Fig. 1. Controlled sketch map of the central Black Coast, Palmer Land.

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PHYSIOGRAPHY AND GLACIAL GEOMORPHOLOGY OF THE CENTRAL BLACK COAST, PALMER LAND

By D. G. SINGLETON

ABSTRACT. The physiography and glacial geomorphology of the central Black Coast, Palmer Land, are described.

Weathering processes are discussed in relation to the widespread disintegration of the bedrock. A comprehensive account of differentially eroded forms is given and some evidence is put forward supporting theories as to the origin and development of cavernous weathering.

Some small-scale patterned ground features are recorded with suggested processes of formation in the Antarctic environment.

The occurrence of ubiquitous glacial erratics and three supraglacial moraines is noted.

THIS paper describes the physiography and glacial geomorphology of the central Black Coast, Palmer Land (lat. 70°50′–71°30′S., long. 64°–61°W.), from observations made during the two austral summers 1972–73 and 1973–74. The work was undertaken from the British Antarctic Survey scientific station at Stonington Island (lat. 68°11′S., long. 67°01′W.).

The central Black Coast (Fig. 1) is bounded on the west by the 2,000–3,000 m. plateau demarcated by the Welch Mountains and Mount Jackson, the highest point on the Antarctic Peninsula. The terrain extends eastward from this scarp for 35–45 km. to the Larsen Ice Shelf; the area is gently undulating and there are many nunataks with occasional steep slopes into the glaciers flowing towards the inlets.

This area was first visited by a sledge party of the United States Antarctic Service Expedition (U.S.A.S.), 1939–41, when geological samples were collected from the Welch Mountains which were mistakenly confused with the Eternity Range (Black, 1945; Knowles, 1945). In November 1940, this expedition pioneered an overland route from the west coast to the east coast of the Antarctic Peninsula and made a full reconnaissance along the Larsen Ice Shelf as far south as lat. 71°48′S. This ground work was further supplemented by air photographs and observations from the air; the photographs extended over the area discussed here and led to the first mapping of Mount Jackson.

The combined sledging operations of the Falkland Islands Dependencies Survey and the Americans during the Ronne Antarctic Research Expedition (R.A.R.E.), 1946–48, resulted in a further survey of the east coast which extended the previous observations southward to lat. 70°42′S. (Ronne, 1948). Valuable information was obtained from the trimetrogon air photograph cover taken during this period.

Since that time, the western fringes of this area have been visited by personnel of the Falkland Islands Dependencies Survey and the British Antarctic Survey during the systematic topographic and geological mapping of Palmer Land.

In October 1972, a geologist, an assistant and two dog teams reached the area overland to commence the reconnaissance geological mapping. In the following year a similar party was flown into the area and the geological mapping was extended.

PHYSIOGRAPHY

As is characteristic of most parts of eastern Palmer Land, the topography is dominated by the prominent peaks which mark the immediate boundary between the high plateau and the east coast. This is typified by the Welch Mountains and Mount Jackson whose summits provide a relief of 1,200–1,500 m. above the highest ice levels immediately to the east. The northern counterpart of these mountains is the Eternity Range. All of these peaks are interconnected by ice-fall zones and steep snow slopes at plateau level.

The Mount Jackson massif (3,178 m.) is the highest part of the Antarctic Peninsula and it consists of two rock masses. The northern half is dominated by Mount Van Buren with its flat-topped to sub-rounded summit (2,865 m.) from which extend two rock ridges. One of these forms a significant arête in an easterly direction and the other trends northward towards the Welch Mountains as part of the high plateau "drop-off" zone. Mount Jackson, which encompasses most of the southern half of the massif, has a classic cirque (Fig. 2) on its eastern side.

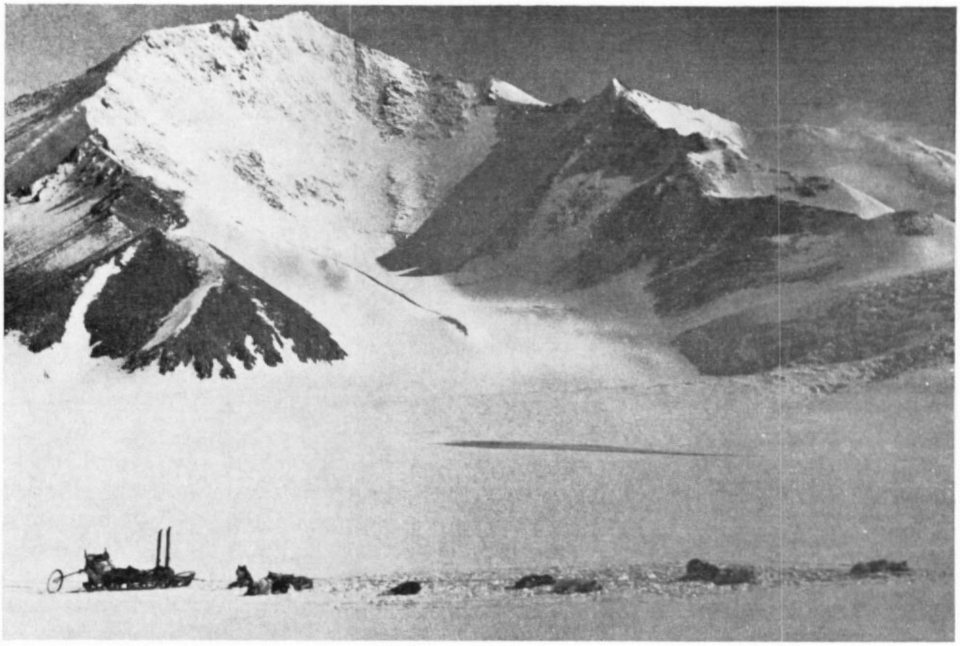

Fig. 2. The eastern side of Mount Jackson showing a large cirque.

The relief of the west-facing slopes is subdued and contrasts with the sheer rock walls (1,000 m. or more high) on the eastern sides. This asymmetry is also observed in the plateau-edge mountains farther north.

The Welch Mountains are a more composite massif consisting of two parallel north–south ridges which are separated by an east–west ridge whose lowest point forms a col. The western ridge, rising to Mount Acton (3,015 m.), extends northward along the plateau edge, a trend which continues through Liston Nunatak and Mount Schimansky. The eastern ridge, which is of similar proportions, is dominated by Mount Nordhill (2,927 m.). The northern section of this has several easterly trending arêtes which enclose cirques proximally.

The greater part of the area, from the plateau edge to the Larsen Ice Shelf, is a gently undulating crevasse-free snowfield descending from 1,500 to 1,100 m. Ill-defined glaciers cut this snow plain and they have characteristically steep rock exposures, screes and ice on their flanks. The nunataks along the margins of the snowfield are the surface outcrops of the subglacial ridges which hold back the ice. The larger mountain chains and nunataks that hold back the vast snow accumulation area of the high plateau are larger-scale examples of this feature. Where this ice has broken through or cascaded over the ridges, bergschrund crevasses and ice falls have developed on the "drop-off" side, feeding a lower level of snow movement towards the Larsen Ice Shelf. Murrish Glacier, for example, is channelled by a number of composite "drop-off" zones and is about 200–300 m. below the main snowfield level. These wide snow-valley features of ice movement have been referred to as glaciers because of their elongate channelled form, confined between rock masses on either side. Glacier movement appears to be particularly slow because crevasse systems are almost absent and are only extensive in the immediate inlet areas where many glaciers converge over a much steeper subglacial topography. Ice flow from this area is in three main directions: south-east towards Odom Inlet, east towards the Palmer Inlet–Morency Island area and north-east towards Lehrke Inlet.

Because the "drop-off" features are non-linear, they are difficult to explain structurally. Their combination with associated glaciers provides natural areas for the funnelling of wind, and this is substantiated by the number of blue-ice patches observed on the air photographs. These areas of blue ice often show surface rippling which is frequently broken by small cracks and cryoconite holes. Loud reports heard during the night while camping on blue ice in the Welch Mountains were probably caused by contraction of the ice due to falling temperatures. Blue-ice areas also occur at the base of windscoops around rock buttresses and nunataks, particularly near the plateau edge where maximum wind speeds are common. These areas probably resulted from the removal of snow by the roller vortex flow of wind caused by a nunatak or a similar obstruction. The snow is deposited at the vortex extremity and eventually, through nivation and constant drift, a windscoop is constructed with its maximum height opposite the rock wall which faces the prevailing wind.

The upper snowfield extends eastward to the coast, where "hogs-back" headlands terminating in sheer rock cliffs form the capes between the inlet areas. Lamplugh and Palmer Inlets are strikingly linear examples and, together with numerous others, they provide the main discharge outlets for the ice from the peninsula. They could possibly be interpreted structurally but are more likely to have resulted from ice retreat with subsequent headwall erosion (Marsh and Stubbs, 1969).

Glacial Geomorphology

Weathering processes on differing lithologies

The most striking geomorphological feature of this east coastal area is the extreme paucity of bedrock exposure, as a result of the intense shattering and general disintegration of the nunataks and mountains. The extent of this process is primarily dependent on the lithology and it is characterized by a close connection between loose blocks and the local bedrock.

General rock break-down is probably due to the combined effects of insolation and freeze–thaw. Rock is a poor conductor of heat so that a thermal gradient develops between the surface and the inner layers. Therefore the outer layer expands more than the interior and stresses are set up to produce a fracture. This differential thermal expansion is further enhanced within the surface layer by the individual response of mineral types to insolation. Different minerals have varying specific heats and coefficients of expansion, and the darker mineral species will absorb heat more quickly than the lighter ones. The extent of the resulting expansion would then depend on other physical properties of crystals as anisotropy, such that movement would take place in one direction preferentially to another. This process would clearly operate at its maximum during the summer when considerable diurnal and nocturnal temperature changes can occur. Mercer (1963) has stated that the surface temperature of diabase reaches at least 20° C under favourable conditions and even light-coloured rocks can reach temperatures above freezing point. Blackwelder (1933) has substantiated this process but he has placed it on a time-scale indicating that granular disintegration will occur through the accumulation of minute strains over centuries. He has further emphasized the importance of hydration in this process, but in Antarctica this would be a variable factor depending on the presence of snow and its consequent melting and freezing on the rock.

The nocturnal freezing of this moisture will produce a force additional to that of simple insolation. Water expands by about 9 per cent on freezing at 0° C (Ollier, 1969) and this would provide a considerable force in disaggregating rock. On exposed surfaces, wind would also assist the removal of debris loosened by the above processes. Wind abrasion is probably more significant than these processes during the autumn and spring months when wind speeds are highest. Further minor processes, which could assist granular disintegration are discussed under cavernous weathering.

It is not known to what extent chemical processes are active throughout this area, especially in view of the fresh appearance of gravels, particularly those in weathering pits. However, various workers (Mercer, 1963; Derbyshire, 1972; Boyer, 1975) have found evidence for chemical weathering in other areas of Antarctica.

Lithological varieties will respond to granular disintegration and allied processes according to their general structure, composition and texture. The response of the thermally metamorphosed rocks to weathering processes is variable; hornfelsic types develop into a coarse autochthonous felsenmeer or, on steeper slopes, coarse scree, whereas the more cleaved slates become highly comminuted through constant splitting along planes of weakness, ultimately producing a silty regolith at depth. The biotic factor is minimal in the disaggregation of rocks but some observations have been made of lichens mechanically lifting cleavage flakes. The surface of the resultant highly weathered slates is often covered by white encrustations. These have also been observed on some mafic gneissic and amphibolitic bands in the metamorphic complex rocks and they resemble a white opaque amorphous mass which on microscopic examination appears to enclose some translucent fibrous crystal areas. X-ray diffraction studies and a comparison with the work of Van Autenboer (1964) confirm that these encrustations are composed of gypsum. It is difficult to assess the degree to which the growth of these gypsum crystals has contributed to fracturing of the rock but, because of its widespread distribution, particularly within the contact-metamorphosed rocks, it cannot be overlooked as a possible minor agent.

Exfoliation occurs on all the exposed surfaces of the intrusive rocks. It takes the form of partings within the upper layers, parallel to the curvature of the rock surface. These partings are invariably discontinuous and hence "flaking" might be a better term. Each flake then seems to be subject to the process of granular disintegration through the mechanisms already discussed, and this results in a gravel end product with occasional residual boulders (Fig. 3).

Fig. 3. Surface flaking of a tonalite showing some residual boulders (E.4065). The hammer shaft is 60 cm. long.

Hamelin and Cook (1967) have attributed exfoliation to frost action, a process which has been further substantiated by Van Autenboer (1964) and Blackwelder (1925), who has differentiated between spalling and chemical exfoliation, the former being effected by ice.

The rocks of the metamorphic complex weather within the two limits set by granular and non-granular to slaty end products. Thus, a particularly granitic gneiss will yield to flaking and granular disintegration whereas a cleaved schistose rock will yield in the same manner as observed in similarly structured slates.

The volcanic rocks are so sheared and fractured that their ultimate disintegration into rubble is largely due to the above-mentioned processes, particularly frost action concentrated along the fractures.

Differential weathering

Differential weathering is largely dependent on structure, lithology and surface differences caused by earlier exogenetic processes such as glaciation, and this occurs on all scales. The large-scale processes give rise to mountains and glaciers, but it is the numerous smaller-scale features which are discussed here.

The more massive metamorphic and similarly structured igneous rocks show many forms of cavernous weathering. This term has usually been employed to denote processes by which recesses are developed (Dahl, 1966). These have been variously called gnammas (Twidale and Corbin, 1963), pits, niches, hollows, dew holes, honeycombs (Smith, 1941; Van Autenboer, 1964; Demek, 1964) and tafoni. However, "weathering pit" is the term used here for cavernous weathering forms on horizontal surfaces regardless of their depth and development. Complex sculpturing and recesses on vertical surfaces are grouped as "erosion cavities".

Weathering pits range in depth from shallow spall depressions to U-shaped hollows up to 30 cm. in depth and 40 cm. in width. The downward penetration of a pit may cut a vertical rock face so that with subsequent wind erosion a cavity will develop. Their shape in the horizontal plane varies from circular to irregular ovoid forms which have inevitably resulted from amalgamation of two or more pits. A honeycomb pattern results from a dense network of pits separated by relatively narrow walls (Fig. 4). It is not easy to suggest a reason for the initiation of weathering pits at the rock surface. It was originally thought that hollowing might be related to xenolith distribution but the widespread occurrence of xenoliths as protruding knobs (Fig. 5) is evidence to the contrary. Only in two instances was a hollow seen to develop across a xenolith and in both cases it extended into the host rock. The distribution of pits at one site is highly irregular so that in one location there may be a honeycomb of pits and, a few metres away, none in an otherwise homogeneous host rock. Smith (1941) suggested spalling of granite chips and shells to produce shallow concavities. A further random process and the most likely in this case is glacierization, where natural depressions are left in the rock surface after recession of the ice. Unlike erosion cavities, weathering pits in their present form are not directly developed by wind action. This is confirmed through examination and staining of gravel samples collected from various hollows. Individual grains have angular shapes with shiny faces which, according to Cailleux (1969), signifies weak or no mechanical weathering. Also, observations showed no evidence for chemical erosion (Twidale and Corbin, 1963; Cailleux, 1969) and all the minerals were found to be present when compared with the composition of the host rock. These include perfect books of biotite, a mineral prone to attrition during wind transport (Ollier, 1969). To account for the vast volume of hollow compared with the small amount of gravel present and the fact that some hollows are deepened towards the east suggesting aeolian influence from the west, it is inferred that wind action was responsible for the removal of debris loosened from the bedrock until the hollow was of such a depth that the effect of wind action became minimal. Hence, the existing gravels in the sampled hollows are not likely to be affected by wind action unless removal is very quick. Van Autenboer (1964) has also stated that the role of wind in cavernous weathering seems to be limited to the removal

Fig. 4. Honeycomb weathering in a (?) granodiorite-gneiss (E.4020). The hammer shaft is graduated at 5 cm. intervals.

of grains loosened by other physical processes. However, wind abrasion may affect the lips of weathering pits. The loosening mechanism for rock fragments in the pits is believed to be two-fold. Diurnal multigelation was observed in some water-filled hollows but not always throughout their full depth, a phenomenon which would depend on the rate and extent of warming and cooling of the rock. Clearly, the surface layers of water would freeze first with a drop in air temperature and this process would tend to widen the pit. The second mechanism is possibly biological but it will only occur where lichens are present. Hollows provide a natural habitat for lichens which can hold a film of water and extract nutrients from the rock-forming minerals by ion exchange. They are swollen when moist and contract when dry, thus possibly breaking off mineral fragments which are absorbed into the lichen tissue (Ollier, 1969). Lichens were observed in some but not in all of the weathering pits. Wellmann and Wilson (1965) have stated that, in general, the insides of holes are not so exposed and therefore

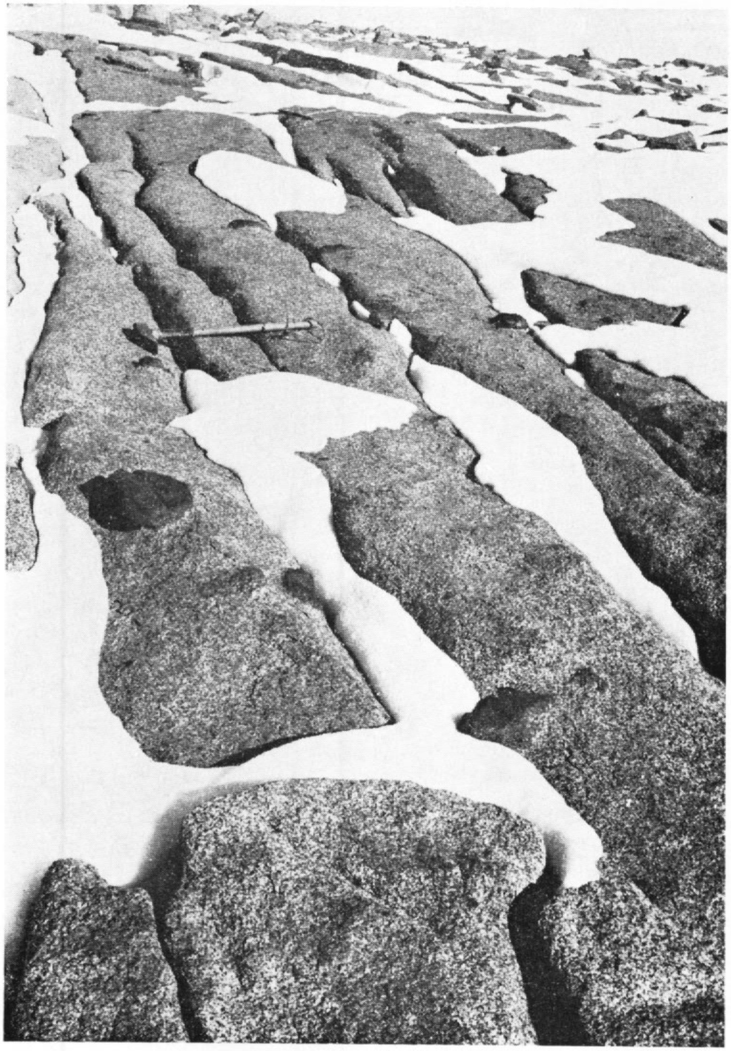

Fig. 5. Snow accumulation in eroded joints in a tonalite with protruding xenoliths (E. 4065). The hammer shaft is 60 cm. long.

temperature changes on these inner surfaces are less than on the outside of the rock. Thus, direct thermal stresses assisting surface granulation are not likely to implement cavernous weathering beyond a critical depth.

The most spectacular examples of erosion cavities are developed close to the plateau edge where the strongest winds occur. The coarse gravel trapped under the windward lips of sastrugi is sufficient evidence of the kind of material carried by the wind. This is further supplemented by ice particles which increase in hardness inversely with a decrease in temperature. (At $-50°$ F [$-45°$ C] they attain a hardness of 6 (Mohs' scale), which is equivalent to that of orthoclase (Teichert, 1939; Blackwelder, 1940; Hamelin and Cook, 1967).) The resultant carved forms have a totally irregular pattern and they vary from a completely sculptured rock face to individual cavities which can be up to 1 m. wide and 50 cm. deep. Farther east, cavities are less spectacular but they are still in evidence at long. 64°45′W., and all face west into the prevailing winds. Many boulders in these eastern areas have adzed or scalloped surfaces and some have become undercut with the consequent development of anvilar forms.

Raised annular rings are found only where destruction of the rock surface by exfoliation proceeds more slowly than by granular disintegration (Blank, 1951). These are raised rims which stand above pit and rock surface. Fig. 6 shows one of the few examples within this area;

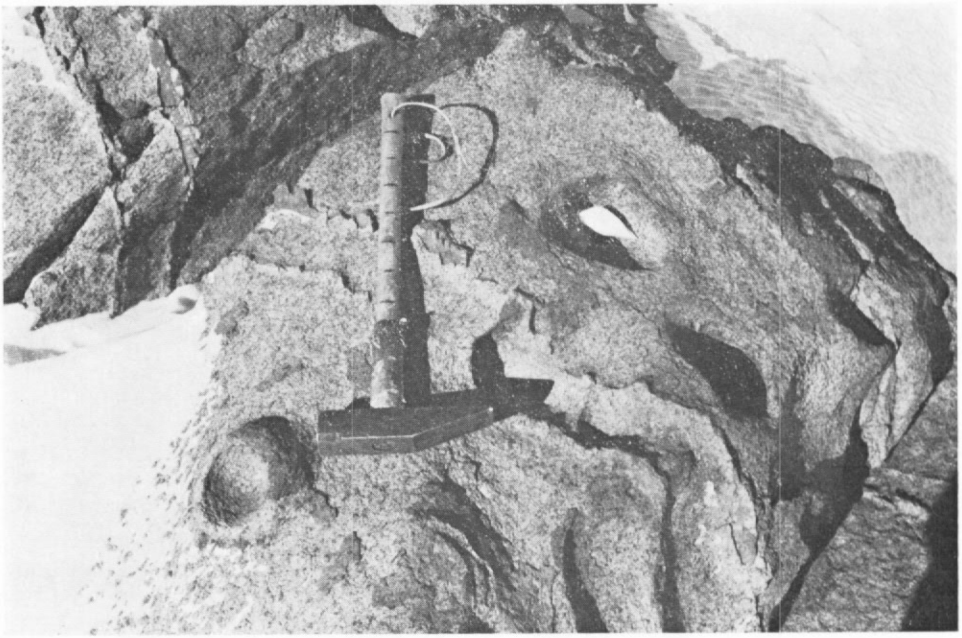

Fig. 6. A raised annular ring in biotite-gneiss (E.4008) cut by a differentially eroded quartz vein. The hammer shaft is graduated at 5 cm. intervals.

this cannot be explained lithologically since there appears to be no obvious compositional difference between the annular ring and the surrounding rock surface.

Occasionally quartz veins may project above the rock surface through differential wind erosion (Fig. 6). Also some dykes have been preferentially eroded to produce a trough feature in the rock surface, whereas others project above the surface as more resistant bands and this is especially so in slate outcrops.

Joints, particularly those developed in the intrusive rocks, have become eroded and subsequently widened at the surface. On vertical and sloping surfaces the main agent is likely to be abrasion by wind but, in addition, on sloping to flat areas, snow will accumulate in the joints (Fig. 5), providing moisture for congelifraction and possibly solution. In one instance, cavernous weathering appears to have combined with joint or fracture exploitation to produce complicated forms with narrow rock bridges (Fig. 7).

Sometimes, in association with areas of pitting, pock marking occurs. These small holes are usually not larger than 1 cm. in diameter on the rock surface (Fig. 8). There is no evidence to suggest their possible origin but wind is thought to be a major force in their development.

Patterned ground

On bare surfaces, frost induces comminution and sorting of the weathered rock debris and at the same time assists an upward movement of stones in the soil to form the surface layer.

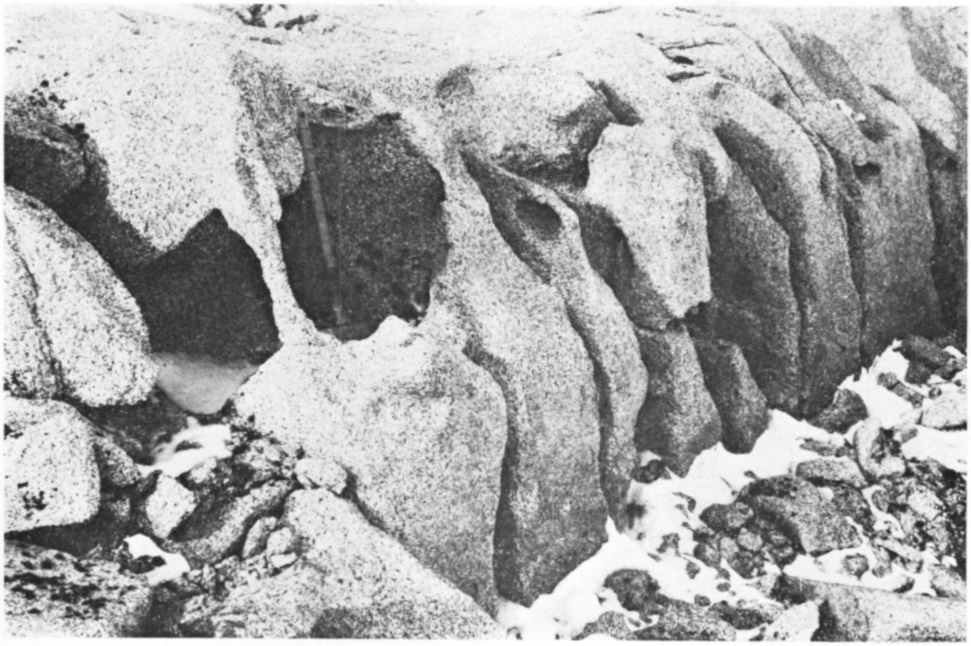

Fig. 7. Thin rock bridges across a complex erosion cavity in granodiorite (E.4076). The hammer shaft is graduated at 5 cm. intervals.

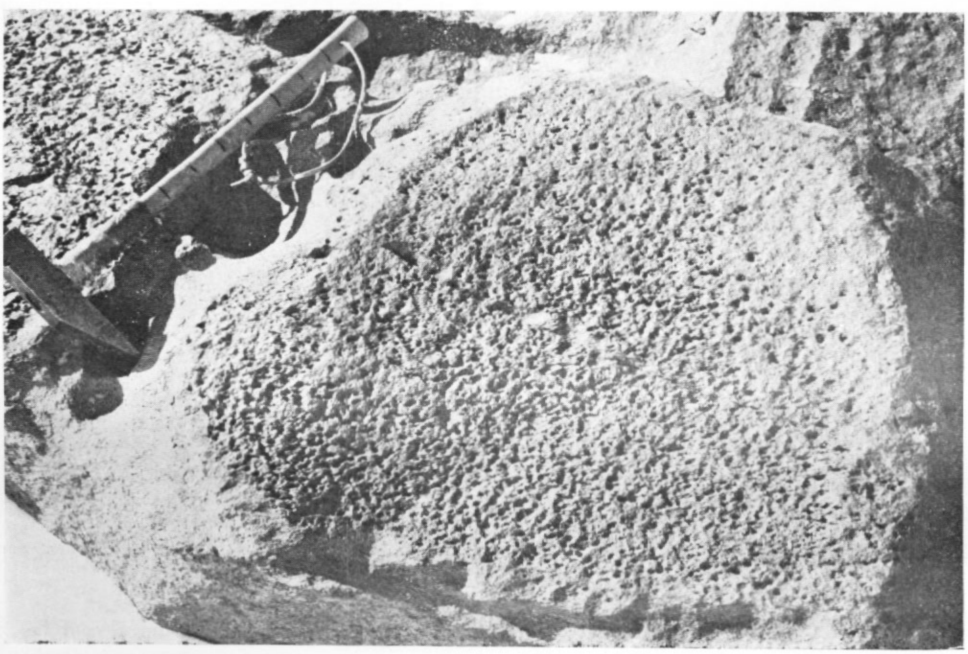

Fig. 8. Pock marks in biotite-gneiss (E.4008). The hammer shaft is graduated at 5 cm. intervals.

This layer is differentiated by various processes to produce patterns (Hollingworth, 1934). Patterned forms are present, mainly within the contact-metamorphic rocks and are mostly crude examples.

Non-sorted circles (Washburn, 1956) were observed at two sites and all of these have distinct domed surfaces enhanced by the surrounding snow cover. They consist of fairly coarse to medium blocks and have an average diameter of 1–2 m.

All other patterns are much smaller scale and they develop on flat to gently sloping areas in finely comminuted material but show a tendency towards sorted forms (Fig. 9). Their maximum diameter is usually not greater than 30 cm., a factor largely attributable to the inconducive environment for their formation. Whether the mechanism be through cryostatic movement or multigelation (Washburn, 1956), a suitable environment must be available for the freezing and thawing of material. For a freeze–thaw cycle to be initiated, a sufficient number of temperature fluctuations across 0° C must occur and these are probably developed through the direct effect of insolation rather than by air masses. During the summer, air temperatures rarely exceed 0° C yet the presence of water during the day in some weathering pits suggests a different temperature regime at the rock surface. The centres of these sorted circles are, in some examples, still active as evidenced by the small silty mud hummocks developed wtthin them. The formation of mounds of fines raised above the level of the stones is probably due to differential frost heaving. The concentrations of fine silt will retain more moisture than the stones and thus undergo greater expansion when freezing takes place (Chambers, 1967). When the frozen silty regolith thaws from above, excess water is lost and the discharge, plus the silty material, rises where resistance is least (Lundquist, 1969). These silt hummocks also occur outside circles (Fig. 10) but they are believed to be produced in the same way. They are distinctly fresh in appearance and are thus very recent in origin.

On the steeper gradients, the down-slope component causes this silty mud to be washed out, producing small linear streams of fresh material on the comminuted regolith surface. Fine silts and gravels have also been found as deposits below semi-permanent snow banks where subsequent melting in the summer provides a sufficient flow of water for the fines to be washed from the coarser scree.

Features of glacial transport

Glacial erratics are common throughout this area, the most frequently observed compositional variety being that of the leucocratic intrusive rocks. These, together with metamorphic complex types are particularly distinguishable on the dark contact-metamorphic rock outcrops but they also occur elsewhere. Sizes range from pebbles up to very large rounded boulders and they indicate that previous ice levels were much higher than the present-day ones.

Three examples of supraglacial moraines were recorded on the east side of Mount Jackson. Debris for these is supplied from scree-forming processes on the eastern flanks of this massif and the material is transported on an almost crevasse-free blue-ice surface. In all cases, these moraines are long narrow features up to 1·5 km. in length and at their distal ends they consist of only one or two boulders.

Conclusions

The central Black Coast is an area dissected by many inlets. The hinterland consists of gently undulating snowfields which descend towards the Larsen Ice Shelf in a series of irregular steps. This topography is dominated at the plateau edge by Mount Jackson and the Welch Mountains.

The vast areas of scree and felsenmeer, which indicate that rock weathering is particularly intense, greatly contrast with the fresher rocks along the west coast of the Antarctic Peninsula.

Fig. 9. Miniature sorted circles in quartz-biotite-gneiss debris (E.4070). The hammer shaft is graduated at 5 cm. intervals.

Fig. 10. Silt hummocks in gravel on weathered tonalite (E.4182). The hammer shaft is graduated at 5 cm. intervals.

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