THE PHYSIOGRAPHY AND SIGNIFICANCE OF THE TRANSITION ZONE BETWEEN GRAHAM LAND AND PALMER LAND

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ABSTRACT. The boundary between Graham Land and Palmer Land occurs in a zone about 150 km. wide where the central plateau is unusually dissected and across which the physiographical character of the Antarctic Peninsula changes in several striking respects. The topography and glaciation of this transition zone are described in detail and the modes of the physiographical transition along its west and east sides are emphasized. The origins of the three levels of planed surfaces in this region are examined and, on the basis of the tentative conclusions, an outline of the evolution of the physiography is constructed. With this background the structural significance of the transition zone is discussed and, in the absence of any positive differences in the geological or tectonic histories of at least southern Graham Land and northern Palmer Land, it is tentatively concluded that this feature is young. However, further study of the continental strip and the ocean floor to the west may reveal a discontinuity at the transition zone similar to those now known in other parts of the circum-Pacific orogenic belt.

THE Antarctic Peninsula is a sinuous cordillera nearly 1,500 km. long extending from the eastern end of Ellsworth Land, western Antarctica, northward to South America (Fig. 1). It consists of an almost continuous, narrow snow plateau drained by deep coastal glaciers but it is bisected by a zone which is physiographically distinctive and across which the physiographical character of the peninsula changes. This transition zone lies between lat. 68°10′ and 69°40′S., and is approximately 150 km. north–south (Fig. 2).

Parts of this area were visited by members of the British Graham Land Expedition in 1936–37, the United States Antarctic Service Expedition in 1940–41 and the Ronne Antarctic Research Expedition in 1947–48. Some trimetrogon air photography was also carried out over the area by members of the last expedition. Particularly in the past 15 years, most of this transition zone has been topographically and geologically mapped as part of the regional survey programmes of the British Antarctic Survey. In 1966–67, the United States Navy covered this area in its systematic trimetrogon air survey of the Antarctic Peninsula south of lat. 68°S.

The physiography of the east side of southern Graham Land north of the transition zone has been described by Marsh and Stubbs (1969) and that of northernmost Palmer Land south of the transition zone by Davies (1975).

PHYSIOGRAPHICAL CONTRASTS BETWEEN GRAHAM LAND AND PALMER LAND

There are five clear physiographical differences between the northern and southern parts of the Antarctic Peninsula:

i. The southern part of the peninsula is about three times wider than the northern part (Palmer Land is 210–240 km. wide and Graham Land is 65–100 km. wide) and even when the off-lying islands are included this contrast in the width of the cordillera is similar. However, this widening is not clearly traced by the continental slopes, although the western slope which trends south-westward off Graham Land turns west between the latitudes of the transition zone to define the margin of continental Antarctica.

ii. The western side of the Antarctic Peninsula is conspicuously different to the north and south of Marguerite Bay. West of Graham Land an archipelago extends up to 70 km. from the mainland, whereas the area west of Palmer Land is dominated by one large island (Alexander Island) which is separated from the mainland only by the 25–50 km. wide George VI Sound. Furthermore, both the geology and structure of Alexander Island are strikingly different from those of the mainland, unlike most of the smaller islands to the north.

iii. Looking northward there is a sharp sinistral curve to eastern Ellsworth Land and southern Palmer Land, and a broad dextral curve to Graham Land. The northern part

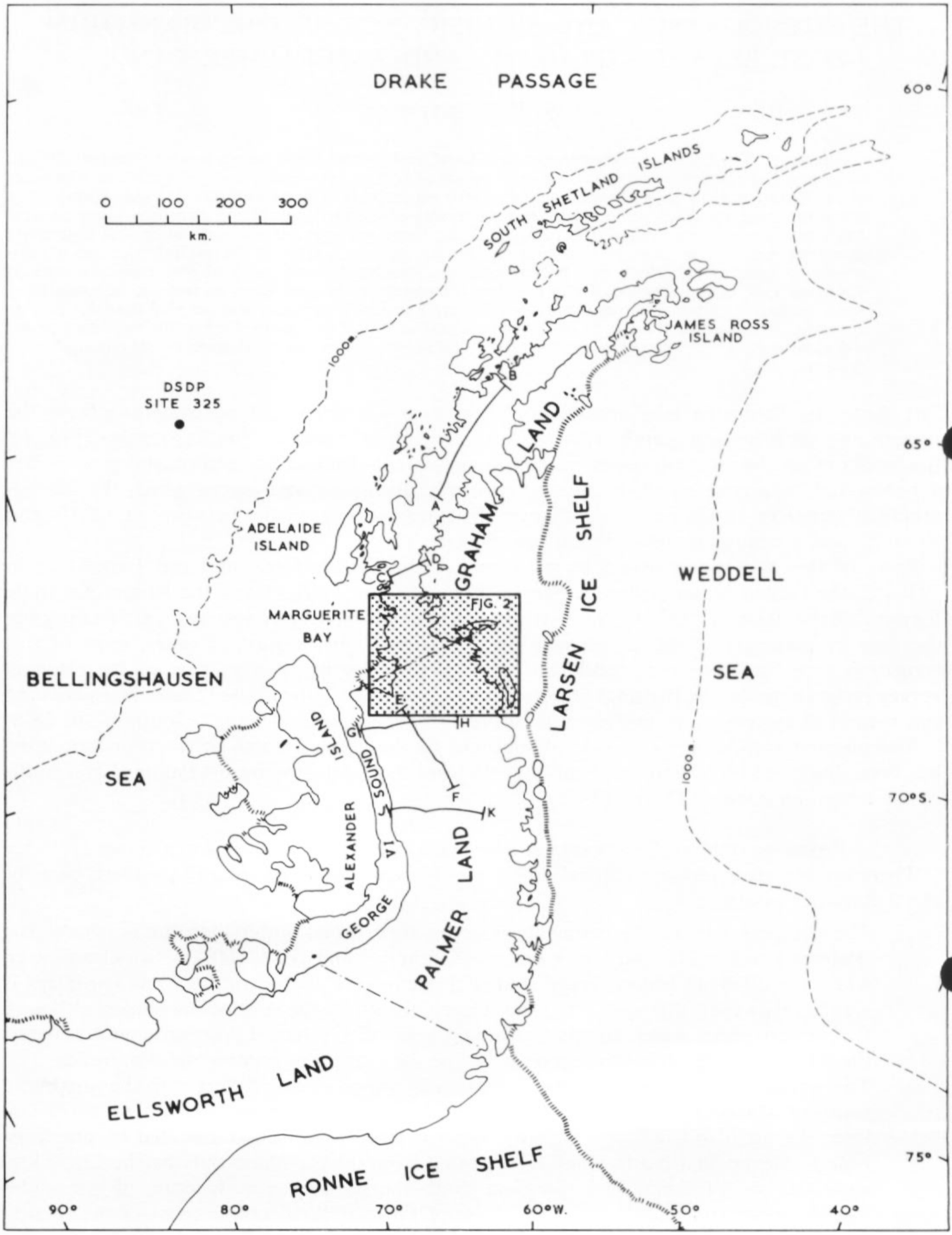

Fig. 1. Map of the Antarctic Peninsula showing the location of the transition zone (Fig. 2). The lines of the ice-depth profiles in Fig. 3 (A–B, C–D, E–F, G–H and J–K) and the position of site 325 of the Deep Sea Drilling Project (DSDP) are also marked.

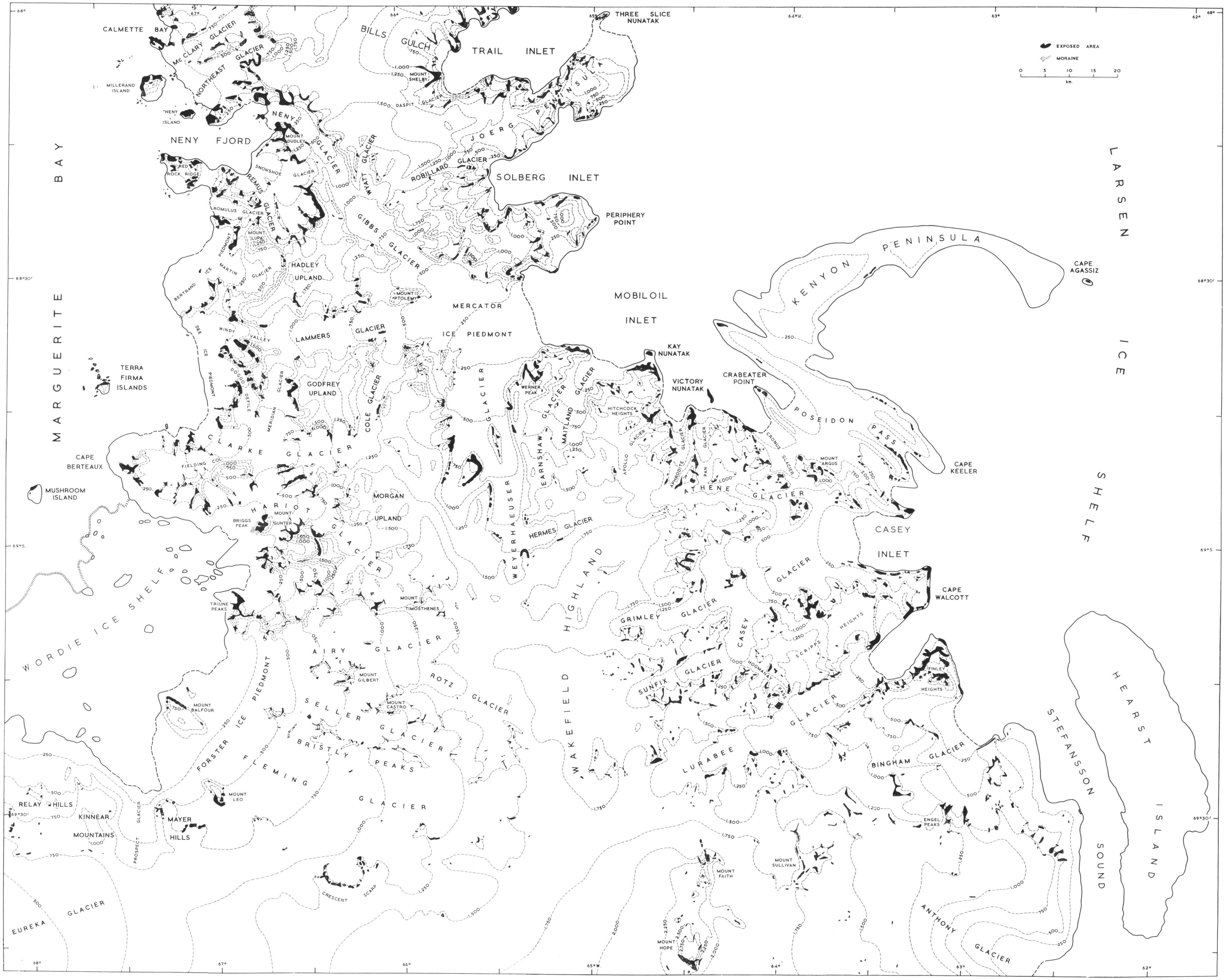

Fig. 2. Topographical sketch map of the transition zone between Graham Land and Palmer Land (see Fig. 1). Approximate contours are at 250 m. intervals.

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of Palmer Land is straight but the transition zone forms the beginning of the arc of Graham Land.

The ages and causes of these two curvatures are not known but the curve of Graham Land has attracted considerable attention recently because of the apparent complementary curve of southernmost Patagonia and its bearing on the origin of the Scotia arc. Dalziel and Elliot (1971, fig. 3) suggested that the curvatures of the cordilleras at either end of the Scotia arc date from the inception of the formation of the arc but Barker and Griffiths (1972, fig. 14) believed that they are primary features. However, Katz (1972, p. 332) suggested the curvature of southernmost Patagonia was due to oroclinal bending of an originally straight orogenic belt, whereas that of Graham Land is more likely to be primary, and this view is supported by recent palaeomagnetic data (Dalziel and others, 1973).

iv. The main glaciers of Graham Land are generally much narrower, shorter and steeper than those of Palmer Land, which are therefore apparently more mature. The coastal valley-glaciated belts of Palmer Land are consequently wider and the central snow plateau is less clearly defined than in Graham Land.

The differences in the typical main glaciers are probably in part due to different climates of Graham Land and Palmer Land sometime in the recent past if not today, but the change in glacier type is sharp and appears to be directly related to the different widths of the two parts of the peninsula. This may have resulted from the greater discharge from the larger ice cap in the south.

v. Although the snow plateaux forming the spines of both Graham Land and Palmer Land are fairly flat and are 1,500–2,000 m. high, the rock surfaces beneath them differ considerably in character. King (1964, p. 63) noted that the snow plateau of the peninsula is smoother in the south but recent radio echo-sounding data show clearly that the ice is much thicker in Palmer Land and that it masks the true nature of the subglacial surface. Profiles along and across the peninsula (Fig. 3), drawn from some of these data (Smith, 1972), confirm that the spine of Graham Land is a fairly flat rock plateau, whereas the surface beneath the snow plateau of northern Palmer Land is very uneven with slopes rising more than 1,000 m. in 2 km. horizontal (which is the spacing of the published ice-depth measurements) and mountains rising to 2,500–3,000 m. on its eastern edge. (Unfortunately no ice-depth data are yet available for southern Palmer Land.) However, it should be noted that there is evidence for a recognizable elevated planed surface in Palmer Land in the bevelled summits and more extensive summit plateaux of some of the coastal mountains. The profiles also show that, although the average altitude of the southern snow plateau is slightly greater than that of the northern one, the average height of the subglacial surface is not higher in the south.

The marked difference in the ruggedness of the terrain beneath the northern and southern ice caps has been attributed by Linton (1964, p. 94) to a difference in their pre-glacial nature.

It may be coincidental that the boundary between the two parts of the Antarctic Peninsula defined physiographically in some of the above ways (and the nomenclatural boundary between Palmer Land and Graham Land) occurs in the transition zone, but at least the differing widths and topographies of the western coasts of the two areas indicate that there is a major structural break in this zone. This is further substantiated by the distinctive physiography of the zone itself.

PHYSIOGRAPHY OF THE TRANSITION ZONE

The transition zone is most easily defined as the unusually dissected area between the main snow plateaux of Palmer Land and Graham Land where the peninsula widens. On its west coast, the peninsula narrows by virtue of a sudden indentation of the coastline to the south of the Wordie Ice Shelf but on the east coast this is more gradual and the coastline trends north-west to south-east throughout the zone (Fig. 1). The southern boundary of this zone is marked by the conspicuous indentation of the western coastline and the northern limit of the main Palmer Land plateau at Fleming and Lurabee Glaciers (Fig. 2). This physiographical break is particularly impressive in the west, where a major rift was reported by Rymill (1938, p. 219–25) and confirmed by King (1964, p. 63), but Fleming and Lurabee Glaciers do not form a continuous trough since they are separated by an indefinite col at a height of 1,700 m. between Wakefield Highland and the Palmer Land snow plateau. The most obvious northern boundary of the zone is the Neny Glacier–Gibbs Glacier trough but the southern limit of the continuous Graham Land plateau is marked by Robillard Glacier, so the dissected area north of the Mercator Ice Piedmont is included in the following description of the transition zone. The zone is considered from north to south in western and eastern parts divided by Neny, Gibbs and Weyerhaeuser Glaciers, and Wakefield Highland (Fig. 2).

Western part of the transition zone

The area between Neny and Fleming Glaciers comprises the southern part of the Fallière Coast but in the north the area discussed here almost reaches the east coast of the peninsula. This part of the transition zone clearly demonstrates a transition in physiography from that typical of Graham Land, which is dominated by plateau scarps, southward to a less spectacular and apparently more mature terrain typical of north-west Palmer Land. It is described in three main parts.

Neny Glacier–Clarke Glacier area

It is in this northern part of the transition zone that the plateau has been most deeply dissected and the glacial divide has been displaced from its usual median position to nearer the west coast. The features of the east and west sides of the divide differ considerably in this area.

The physiographic structure of this area is dominated on the east coast by the Mercator Ice Piedmont into which four main glaciers flow: Gibbs, Lammers, Cole and Weyerhaeuser Glaciers. Two of these are the eastward-draining parts of the two almost straight troughs that cut across the peninsula in the transition zone, the north-west to south-east Neny Glacier–Gibbs Glacier trough and the west–east Windy Valley–Lammers Glacier trough. Snow cols rise to over 1,000 m. in both troughs but even at these high points the troughs are sharply defined. Cole Glacier and Weyerhaeuser Glacier (about 45 km. long) trend north–south like the other two glaciers on the south side of the Windy Valley–Lammers Glacier trough. All these glaciers except Weyerhaeuser Glacier rise not on plateaux but at interglacial cols. This intersecting glacial pattern has left isolated plateau blocks which are still as high or almost as high as the main snow plateaux bordering the transition zone. Their ice caps are drained into the main glaciers by many minor glaciers which are generally short and steep but some are larger such as the two that have developed on Godfrey Upland and have almost bisected that massif (Fig. 4).

Like most of the coast of Graham Land to the north, the west coast between Neny Fjord and the Cape Berteaux peninsula is dominated by the plateau scarp scalloped by cirques and breached by a few short, steep, severely crevassed glaciers. Narrow, sharply defined ice piedmonts occupy much of this coastal strip but they differ fundamentally from the Mercator Ice Piedmont on the east coast since they are not nourished by much inland ice. The northern part of this area is particularly deeply eroded and the cirque glaciers have developed into coastal glaciers with subsidiary cirques. Only ridges remain of the former plateau surface except for Mount Lupa, an isolated plateau remnant 1·5 km. across.

The snow plateaux are undulating and they have a few distinct snow summits but rock is

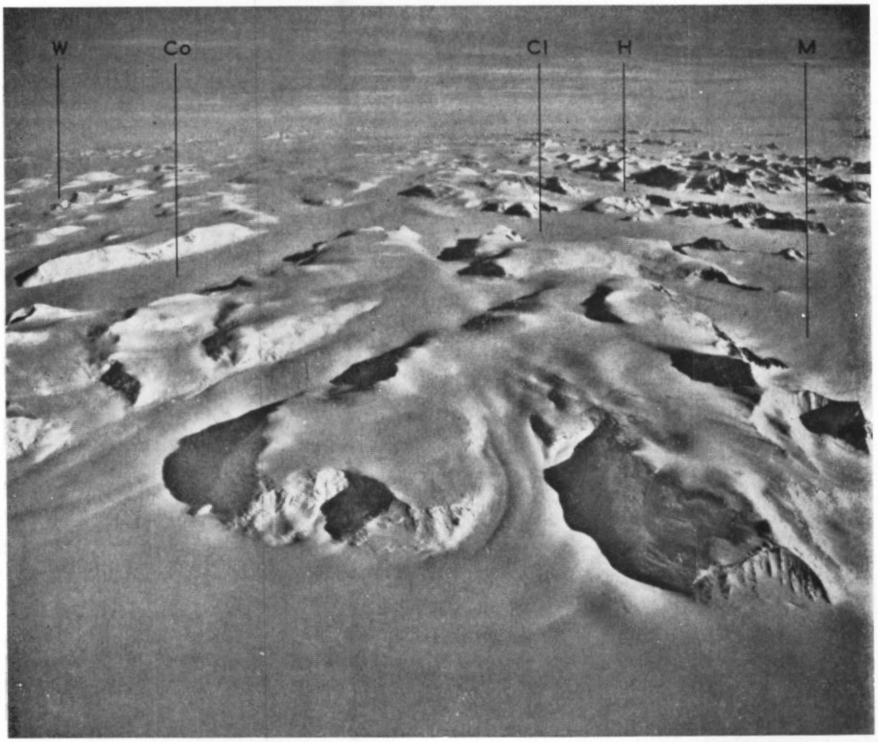

Fig. 4. South-eastward view of the plateau blocks of Godfrey Upland and Morgan Upland, east of the dissected coastal belt. The main glaciers are Weyerhaeuser Glacier (W), Cole Glacier (Co), Clarke Glacier (Cl), Lammers Glacier (foreground), Hariot Glacier (H) and Meridian Glacier (M). (U.S. Navy photograph.)

only exposed on their marginal scarps, particularly in the lower parts of the main valleys and the headwalls of coastal cirques. Many of the coastal ridges are also mostly ice-free.

Clarke Glacier–Airy Glacier area

The dissection of the plateau in this area has not reached as advanced a stage as it has north of Clarke Glacier and the plateau blocks have not yet been completely isolated, but the west side of the plateau has been eroded back from the coastline as far as it has been north-west of Hadley Upland.

The plateau in this area is nearly cut off from Wakefield Highland by Weyerhaeuser and Airy Glaciers, and it is partly incised by Hariot Glacier and many smaller glaciers (Fig. 4). As in the north, the coastward sides of the plateau blocks are indented by cirques and breached by a few minor, severely crevassed glaciers 2–3 km. wide, and the north-west part of Morgan Upland and the plateau west of Hariot Glacier have been largely reduced to complex ridge systems (Fig. 5). Much of the area between Clarke and Airy Glaciers is drained by Hariot Glacier into the north-east corner of the Wordie Ice Shelf, which is 400 m. deep and is the thickest part of the ice shelf (Smith, 1972, map M).

The plateaux are as barren of rock exposure as in the north except at the southern end of Morgan Upland, where there is a group of nunataks around Mount Timosthenes (2,058 m.). These are the northernmost plateau nunataks which are typical of the margins of the snow plateau of north-western Palmer Land. Again, much more rock crops out in the coastal belt, on the ridges extending west from the plateau scarps.

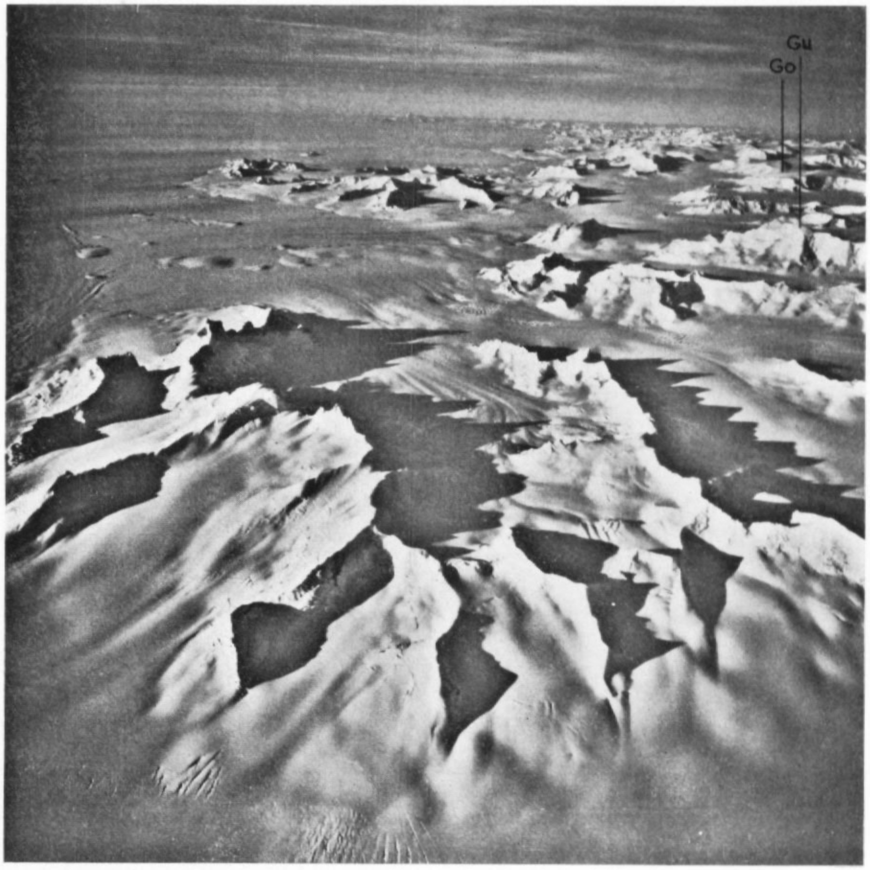

Fig. 5. North-westward view of the dissected coastal belt west of Godfrey Upland (Go) and Morgan Upland, including Mount Gunter (Gu; 1,970 m.) and Triune Peaks (centre left; 1,130 m.). On the north side of the Wordie Ice Shelf (where there are several ice rises) the Cape Berteaux peninsula projects into Marguerite Bay. (U.S. Navy photograph.)

The *Cape Berteaux peninsula* projects about 25 km. into Marguerite Bay, confining the Wordie Ice Shelf to the south. It is physiographically independent of the mainland (Fig. 5). Its main topographic features are an east–west median ridge 1,000–1,150 m. high with subsidiary ridges trending north and south, and, isolated at the end of the peninsula, a degraded block mountain rising to 1,200 m. and surrounded by a narrow ice piedmont. On the northern side of the longitudinal ridge a trough breaches the ends of the northern ridges allowing a local ice cap, Fielding Col, to drain both westward and eastward. Little rock is exposed on the longitudinal ridge but much is seen on the other ridges, particularly near their coastward ends.

Airy Glacier–Fleming Glacier area

There is a dramatic change in physiography on the northern side of Airy Glacier. The deeply dissected plateau surface with residual ridges to the west (described above) gives way abruptly southward to more open terrain with larger glaciers and discontinuous ridges merging more gradually into the central snow plateau. This ice cap, Wakefield Highland, is a narrow northerly extension of the main Palmer Land snow plateau which also dips westward. It drains westward by two main glaciers, Fleming and Airy Glaciers, into the Forster Ice Piedmont at the head of the southern half of the Wordie Ice Shelf.

The northern margin of Airy Glacier is an almost continuous barrier formed by the south sides of plateau blocks including Morgan Upland and the longest and most complex ridge extending westward from them, which reaches the coast at Triune Peaks (1,130 m.). The southern margin is less definite since the plateau is more rapidly reduced westward in the form of the degraded massif that emerges from Wakefield Highland and terminates west of Mount Castro (1,630 m.), the eroded block mountain of Mount Gilbert (1,420 m.; Fig. 6), and finally a very subdued horseshoe ridge still more than 20 km. from the coastline. Although it is transected at its eastern end, the eastern part of the next ridge south (Bristly Peaks) is also a remnant of the plateau and it is also progressively degraded westward. Seller Glacier lies between these two ridges but it does not drain Wakefield Highland directly, since there is a group of plateau-edge nunataks 15 km. wide at the head of the glacier (Fig. 6). It is not nourished either by Fleming Glacier with which it is almost joined by a low col at its head but, although it is only 2–4 km. wide like the local glaciers to the north, it is known to be 400–900 m. deep (Smith, 1972, map E). Although rock exposure is moderate around Seller Glacier, there is generally less rock in this area than north of Airy Glacier.

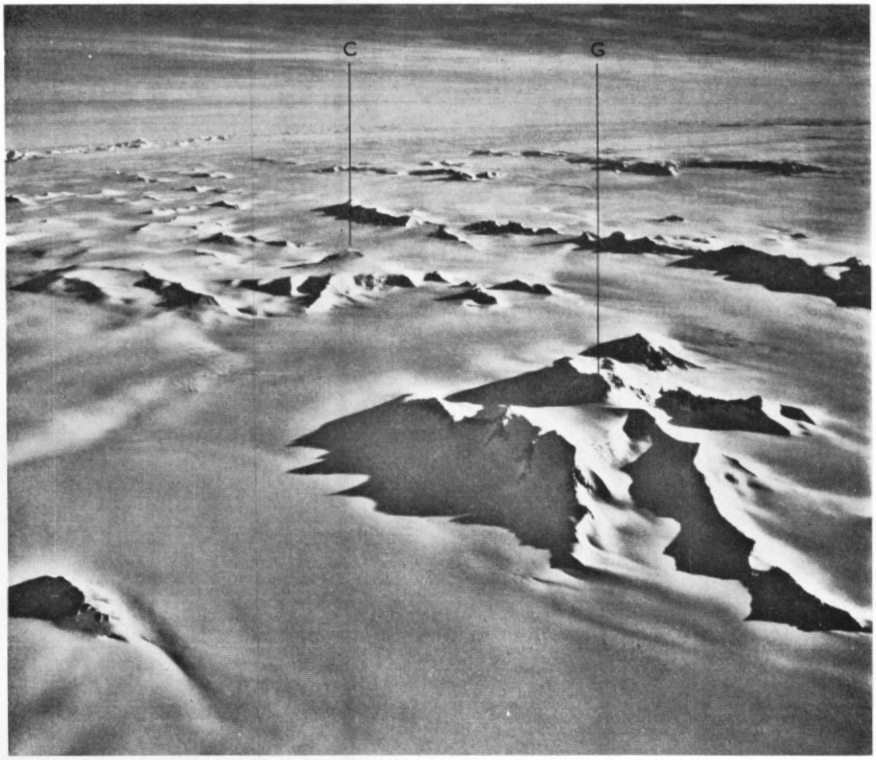

Fig. 6. South-eastward view of the eroded massifs between Airy Glacier (foreground) and Fleming Glacier (background) including Mount Gilbert (G; 1,420 m.) and Mount Castro (C; 1,630 m.). The discontinuous scarp (right horizon) on the south side of Fleming Glacier forms the northern margin of the Palmer Land snow plateau; the Eternity Range forming its eastern margin is on the left horizon. (U.S. Navy photograph.)

Fleming Glacier is the southernmost and largest glacier in the transition zone. It is 60 km. long and in its central reaches, where it is widest, it is 1,300 m. deep (personal communication from C. W. M. Swithinbank) and its floor is 300 m. below sea-level. The south side of the glacier is a broken scarp which continues westward along the south side of the ice piedmont and ice shelf, demarcating the northern margin of the Palmer Land snow plateau. It is known that the ice in the breach in this rampart east of Crescent Scarp is as deep as 1,600 m. in places (personal communication from C. W. M. Swithinbank).

The Forster Ice Piedmont, like the Mercator Ice Piedmont, is a large lowland snowfield nourished by major inland glaciers and merging seaward into an ice shelf, but it is extensively crevassed and less clearly defined inland. The boundary between the ice piedmont and ice shelf is also not clear but radio echo-sounding has shown that the ice at the edge of the ice piedmont is 500–600 m. thick, whereas near the adjacent edge of the ice shelf it is only about 250 m. thick (personal communication from C. W. M. Swithinbank). Although the ridges west of Wakefield Highland peter out westward to give way to the ice piedmont, an isolated block about 10 km. long and 1,000 m. high stands by the coastline.

Eastern part of the transition zone

This area between Robillard and Lurabee Glaciers comprises the southern end of the Bowman Coast and the northern end of the Wilkins Coast, meeting at Kenyon Peninsula. The eastern part of the transition zone also demonstrates the physiographical transition between Graham Land and Palmer Land but not as clearly as the western part. For the following description, it is divided into three main parts mainly on the basis of their dominant glacial trends.

Area east of Gibbs Glacier

Between Robillard Glacier and the Mercator Ice Piedmont, the Graham Land plateau is deeply dissected by closely spaced, north-east to south-west trending glaciers and cirques (Fig. 7) but as many snow peaks higher than 1,700 m. survive as on the main plateau to the north. The incised nature of this area, which is less than 25 km. wide, is the result of glacial erosion from low base levels on both sides: Solberg Inlet on the north-east side and Gibbs Glacier and the Mercator Ice Piedmont on the south-west side. The Periphery Point peninsula, which separates Solberg and Mobiloil Inlets, is as mountainous as Joerg Peninsula to the north and both are bisected by passes at about 400 m. and terminate in massifs about 1,100 m. high.

South coast of Mobiloil Inlet

The area between Weyerhaeuser and Cronus Glaciers is occupied by many minor glaciers draining northward to the Larsen Ice Shelf in Mobiloil Inlet. Its southerly limit is marked by the long narrow east–west trough of Athene Glacier, which prevents much of the Wakefield Highland ice cap from draining northward. The western glaciers in this area are separated by subdued snow ridges and, although the eastern ones are generally more deeply incised, little rock is exposed except adjacent to Aphrodite Glacier. On the other hand, rock exposure is exceptional in the coastal cliffs and off-lying Kay and Victory Nunataks.

Kenyon Peninsula is a conspicuous feature on the map but not on the ground; although it extends about 80 km. into the Larsen Ice Shelf, it is a snow plateau only about 300 m. high. It is generally clearly delineated in the ice shelf by a marginal steep snow slope but very little rock is exposed in this scarp. Kenyon Peninsula is therefore very unlike the mountainous peninsulas already described to the west and north but it resembles the low islands off the east coast of Palmer Land, such as Hearst Island which is 70 km. long. Its hinterland is a dissected snow plateau about 500 m. high which increases in height and irregularity south-westward to Mount Argus (1,220 m.). Large parts of the scarps of the plateau ridges, particularly in the eastern faces, are snow-free. Although the troughs are not continuous from coast to coast, this area has a strong north-west to south-east grain which is in marked contrast to the north-

Fig. 7. North-westward view of the dissected part of the Graham Land plateau south of Robillard Glacier (R). The peaks of Adelaide Island are on the left horizon, but the Larsen Ice Shelf in Solberg Inlet (S) and Trail Inlet (T) is obscured by low cloud. (U.S. Navy photograph.)

trending glaciers to the west and approximately east-trending glaciers to the south, but it is probably significant that it parallels the general strike of the south coast of Mobiloil Inlet.

Area east of Wakefield Highland

The glaciers in this area are generally longer than those to the north and in some respects the general physiography resembles that of the area west of most of Wakefield Highland, but the east side of this part of the peninsula has not been eroded to the same extent as the west side and plateau remnants 1,000 m. high reach the coast.

Although several glaciers drain Wakefield Highland, only two debouch into the ice shelf, both carving marked indentations in the coastline. Lurabee Glacier, the southernmost one, is the only glacier flowing directly from the plateau to the ice shelf and, although it is almost as long as Fleming Glacier, it is much narrower and a little steeper (with an average gradient between 500 and 1,500 m. of 1 : 48 compared with 1 : 56 in Fleming Glacier) and it is extensively crevassed. Most of the glaciers north of Lurabee Glacier are even narrower and steeper, and less than 30 km. long, because they discharge into Casey Glacier which transports most of the ice draining eastward from Wakefield Highland around Scripps Heights into Casey Inlet. The plateau blocks of Scripps Heights and Finley Heights, which are truncated by the coastline, are largely surrounded by impressive rock walls and, as in the hinterland of Kenyon Peninsula, the summit plateaux become lower and flatter towards the coast, sometimes descending in distinct steps (Fig. 8).

Fig. 8. Westward view of Scripps Heights between Lurabee Glacier (centre) and Casey Glacier (far right). Note the coastward stepped drop in the height of Scripps Heights towards the right. (U.S. Navy photograph.)

The area south of Lurabee Glacier is much less dissected than that to the north in the transition zone; east of the main Palmer Land snow plateau, ridge systems and nunatak groups occur in open snowfields with glaciers generally only becoming well defined near the coast, and the eastern edge of the central ice cap south of the transition zone is marked by a discontinuous string of high mountains. However, the tendencies for the snow plateau to fall away eastward in a series of steps, and for rock exposure to be restricted to patches and ribbons of *felsenmeere* on ridge crests, are also found north of Lurabee Glacier, particularly in the higher western area (Fig. 9).

PLANED SURFACES

The existence of a central plateau and many coastal ice piedmonts in Graham Land was clear to early visitors to the area but their origins are still debated. The systematic mapping of much of the Antarctic Peninsula over the past 25 years has revealed other planed surfaces between these two levels, and these are equally puzzling. It is difficult to deduce the modes and times of formation of any of these surfaces with confidence because their characters are still masked

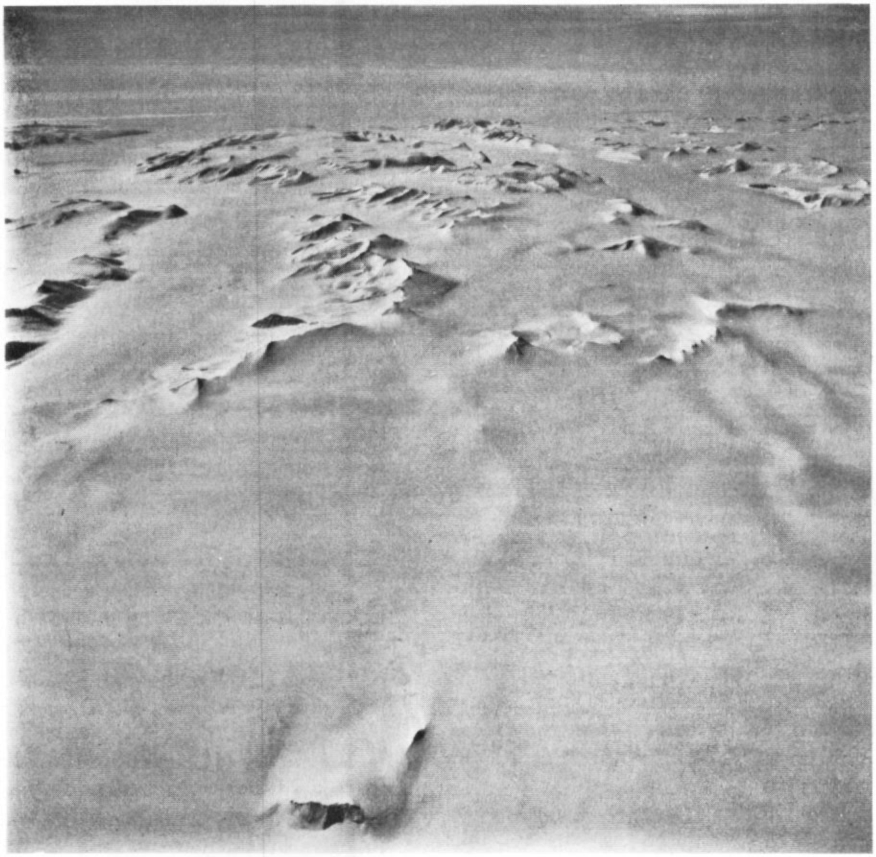

Fig. 9. Eastward view of the low poorly exposed ridges of western Scripps Heights (far centre) and between Sunfix and upper Lurabee Glaciers. (U.S. Navy photograph.)

by an ice mantle. Since their interpretation is of major importance in deducing the geomorphological history of the region, the main evidence relating to the origins of all three levels of planed surfaces warrants detailed attention.

The plateau surface

Although ice-depth radio echo-sounding has shown that the subglacial surface of the northern Palmer Land plateau is much less even than the Graham Land rock plateau, it has been pointed out that there is evidence for an elevated flat surface in Palmer Land (p. 41), and there is also a similar surface in Alexander Island, although it seems to be tilted from about 3,000 m. in the north to 300 m. in the south (Horne, 1967, p. 19). Another plateau surface of a similar age, 1,100–1,800 m. high, has been described from eastern Ellsworth Land (Laudon and others, 1969), and therefore a high narrow plateau appears to continue for about 1,700 km. along the Antarctandes. It should be noted that Tertiary erosion surfaces have also been found in western Ellsworth Land and parts of eastern Antarctica (Nichols, 1970).

Although they had no firm evidence, Nichols (1953, p. 10–12) and most subsequent workers have assumed that the plateau surface was formed as a fluvially eroded peneplain. However, in discussing northern Graham Land, Linton (1964, p. 94) stated that "as the plateau surface undulates the convexities and culminations correspond with the divides between the glacier troughs", suggesting to him that the glaciers were initiated in pre-glacial valleys. He also believed that the surface in northern Palmer Land is less even than to the north, and suggested that the southern area was located in the pre-glacial headwater area.

The erosion of the surface was completed some time *after* the youngest intrusions exposed on it were emplaced and this may have been as late as the Eocene (Rex, 1971, p. 133). No younger rock formations (which are mainly volcanic) provide time limits to the age of the surface in this region; although volcanic rocks apparently fill deep valleys in a plutonic terrain on Anvers Island (Hooper, 1962, p. 66), their age and relation to the summit plateau are not known. If the plateau surface had been eroded fluvially, it must have taken place *before* the onset of glacial conditions in this region. Substantial evidence of several types has recently indicated that at least continental Antarctica has been enveloped in a major glaciation since the late Tertiary (Dorman, 1966; Alt, 1968; Goodell and others, 1968; Denton and others, 1970; Rutford and others, 1971) or even since the early Tertiary (Margolis and Kennett, 1970; LeMasurier, 1971). If the glaciation began on the Antarctic Peninsula (or even only in eastern Ellsworth Land) by the mid Tertiary, the time available for the planation may thus be limited. From a study of some British unconformities, Linton (1957, p. 61–62) inferred that "on continental margin an erosion cycle can be carried to peneplanation in twenty to forty million years according to the magnitude of the initial relief", and this statement has been more or less endorsed by Schumm (1963, p. 9). However, the early Tertiary relief of this area may have only been moderate and it was apparently not entirely reduced to a peneplain. Furthermore, the Antarctic Peninsula and off-lying islands have formed an isolated continental strip at least throughout the Cenozoic and they have therefore been liable to particularly rapid erosion.

Another implication of the formation of the plateau surface by fluvial erosion is that it was uplifted remarkably uniformly by about 1·5 km. to its present level. Linton (1964, p. 97) believed that it was uplifted in a broadly arched form but the model of King (1964, p. 63) invoked extensive block faulting. In spite of the difficulty of proving faults in this area where rock exposure and suitable stratigraphical control are sparse, major faults parallel to the structural trend of the Antarctic Peninsula have been found on the west coast of Graham Land by Goldring (1962, p. 49), Hooper (1962, p. 66), Curtis (1966, p. 49–50) and Dewar (1970, p. 59–62), all of whom inferred that the peneplaned peninsula was uplifted between series of longitudinal faults. Some longitudinal faulting has also been reported from the west side of Palmer Land (Rowe, 1973, p. 71; Skinner, 1973, p. 20), although it is not as clear as farther west in Alexander Island, where major north–south faults downthrow to the east as well as to the west (Bell, 1974, fig. 3). Fewer analogous faults have so far been found on the east side of the peninsula, but Fraser and Grimley (1972, p. 55) believed that major faults on the east coast of the transition zone were involved in the elevation of the plateau, and Knowles (1945*a*, p. 136) implied that the large low islands off the east coast of Palmer Land have been down-faulted.

The age and the downthrow direction of many of these reported longitudinal faults are often not clear, but it is suggested that the difference in the altitudes of adjacent high erosion levels generally indicates post-planation faulting (although it is pointed out below that not all planed surfaces at intermediate levels are down-faulted parts of the plateau surface). Using this criterion, it is found that the most important tilting, folding and faulting of the Mesozoic rock formations occurred before the erosion of the plateau surface. There is good evidence for this in the transition zone, where a particularly clear longitudinal fault system can be traced for about 250 km. along the entire length of the Fallières Coast. Since it divides a western Mesozoic volcanic terrain from a topographically higher metamorphic terrain to the east, a considerable upthrow inland is indicated, and at least most of this movement preceded the regional planation because the plateau remnants on Arrowsmith Peninsula and peaks on nearby islands in Marguerite Bay are as high as the mainland snow plateau. Also, much of the faulting in

Alexander Island, which has the most obvious north–south structure in this region, is thought to have occurred before the formation of the elevated surface (Bell, 1974, p. 12).

The evidence for a great post-planation uplift of this region from known faulting is thus equivocal but the predominant longitudinal trend of the Tertiary post-batholith dykes over most of the Antarctic Peninsula confirms that the region was at least subject to east–west tension for much of this period. More positive evidence of uplift during the past few million years is found at the northern end of this region, although this is not conclusive either because it is a volcanic terrain which cannot be assumed to have been stable. It comes from the south-eastern side of Trinity Peninsula in the outcrop of the James Ross Island Volcanic Group, whose stratigraphical relations indicate a Miocene age (Adie, 1953) but radiometric dating has suggested the formation may be younger (Rex, 1971, p. 135). In the type area it comprises five alternations of olivine-basalt lavas and palagonite-breccias totalling at least 900 m. and perhaps as much as 1,500 m. (Nelson, 1965, p. 182). Similar sequences in Antarctica have recently been interpreted as due to subglacial eruptions (LeMasurier, 1971) but structures at the breccia/lava boundaries have been thought to indicate that these particular rocks were deposited near sea-level (Jones and Nelson, 1970). An intermittent submergence approximately equal to the total thickness of the volcanic succession is therefore implied and a post-late-Tertiary uplift of approximately twice this amount has raised the rocks to their present levels. Small outcrops of the same formation occur on Trinity Peninsula (although not above 500 m. a.s.l.) and Nelson (1965, p. 182) believed the mainland was submerged and then uplifted at the same time as James Ross Island, but it is unlikely that a final regional uplift of up to 3,000 m. is indicated.

A wave-cut platform 220–250 m. a.s.l. on Cockburn Island, east of James Ross Island, has been cut into lavas of the James Ross Island Volcanic Group and resting on this is the Pliocene or Pleistocene "Pecten Conglomerate" (Adie, 1964*a*, p. 28), indicating an uplift of this order probably in the Quaternary. A similar shore deposit has been found on a wave-cut bench on the south coast of King George Island in the South Shetland Islands but this area has been unstable for much of the Cenozoic and the bench has been tilted and faulted (personal communication from M. R. A. Thomson).

Since Arçtowski (1908) first realized that the ice sheet in the Antarctic Peninsula region was once much thicker than it is at present, further evidence in support of this has accumulated (Fleming, 1940, p. 95; Knowles, 1945*b*, p. 174; Nichols, 1953, p. 26–28; Adie, 1964*b*, p. 155; Dewar, 1967, table I). It also now seems that the glaciation of this region is older than previously thought and therefore it is possible that glacial erosion has been effective more widely than believed by Linton (1963) and other workers. The possibility must therefore be considered that ice planed the plateau surface of the Antarctic Peninsula region, which may have been domed before the ice level was reduced and cirque erosion could become effective and force the retreat and sharpening of the plateau escarpment. Marginal faulting and regional uplift of the peninsula are neither a prerequisite to nor precluded by this hypothesis. However, planation by glacial scouring would presumably become less efficient farther away from the South Pole and thus does not explain why the peninsula's plateau surface is much less regular in northern Palmer Land than in Graham Land. (The available data (Blundell, 1962, p. 22; Dalziel and others, 1973, fig. 4; Kellogg and Reynolds, 1974; Reynolds and Kellogg, 1974; Scharnberger, 1974) indicate that the orientation and palaeolatitude of the Antarctic Peninsula have not changed much since the beginning of the Tertiary.) Furthermore, although recent extensive flat glaciated surfaces are now exposed in the north of the northern continents and surfaces eroded during earlier glaciations have been reported (including the Upper Palaeozoic Maya erosion surface of McKelvey and others (1970) in Victoria Land, Antarctica), it has not been established that an ice sheet could flatten moderate relief in the duration of an ice age.

The absence of an obvious summit-plateau surface in the southern Andes (which is believed to have had a broadly similar history to the Antarctic Peninsula) may appear to support a glacial origin of the peninsula's plateau, but a similar surface in the Andes may have been obliterated by thick Cenozoic volcanic forms and rapid subaerial erosion. Favourable circumstances have, however, preserved Cenozoic erosion surfaces in north-central Chile and they indicate that much of the Andean relief formed earlier in the Tertiary than previously thought (Sillitoe and others, 1968; Mortimer, 1973).

Since the Antarctic Peninsula may have been subjected to periglacial conditions for much of the Tertiary, the possibility that the formation of the plateau was due to cryoplanation (Bryan, 1946; Peltier, 1950) should also be considered. Although this process may theoretically have eroded the surface at its present level, the plateau would be expected to become higher and less regular northward if this had been the case. Moreover, even if cryoplanation is considered a possible major agent of erosion, it is unlikely that it could level a whole region with such a latitudinal range.

Although it now appears that the summit erosion surface of the Antarctic Peninsula and off-lying islands represents only a partial peneplain, it is concluded that it was probably formed by a process of fluvial erosion. In spite of the weight of opinion previously expressed in favour of this hypothesis, it should be stressed that the available evidence is still not decisive.

Sub-ice-piedmont surfaces

Ice piedmonts are commonest around the coasts of Graham Land, on the off-lying islands and on Alexander Island where they are particularly extensive. There are two types and they are both represented in the transition zone:

i. The Forster and Mercator Ice Piedmonts are true piedmont glaciers nourished by several major drainage glaciers. They are deep and almost flat. The Forster Ice Piedmont is generally 650–950 m. deep in the north (personal communication from C. W. M. Swithinbank) and the rock floor beneath it is approximately 500 m. below sea-level. This underlying surface has clearly been eroded by the ice discharging over it.

ii. The Bertrand and Dee Ice Piedmonts are narrower and are backed by high plateau walls which are sculptured by cirques but are not breached by glaciers draining much of their hinterlands. Since their ice masses are locally derived, these ice piedmonts do not constitute piedmont glaciers in the genetic sense, and Fleming (1940, p. 93) called them "fringing glaciers". Especially near the coast, they are generally steeper than true piedmont glaciers, and in places solid rock can be seen at the base of their ice fronts suggesting that only parts of their floors are below sea-level; for technical resons, echo sounding of ice piedmonts has proved difficult (Smith, 1972, p. 9) but reliable results obtained from the western part of the Bertrand Ice Piedmont confirm that it is only about 200 m. thick there. The supposed slow movement of the "fringing glaciers" and the apparent lack of rock debris seen in their ice fronts suggest (but do not prove) that they have not eroded their rock benches, so their origin has been disputed.

Holtedahl (1929, p. 17–18) believed that "fringing glaciers" rest on strandflats formed by rapid headwall erosion of cirques by freeze–thaw action. Koerner (1964, p. 5) could not accept this hypothesis because he considered that the margins of the plateau of Trinity Peninsula are too even to have been produced by piece-meal headwall erosion, and most of the scarps are ice-covered and thus presumably protected from freeze–thaw action. He concluded that these piedmonts are pre-glacial in origin. Marsh and Stubbs (1969, p. 70) decided that the wide extent of the erosion levels at about 305 m. on Jason and Churchill Peninsulas on the east coast of Graham Land also precludes a strandflat origin and suggested they were fluvially eroded, but the inland margins of flood plains are probably less definite and even less straight than those produced by headwall erosion. Dewar (1967, p. 46–47) also did not favour Holtedahl's hypothesis for the formation of the sub-ice-piedmont surfaces on Adelaide Island but he demonstrated that the plateau and piedmont surfaces there could not have originated as one planed surface which has subsequently been faulted to different levels.

When the ice mantle in the Antarctic Peninsula region was much thicker than it is now, there was a large ice shelf along the west coast of the peninsula similar to the present-day Larsen Ice Shelf on the east coast. Fleming (1940, p. 95), pointing to the ice piedmont around Cape Berteaux as the landward remnant of a formerly more extensive Wordie Ice Shelf, inferred that the "fringing glaciers" on the west coast of Graham Land and the ice caps on the low off-lying islands are the stranded remnants of this earlier western ice shelf. This explains the unexpected thickness of the "fringing glaciers" but not the origin of their benches and, although he was critical of Holtedahl's hypothesis of their origin, Fleming could not abandon it without a more acceptable alternative. Perhaps these sub-ice-piedmont surfaces were glacially eroded when the amount and movement of ice on them (and thus the efficiency of glacial scouring) were greater than at present. Headwall erosion may also have been more vigorous under a previous ice regime. Since the formation of these piedmonts is apparently related to their proximity to sea-level, it is important that the general sea-level is probably less than 100 m. higher now than it was during a world-wide glacial maximum (Flint, 1971, p. 318) and the effective rise in sea-level around Antarctica has presumably been less than this because of isostatic recovery.

Intermediate planed surfaces

Many continuous platforms, bevelled summits and mountainous areas with striking accordances of summit heights have been interpreted as the remnants of planed surfaces intermediate in position and altitude between the central snow plateau and the coastal ice piedmonts. However, they are often difficult to recognize and they sometimes cannot be confidently distinguished from the surfaces already described.

Dewar (1967, fig. 1) found several benches between 730 and 990 m. a.s.l. on Adelaide Island and Marsh and Stubbs (1969, p. 59) noted many others on the east coast of Graham Land, especially those at about 450 and 900 m. a.s.l. in the hinterland of Jason Peninsula. The 250 m. platform on Kenyon Peninsula which rises to 500 m. in its hinterland may be an intermediate planed surface but, although they are partly demarcated by snow scarps, this and the islands off the east coast of Palmer Land may be remnants of exceptionally wide ice piedmonts similar to those seen today in western Alexander Island. In the area immediately south of the transition zone, Davies (1975) has listed five uncovered planed surfaces but none occurs between 250 and 1,250 m. a.s.l. and there seem to be three plateau levels. The snow surface east of the Palmer Land snow plateau appears to decline towards the east coast in a series of steps, and for about 200 km. the western margin of the plateau is marked by an irregular discontinuous scarp at the foot of which there is a bench of variable width. The plateau surface of Alexander Island seems to dip southward but the flat summit of the Elgar Uplands is an intermediate planed surface, since it is almost 1,000 m. lower than the accordance of summit heights of the northern Douglas Range to the east (Bell, 1974, p. 12).

Marsh and Stubbs (1969, p. 70) believed that the benches at intermediate levels on the east coast of central and southern Graham Land, like the unusually large ice piedmonts in parts of this area, were formed by fluvial erosion during periods of stability in the spasmodic uplift of the whole region in the Tertiary. Although some of the intermediate planed surfaces in this region are very extensive, they do not seem to occur at consistent levels which would be expected if this were the case. Dewar (1967, p. 45) showed that the intermediate benches on Adelaide Island are up-faulted parts of the surface beneath the Fuchs Ice Piedmont but, in the absence of any other positive evidence to the contrary, it is suggested that most of the intermediate planed surfaces in this region are down-faulted parts of the plateau surface, which is the largest and most widely distributed of the two main genetic types of erosion level. If the peninsula was uplifted between fault complexes, it is possible that the marginal parts were not elevated as high as the axial section.

LANDSCAPE EVOLUTION

The origins of the planed surfaces on the Antarctic Peninsula and its off-lying islands help to elucidate the post-Mesozoic tectonic and hence geomorphological histories of this region but, since it has been concluded that they are still uncertain, the following deductions can only be tentative.

During and perhaps after the emplacement of the batholithic complex of the Antarctandes, major faulting and uplift of the whole region occurred, probably as a direct result of the intrusions or due to isostatic recovery after the cessation of subduction on the Pacific side of the magmatic belt. The outcrops of the metamorphic complex and Trinity Peninsula "Series" may suggest a northerly tilt of the north and central parts of the Antarctic Peninsula, but the differences in the depth of erosion of the batholithic complex in the southern part of the peninsula (Rowley, 1973) and the structure of the Mesozoic sediments in Alexander Island (Bell, 1975) suggest pre-planation southerly tilts of the southern part of the region had occurred. The major longitudinal faulting and the relative uplift of the axial part of the Antarctic Peninsula were probably initiated at this time, if not earlier.

There followed a period of relative stability when most of the region was almost reduced to a peneplain by subaerial denudational processes which left broad valleys and a few monadnocks, and in northern Palmer Land a backbone of high mountains. Tertiary marine sedimentation occurred during the Lower Miocene in the James Ross Island area and the subsequent volcanicity there is also believed to have been accompanied by major subsidence, but this was probably localized.

The semi-peneplain was then uplifted to its present position, the axial part of the peninsula rising higher than some marginal areas where the surface now lies at intermediate levels. This epeirogeny was remarkably uniform, since the northward tilt of the plateau surface first noted by Nichols (1953, p. 7) is more apparent than real and the reported southward tilt of Alexander Island (which may alternatively be accounted for by transverse faulting or may even be partly primary) amounts to much less than 1°. The time and cause of this second uplift during the Cenozoic are not known but it may have been related to the late Cenozoic volcanism in the Antarctandes. It should be noted that the onset of the glaciation would account for little of the emergence of the peninsula because the world-wide sea-level is probably less than 50 m. lower now than it was before any ice caps formed in either hemisphere (Flint, 1971, p. 318).

The occurrence of two major uplifts of the Antarctic Peninsula region during the Cenozoic is mainly deduced from the evidence of major longitudinal faulting and tilting prior to the formation of the plateau surface and the interpretation of this surface as a fluvially eroded semi-peneplain. However, more direct evidence of these earth movements and their respective ages may be gained from the data being obtained from site 325 of the Deep Sea Drilling Project (Fig. 1), for example, as a marked minimum in the sedimentation rate during the Tertiary. The age of the first ice-rafted debris at this site will also be interesting, although its reflection of the onset of the major glaciation in the Antarctic Peninsula region will be difficult to prove. At present it appears that the glaciation began at approximately the same time as the formation of the present cordillera. (Kremp (1964) has suggested a causal relationship between the rising of the mountain ranges of Antarctica and the commencement of glaciation.)

Although it is not yet known whether the main agents of erosion were fluvial or glacial at the time of the last uplift and rejuvenation of erosion, all the features in the dissected coastal belts today are believed to be the products of glacial erosion. The strandflats beneath the relatively inactive "fringing glaciers" may have been formed when the ice mantle was thicker than it is today and there was an extensive ice shelf along the western coast of the Antarctic

Peninsula. The other ice-piedmont surfaces are probably as old as the strandflats but they have been further eroded more recently. During the glacial maximum (or maxima), erosion by the plateau ice caps may have been substantially more vigorous than it seems to be now but the character of the semi-peneplain has probably not been greatly modified. Although the ice mantle is probably much thinner than it has been in the past, glacial erosion is still active in the coastal belts of the peninsula and on the off-lying islands.

The faulting of the main ice-piedmont surface on Adelaide Island shows that shallow tectonism has occurred in recent times but today the Antarctic Peninsula region is seismically quiet except around the South Shetland Islands.

Structural Significance of the Transition Zone

Although the physiographic dissimilarities between Palmer Land and Graham Land have long been apparent, their significance has received little attention. Mason (1950, p. 412) first commented on the differences in the east coast topography north and south of Joerg Peninsula, and Marsh and Stubbs (1969, p. 72) noted that this change coincides with a change in the trend and the increased width of the peninsula. They suggested these differences "have resulted mainly from variations in the degree of tectonism, due to the change in the form of the arc of the Antarctic Peninsula" but this cause was not clarified. Linton (1963, p. 282) suggested that the discrepancy between the respective widths of the two parts of the Antarctic Peninsula resulted from differing degrees of glaciation in the past and differing stages of deglaciation today, but the transition zone is more dissected than the rest of the long narrow plateau and the three-fold change in the width of the peninsula takes place over a fairly narrow and sharply defined area.

The rapid major physiographical changes across the transition zone and the dissected nature of the zone itself strongly suggest the occurrence of large-scale transverse faulting there. The Neny Glacier–Gibbs Glacier and Windy Valley–Lammers Glacier troughs are clearly fault lines, particularly the former, which may continue north-west to Calmette Bay and south-east along the discontinuous rock scarp at the head of Mobiloil Inlet which Fraser and Grimley (1972, p. 5) believed to be one of several "unmistakable physiographic expressions of major fault systems" in this area. They found geological evidence of north-west trending faults south of Kay Nunatak and in the immediate hinterland of Kenyon Peninsula (Fraser and Grimley, 1972, fig. 3), but there is little readily accessible rock exposure adjacent to Neny and Gibbs Glaciers and no geological evidence of faulting has been found there. Adie (1971, p. 122) has suggested that a tectonic break between Graham Land and Palmer Land occurs in the vicinity of "Neny Trough", but the physiographic changes between these two areas occur over a 150 km. wide zone, suggesting that the tectonic break is complicated.

Although there is little detailed geological evidence for major faults occupying the continuous troughs across the transition zone, there is some regional evidence for a geological discontinuity in this zone. The (?) Palaeozoic–Mesozoic metamorphic–volcanic belt which forms most of the central part of the peninsula apparently crosses this zone essentially unchanged, but the (?) Lower Palaeozoic strato-tectonic belt to the west comprising metasediments and locally intercalated lavas (and perhaps also the adjacent Upper Mesozoic sedimentary belt) appears to be transected abruptly at the northern end of Alexander Island, between the latitudes of the transition zone on the peninsula. The preliminary results of a reconnaissance aeromagnetic survey of the Marguerite Bay area suggest that the west side of a line between George VI Sound and the western side of Adelaide Island is relatively non-magnetic and magnetically featureless compared with the mainland side; it may therefore be sedimentary with little shallow more basic volcanic or plutonic rock (personal communication from R. G. B. Renner). Even if the sedimentary sequences of Alexander Island continue northward, it is likely they are considerably thinned. It is also possible that most of the sediments on the east coast of Palmer Land constitute a formation that is confined to the area south of the transition zone.

At present there is no definite evidence that the respective histories of southern Graham Land and northern Palmer Land are *not* similar at least until the late Mesozoic, and no major net transcurrent or vertical faulting between them can be deduced from the known geology. Little has been found in the field of any post-Mesozoic geological record in this region apart from widely scattered outcrops of volcanic rocks, and the transition zone forms no clear division in type, age or distribution pattern of magmatism. The Cenozoic tectonic history outlined above also indicates no important differences between the two parts of the peninsula, and they were finally uplifted by similar amounts. The comparable heights of the plateau blocks in the transition zone and the adjacent main plateaux demonstrate that there was no major net transverse vertical faulting of the peninsula during this last uplift but there is probably a major transverse fault along the northern margin of Alexander Island.

So far, geophysical work has not shown any conspicuous features in the transition zone or any evidence of a change in the structure of the Antarctic Peninsula there. Except around the South Shetland Islands in the north, the whole region is volcanically and seismically inactive, so the transition zone is not a discontinuity in present-day tectonic activity either.

The paucity of geological evidence in support of the physiographical indications that an important tectonic break in the Antarctic Peninsula occurs in the transition zone described here suggests that this feature, at least in its present form, is young and it may only date from the last epeirogeny. The differing widths of Graham Land and Palmer Land, and the contrasting features of their western sides may be explained by different spacings of the main longitudinal faults between which the peninsula was uplifted, and different effects of other longitudinal faults north and south of the transition zone. To accommodate these differences in the longitudinal faulting, transverse faulting would have been initiated in this zone. If this occurred, little net movement along the transverse faults would be expected between most of Graham Land and Palmer Land, since they were apparently uplifted by similar amounts, but considerable vertical throws may have developed in the marginal areas, particularly west of the mainland. Such a dramatic change in the effects of the latest earth movements is likely to have been controlled by older structures, but the only obvious factor in the location of the transition zone is the marked widening of the continental crust on its south side, apparently due to the deposition of the sedimentary and volcanic formations of Alexander Island.

Gansser (1973) has subdivided the Andes into three (or four) physiographically and geologically distinct blocks separated by transverse boundaries, which coincide with the junctions of major oceanic features with the continent. On the basis of present seismicity, Lomnitz (1962) had previously subdivided Chile into four provinces which he noted corresponded with physiographic units and, also emphasizing the divisions of the modern volcanic belts and the marginal trench after Carr and others (1973), Sillitoe (1974) has recently subdivided the Central Andes into segments 100–300 km. long which he believed reflect the divisions between sections of oceanic lithosphere being subducted individually. Similar criteria cannot be used to subdivide the Antarctic Peninsula, because of its recent seismic and volcanic quiescence (except in the extreme north) and the obscurity of the old trench (Houtz, 1974) believed to lie along its western margin. However, it is possible that further geological and geophysical work on and around the Antarctic Peninsula will reveal subduction-related sub-divisions similar to those found to the north. Detailed geochronological and geochemical studies may indicate transverse breaks in the magmatic pattern of the peninsula's batholith and main volcanic belt, detailed gravity and magnetic surveys perhaps supported by sampling of the western continental shelf may show the nature of the northward continuation (if any) of the geology of Alexander Island, and similar surveys of the adjacent ocean floor may yield its age patterns and discontinuities. However, the present understanding of this region is such that the structural significance of the transition zone is still a subject of speculation.

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