow dome only 1,150 ft. (351 m.) high. The exposed rock is mainly restricted to the terminations of the ridges and to a few terminal nunataks. The low, rounded snow-covered relief of Francis Island could be ascribed to modification of a pre-glacial topography by a local ice cap. It provides a striking contrast with the pyramid-shaped glacially sculptured features along the base of the plateau escarpment to the west.

The coastal topography in Seligman Inlet is less extreme than that of Whirlwind Inlet, and the preservation of Tonkin Island indicates that ice flow from the inlet could not have been great even at the maximum of glaciation. The ice from Lewis and Ahlmann Glaciers has truncated the northern and southern tips of Tonkin Island, which are formed by straight vertical cliffs with a crag-and-tail shape, descending from 1,500 ft. (457 m.) in the west to the level of the ice shelf in the east. Between these terminal bluffs is a snow-covered col to the east of which rock is exposed in a deep windscoop.

Trail Inlet is comparable with inlets to the north in the contrast between its northern and southern sides. To the north of the inlet is the cirque-dissected plateau escarpment, while Joerg Peninsula forms the southern boundary. Only two outlet glaciers enter Trail Inlet, but Three Slice Nunatak (1,650 ft.; 503 m.; Fig. 14) is a crag-and-tail feature orientated parallel to the flow of Daspit Glacier, indicating that it marks the southern margin of the earlier larger glacier in Trail Inlet.

The east-north-east trend of Joerg Peninsula is different from that of the other major eastern promontories (Jason Peninsula, Churchill Peninsula and Cape Robinson). There is no geological evidence for any structural discordance between the plateau and Joerg Peninsula, which slopes gently and regularly eastwards from 5,850 to 3,720 ft. (1,783 to 1,134 m.) over a distance of 21 miles (34 km.). A prominent feature is a narrow, steep-sided southerly pass, about 1,000 ft. (305 m.) high, which may occupy a fault. Pylon Point (2,848 ft.; 868 m.; Figs. 11 and 12) is a northerly extension of the pediment of Joerg Peninsula, although it has been almost severed from the peninsula by two cirque glaciers. Northern Joerg Peninsula is being attacked by the headwall erosion of cirque glaciers, resulting in the formation of long ridges and occasional isolated pyramid-shaped nunataks (Fig. 14). In contrast, the peaks on the snow-covered pediment of Joerg Peninsula are obtusely conical (Fig. 11) and they are identical in shape to those on the plateau, suggesting that they may also be pre-glacial in origin. Southern Joerg Peninsula comprises a few, major steep cirques separated by wide abruptly terminated ridges with rectilinear, near vertical, faces (Fig. 12). Rock is seldom exposed in the headwalls of these cirque glaciers which descend to the ice shelf in ice falls. The smaller cirques, which have established themselves on the major inter-cirque ridges, have steep head-walls and wide bergschrunds.

In the Trail Inlet area there appears to be a significant erosional level at about 1,500 ft. (457 m.). This is the height of the base of isolated cirques on Mount Shelby, the break in slope in Bills Gulch and Daspit Glacier, the highest point on Three Slice Nunatak and the summits of many of the long ridges and nunataks on northern Joerg Peninsula, e.g. east of Pylon Point. That erosion to at least two levels has occurred is suggested by one of these ridges (Fig. 11) which has a planar snow-covered surface at about 1,500 ft. (457 m.) on which stands a single, residual glacial pyramid.

Fig. 11. Joerg Peninsula eastward from Pylon Point (W) viewed from Three Slice Nunatak, showing the presence of many cirques between the ridges which are up to 2,500 ft. (762 m.) high. The morphology of the ridge (X) suggests that there were at least two levels of cirque erosion. The earlier, higher erosional level is represented by the small residual pyramid; a small bench occurs at similar levels in all of the cirques. (Photograph by E. Thornton.)

Glacierization

The major valley glaciers of eastern Graham Land are "outlet glaciers" draining the central plateau (Figs. 1, 4, 6 and 9b). True alpine glaciers have not been observed but cirques are common at various levels. Small apron glaciers are extensively developed above the south-facing cirques (Fig. 12). On north-facing features, some of the former cirques have been reduced to hanging glaciers. As in western Graham Land (Fleming, 1940), the lowland areas are covered by extensive flat or gently undulating ice piedmonts. Horizontal rock outcrops are very uncommon and they are always covered by *felsenmeere* (Fig. 9a) which obscure all direct evidence of glacial movement, e.g. glacial striae, etc. Englacial material has not been seen in any of the glaciers, although only a few isolated crevasses have been examined, and supraglacial moraines have been mapped in only three localities.

Fig. 12. Air photograph of a large snow-filled cirque on south-eastern Joerg Peninsula viewed from the south toward Pylon Point (W), showing the small apron glacier (Y) which has developed on the sidewall above the bergschrund. Apart from its steep rock cliffs, southern Joerg Peninsula is predominantly snow-covered. (Photograph by J. Ross.)

Outlet glaciers

The outlet glaciers are active sub-polar glaciers (Ahlmann, 1948, p. 66) with considerable surface melt in summer, particularly where they are exposed to the sun. None of these glaciers have been mapped in detail and only Leppard and Eden Glaciers and Bills Gulch have been completely traversed by dog-sledge; travel over the others would be impracticable. Because of the differences in the width of the hinterland, the northern glaciers are long and broad, with gently sloping profiles, whereas those in the south are steep and constricted.

Flask, Attlee, Friederichsen and Ahlmann Glaciers, and most of the other larger glaciers, are probably type I valley glaciers in Ahlmann's (1948) classification, but some of the glaciers with extensive firn basins and constricting sidewalls in their lower reaches (e.g. Bills Gulch) are type II glaciers. Leppard Glacier possesses lateral tributaries and therefore it belongs to Ahlmann's type IV, though it is essentially similar in profile to the commoner type I glaciers.

All of these glaciers have stepped profiles and the main variation between them is in the distance between the steps which seems to be dependent on the width of the hinterland. Leppard Glacier is steep at its head but it has a relatively smooth profile, having modified the shape of the stepped ice piedmont (p. 57). In contrast, the shorter Bills Gulch descends from a similar height (5,000 ft.; 1,524 m.) in a series of steep steps, the first of which, at the plateau edge, is relatively smooth and similar to the one at the head of Leppard Glacier. The main break in slope is sufficiently steep to form an ice fall which descends from about 1,500 ft. (457 m.) to the smoother tongue of the glacier (*c.* 750 ft.; 229 m.) in about 1 mile (1·6 km.).

The crevasse patterns on the surfaces of these glaciers reflect both the shapes of the sidewalls and the profiles of their beds. These patterns are similar in the upper parts and at the snouts of most of the glaciers, the crevasses commonly being parallel and perpendicular to the direction of ice movement, whereas in the central parts the crevasse patterns vary considerably. In the steeper glaciers (e.g. Daspit Glacier) the tumbled blocks of ice in the ice falls are delineated by transverse lenticular crevasses, the product of longitudinal tensile stress, indicating that the glacier has a convex bed (Nye, 1952), i.e. a riegel and basin profile (Flint, 1957). A fanned crevasse pattern, as displayed on the surfaces of Friederichsen Glacier, is probably produced by constriction of the glacier by its sidewalls (longitudinal compressive stress).

Cirque glaciers

Cirque glaciers have developed at several levels but a number of cirques commonly occur at the same level. On Mount Denucé (Fig. 13) they have formed at three distinct levels, whereas on Mount Shelby they all occur at the same level as the main break in slope of Bills Gulch and Daspit Glacier (*c.* 1,500 ft.; 457 m.). Two distinct types of cirque glacier have been recognized:

i Almost ice-free cirques which have extensive, predominantly snow-free headwalls and have formed on north-facing landforms, e.g. northern Cape Casey, Mount Hayes (Fig. 7) and northern Joerg Peninsula (Fig. 14).

ii. Snow-filled cirques which are restricted to south-facing landforms. They are generally broader and larger with snow-covered headwalls which often have small apron glaciers (Fig. 12).

The ridges between the almost ice-free cirques are long, low and narrow, whereas the inter-cirque ridges between the snow-filled cirques are correspondingly massive, and small subsidiary cirques have been established on them. This contrast in ridge morphology is clearly displayed on

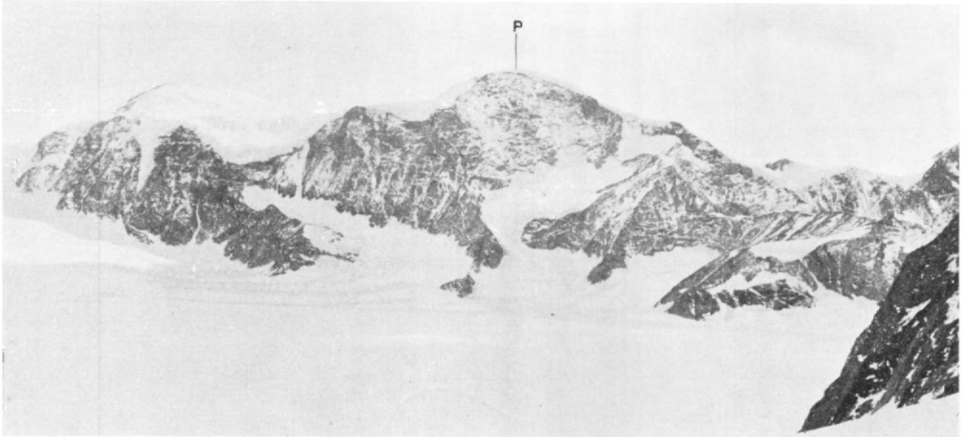


Fig. 13. Mount Denucé (P) from the east, showing cirques established at three levels probably reflecting fluctuations in the elevation of the firn line. The uppermost is a hanging cirque. (Photograph by E. Thornton.)

Mount Hayes (Figs. 6 and 7) where the ridges on its north side have usually been reduced to low saddles connecting the mountain to small isolated terminal "monuments". In contrast, the ridges extending from southern Mount Hayes have been considerably less eroded and there is no truncation by cirques (Fig. 7).

Cirque glaciers are apparently enlarged by headwall shattering and basal scouring. Battle (1960) has shown that maximum frost action, leading to shattering of the headwall, occurs when most of the headwall is exposed to diurnal freezing and thawing. This condition is satisfied in the ice-free cirques, and considerable volumes of melt water have been observed on north-facing rock exposures in this area in summer. Fisher (1955) has shown that the freeze-thaw process envisaged by Johnson (1899) can operate beneath the ice surface of "cold" glaciers but this process is governed by fluctuations in the geothermal balance of the Earth and not by changes in the ambient temperatures. In the snow-filled cirques, basal scouring may be a more important factor than headwall erosion (Thompson and Bonnlander, 1956), though McCall (1960) has shown that in temperate glaciers an increase in ice thickness beyond

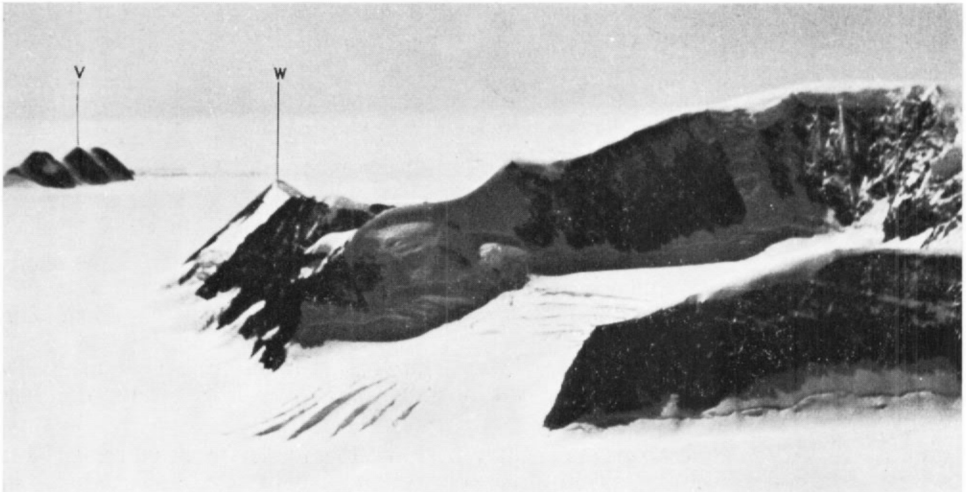


Fig. 14. Air photograph of Three Slice Nunatak (V) and a cirque south-west of Pylon Point (W) viewed from the south-west. The cirque has extensive, long low sidewalls and a narrow headwall. The measurement of major vertical joints on the terminal rock ridges shows that their strike coincides with the orientation of the headwall and sidewalls of the cirque. A small apron glacier (right foreground) has formed on the rock cliff above a prominent bergschrund. (Photograph by J. Ross.)

72 ft. (22 m.) will not increase the amount of basal abrasion. No direct evidence indicating the process of erosion in south-facing cirques has been found because englacial material is apparently absent from them.

It appears that erosion is headward in north-facing cirques, because the inter-cirque ridges are long and narrow. Thus the shape of the cirques, i.e. the ratio of width to length, seems to be dependent only on whether the feature faces north or south. It is clearly independent of the bedrock lithology, because the relatively ice-free cirques have formed on north-facing features in plutonic, volcanic, sedimentary and metamorphic rocks. Snow-filled cirques have only been observed in plutonic and volcanic rocks but this is probably a consequence of the dominance of these rock types in this area.

However, lithology is a factor limiting the size to which a cirque will increase (Embleton and King, 1968), because the competence of a rock to withstand failure will limit both the height of the cirque and the inter-cirque walls. This could account for the low relief of the north-facing ridges of Mount Hayes, which are composed of close-jointed volcanic rocks. The "monuments" at the termini of these ridges, which commonly form separate nunataks, are composed of more competent plutonic rocks.

From field observations, the orientation of cirque walls appears to be largely independent of lithology, which is in accordance with the views of workers in other areas (Helland, 1877; Harker, 1901; Lewis, 1938; Battey, 1960). In this area the cirques cut across lithological boundaries, igneous contacts, faults, bedding and foliation, but there appears to be a strong coincidence between the strike of the major vertical joint planes and the sidewalls and headwalls of the cirque glaciers. This suggests that joint-block removal (Lewis, 1954) may be the most important single factor controlling cirque orientation. The joints are believed to be largely tectonic in origin but they have probably been opened up by pressure release and augmented by dilation jointing (Battey, 1960). Joints facilitate both headwall erosion by freeze-thaw processes and spalling, and they also provide material which can be used by the glacier to shape its walls (McCall, 1960; Glen and Lewis, 1961).

Qualitative analyses (e.g. Lewis, 1938; Schytt, 1959; Temple, 1965) on the orientation of cirques in other areas indicate that they tend to form on the lee sides of landforms, commonly on the side shaded from the sun. However, on Cape Robinson and Joerg Peninsula there are more cirques on the north-facing features where solar radiation is greatest. Under the present climatic conditions, snow is deposited on the south-facing sides of the ridges and it tends to be ablated by wind and sun from the northern sides. It therefore appears that although the cirques formed on south-facing features are fewer in number they are larger in size because of the higher precipitation. It is very probable that on the more heavily glacierized southern sides of ridges cirque nivation has reached a more mature stage and that the large smooth cirques represent the coalescence of several smaller ones (Linton, 1963*a*).

Glacial deposits

Below 500 ft. (152 m.) on the south side of Leppard Glacier there is a locally derived lateral moraine, about 100 ft. (30·5 m.) wide and 800 ft. (244 m.) long (Fig. 15), composed of frost-shattered boulders and occasional blocks up to 5 ft. (1·5 m.) in diameter. A terminal moraine, composed of rounded boulders of granite and gneiss (up to 10 in. (25 cm.) in diameter), extends for about 5 miles (8 km.) from the snout of Friederichsen Glacier. The probable present limit of glacier flow is marked by several large angular boulders (over 10 ft. (3 m.) in diameter), which lie on the ice in a line parallel to the moraine and about 650 ft. (198 m.) from it.

Along the southern margin of Adie Inlet, a double-ridge moraine fingers north-eastwards for about 2 miles (3·2 km.) across the surface of the Larsen Ice Shelf. Though this moraine may be broken in local accumulation areas, it does not vary in width, and therefore, it is probably a shear moraine (Schytt, 1956) marking an earlier boundary between a small glacier and the ice shelf in Adie Inlet. Schytt (1961, p. 190) has also described the occurrence of shear moraines in local ablation areas in Dronning Maud Land; this moraine in Adie Inlet is remarkably similar to arcuate "necklace" moraines described from some of the ablation areas below small valley glaciers in the Shackleton Range which Stephenson (1966, p. 16) considered are the surface expressions of thrusts in the ice. Also, the "railroad moraine" on Boomerang Glacier, Victoria Land (Wright and Priestley, 1922, pl. 138), appears to be similar to this moraine in Adie Inlet.

These are the only moraines that have been recorded in this area. The fact that all the glaciers have weathered rock walls and lichens are established close to the rock/ice junctions indicates that this area is at present almost in glacial equilibrium. Koerner (1964) found a similar equilibrium state in north-east Graham Land and this is probably true for most of Antarctica

Fig. 15. The locally derived lateral moraine on the south side of Leppard Glacier. Jason Peninsula is just visible in the background. (Photograph by A. J. Schärer.)

(Schytt, 1953; Gunn and Warren, 1962). In summer, blue ice, with scalloped surfaces indicative of wind erosion, occurs extensively in the lower reaches of most of the glaciers, and also in southern Adie Inlet and at the bottom of many windscoops around isolated rock outcrops. As in the Sør-Rondane, there are no intermediate transitional layers at the sharp boundary between snow and blue ice, which Van Autenboer (1964, p. 50) has suggested might indicate a relatively recent rapid change in the rate of accumulation.

Geomorphological History

The dominant feature of eastern Graham Land is the plateau and its escarpment. Holtedahl (1929), Nichols (1960) and Linton (1964) considered that the plateau is a pre-glacial peneplane eroded during the Middle or late Tertiary. Linton (1964, p. 95) has suggested that the ice surface follows the rock topography about 300 ft. (91 m.) beneath. Recent measurements by Dr. C. W. M. Swithinbank (Renner, 1969) indicate that the ice west of Joerg Peninsula is 1,310 ft. (399 m.) thick, which is considerably greater than estimates from elsewhere in Graham Land. The ice appears to have a protective role on the plateau and an extremely erosive one in the glaciers, because ice moving across the plateau loses only a small amount of its potential energy compared with ice descending about 3,000 ft. (914 m.) to the Larsen Ice Shelf (Linton, 1963*a*).

In western Graham Land, extensive block faulting has been proved (Hooper, 1962; King, 1964; Curtis, 1966), and this may explain the escarpment, but on the east coast there is no geological evidence for large-scale block faulting. The cliffs of Whirlwind Inlet may represent scarp faces, but Mount Hayes and Joerg Peninsula appear to be extensions of the plateau without obvious breaks. The Avery Plateau domes gently from its highest point, Slessor Peak (7,640 ft.; 2,329 m.), to Mount Hayes (3,740 ft.; 1,140 m.). This doming could be ascribed to an easterly tilt of Graham Land resulting from uplift of the west coast by block faulting.

In addition to the planation of the plateau, flat surfaces have been recognized at other levels in eastern Graham Land. There are extensive ice piedmonts at about 1,000 ft. (305 m.) on Jason and Churchill Peninsulas and in the area west of Cape Robinson, and there is a rather higher but more undulating ice piedmont on Francis Island. In their morphology they resemble the extensive low peneplanes of western Graham Land such as the Marr Ice Piedmont on Anvers Island (Holtedahl, 1929) and the Fuchs Ice Piedmont on Adelaide Island (Dewar, 1967). Holtedahl considered that they had been eroded by "strandflat glaciers" established below cirques which formed in a major fault scarp. Both Fleming (1940) and Dewar (1967) have drawn attention to the absence of subglacial morainic material between the ice and the rock and, as static ice has a mainly protective function, it is very unlikely that Jason Peninsula, which is 40 miles (64 km.) long with its highest peak in the east, could have been eroded in this way. These low planed surfaces, together with similar features in north-east Graham Land (Koerner, 1964), are probably pre-glacial in origin.

Other extensive planed surfaces have been described from the area west of Jason Peninsula (p. 57-59). North of Leppard Glacier distinct planed surfaces have been recognized at 1,500 ft. (457 m.) (the lower ice piedmont), 3,000 ft. (914 m.) (the coastal mountain summits) and 5,000 ft. (1,524 m.), the level of the Bruce Plateau. South of Leppard Glacier there is a gentle slope from the Bruce Plateau to a "sub-plateau" at 3,000 ft. (914 m.) which extends southwards to form the "root" of Churchill Peninsula. To the north-west of Adie Inlet distinct planed surfaces have been recognized at 1,500 and 3,000 ft. (457 and 914 m.) but on the north side of the inlet there is a gentle eastward slope from 3,000 to 1,500 ft. (914 to 457 m.).

It is probable that faulting and tilting took place intermittently during the Tertiary and that these planed surfaces were eroded in periods of stillstand and, in some cases, were subsequently tilted. They have only been revealed by the relative deglaciation of this part of eastern Graham Land. Because these planed surfaces are so extensive, they are more likely to be of fluvial origin. Although the present topography of Victoria Land appears to have resulted exclusively from glacial erosion (Priestley, 1923; Gunn and Warren, 1962), Nichols (1966, p. 7) has suggested that a long period of fluvial erosion preceded glaciation in the Antarctic Peninsula, and Gould (1940, p. 731) considered that in the Queen Maud Mountains faulting "began in Tertiary time long before the inception of glacial conditions". In the Sierra Nevada, erosion surfaces are also believed to have been the result of alternating Tertiary uplift and fluvial erosion (King, 1965).

It is probable that an eustatic drop in sea-level during earlier phases of the Pleistocene glaciation of Antarctica (Hollin, 1962) led to an increase in the extent and thickness of the Larsen Ice Shelf and increases in glacier thicknesses. Glacial benches, such as those of Victoria Land (Bull and others, 1964), may arise in this way and, with the absence of major horizontal sills in the Antarctic Peninsula, they are unlikely to be mistaken for structurally controlled planed surfaces (McGregor, 1963), although they could easily be confused with glacially eroded pre-glacial erosion surfaces.

The truncation of the marginal spurs and ridges surrounding Cabinet Inlet suggests that it was once filled to a much higher level by a major outlet glacier. Deep troughs cut by major outlet glaciers may be significant in the lowering of plateau levels and the starving of smaller outlet glaciers (Grindley, 1967). Ice "piracy" by Attlee and Bevin Glaciers could explain the considerable retreat of Anderson and Sleipnir Glaciers. The local effect of a major outlet glacier would also tend to obscure evidence for regional glacial fluctuations which have been recognized in Victoria Land (Grindley, 1967, p. 570).

The smaller planed surfaces at 2,000 ft. (610 m.) include the break in slope of the glaciers in Cabinet Inlet and the presence of ice-covered benches on the eastern slopes of Frigga Peak and Mount Odin. Whereas these planed surfaces may have resulted from an increase in the apparent thickness of ice in Cabinet Inlet, they could also be remnants of pre-glacial erosion surfaces. The peaks west of Cape Casey, also about 2,000 ft. (610 m.) high, were probably eroded from an extensive pre-glacial surface at about 3,000 ft. (914 m.).

Cirques tend to become established at or slightly above the snow line and therefore the existence of cirque glaciers at various levels on Mount Denucé (Fig. 13) suggests fluctuations in the glacial maximum. An apparent increase in the thickness of the ice level in Antarctica by about 984 ft. (300 m.) has been suggested by Rutford and others (1968), and Grindley (1967, p. 579) has discussed the possible causes of this.

In this part of eastern Graham Land, Nordenskjöld (Andersson, 1906) found erratic boulders at a height of 975 ft. (297 m.) above the ice shelf on Borchgrevink Nunatak, and they have been found at a slightly lower height on Stanley Island. The only glacial bench recorded in this area is at about 1,100 ft. (335 m.) on north-eastern Cape Robinson (Fig. 9a). If the ice were about 1,000 ft. (305 m.) thicker during a glacial stillstand than at present, cirque nivation would only be effective down to this level. The presence of ridges and isolated nunataks at this height provides some evidence for the elevation of the ice level by 1,000 ft. (305 m.), though it is also possible that these lower features were derived by erosion of a pre-glacial surface. Examples of features planed to this level include the nunatak south of Mount Hayes (1,240 ft.; 378 m.), the ridge north-east of Karpf Point (Fig. 10) and Tonkin Island, though its highest point reaches 1,700 ft. (518 m.). On Cape Robinson, two nunataks to the east of Mount Hayes have been separated from it by the truncation of an arête, at about 1,200 ft. (366 m.), between two large cirques. The height of the arête corresponds to the base level of erosion of the cirques. On eastern Joerg Peninsula there is a small cirque-eroded peak (Fig. 11) with its base at about 1,500 ft. (457 m.), which is also the height of many cirque floors in this area.

Ice may reduce or eradicate all vestiges of inter-cirque ridges and arêtes (Linton, 1963*a*). It is probable that the escarpment in Whirlwind Inlet shows this process in extremes because all traces of the hinterland or ridge connections with Francis Island have been removed. In all of the inlets the ridges between parallel glaciers have been eradicated or cut back to abrupt cliffs, e.g. Bastion Peak and Mount Shelby. Linton has suggested, however, that isolated nunataks (e.g. Three Slice Nunatak, Tonkin Island) will survive in a heavily cirque-glaciated environment because once they are isolated pressure release and dilation jointing (Battey, 1960) no longer provide planes of weakness which can be exploited by the agents of erosion.

Conclusions

Although it is possible that the ice was formerly at least 1,000 ft. (305 m.) thicker than at the present time, there is clearly insufficient evidence to postulate multiple glaciation in this area. It is suggested that the topography of this part of Graham Land is the result of pre-glacial peneplanation and uplift, upon which the effects of glacial erosion and cirque nivation have been superimposed.

Joerg Peninsula forms the boundary between the two physiographic provinces described by Mason (1950*b*, p. 410), who contrasted the area dominated by the plateau escarpment to the north of Three Slice Nunatak with the broken topography to the south. This change in the topography also coincides with the junction between the dextral and sinistral arcs of the Antarctic Peninsula. Also, at this latitude there is a sharp increase in the width of the Antarctic Peninsula, which Linton (1963*b*) has suggested is the result of differing degrees of deglaciation. There is only a narrow transitional zone between these two provinces and it is now suggested that the physiographic differences within the Flask Glacier–Joerg Peninsula area and between the areas to the north and south of Three Slice Nunatak have resulted mainly from variations in the degree of tectonism, due to the change in the form of the arc of the Antarctic Peninsula.

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